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## Glacial oscillations during the Bølling–Allerød Interstadial–Younger Dryas transition in the Ruda Valley, Central Pyrenees

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2 1 **Glacial oscillations during the Bølling-Allerød Interstadial-Younger Dryas**  
3 2 **transition in the Ruda Valley, Central Pyrenees**  
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## Abstract

The Upper Garonne Basin included the largest glacial system in the Pyrenees during the last glacial cycle. Within the long-term glacial retreat during Termination-1 (T-1), glacier fluctuations left geomorphic evidence in the area. However, the chronology of T-1 glacial oscillations on the northern slope of the Central Pyrenees is still poorly constrained. Here, we introduce new geomorphological observations and a 12-sample dataset of  $^{10}\text{Be}$  Cosmic-Ray Exposure (CRE) ages from the Ruda Valley. This U-shaped valley, surrounded by peaks exceeding 2800 m a.s.l., includes a sequence of moraines and polished surfaces that enabled to reconstruct the chronology of the last deglaciation. Following the maximum ice extent, warmer conditions prevailing at ~15-14 ka, during the Bølling-Allerød (B-A) Interstadial, favoured glacial retreat in the Ruda Valley. Within the B-A, glaciers experienced two phases of advance/stillstand with moraine formation at 13.5 and 13.0 ka. During the early YD, glacial retreat exposed the highest surfaces of the Saboredo Cirque (~2300-2350 m) at 12.7 ka. Small glaciers persisted only inside the highest cirques (~2470 m), such as in Sendrosa Cirque, with moraines stabilizing at 12.6 ka. The results of this work present the most complete chronology for Pyrenean glacial oscillations from the B-A to the YD.

**Key words:** Central Pyrenees, Bølling-Allerød, Younger Dryas, Cosmic-Ray Exposure dating, deglaciation

## 1. Introduction

The present-day landscape of the highest mid-latitude mountain environments is dominated by the effects of glacial dynamics that prevailed during the last glacial cycle and of post-glacial processes that reshaped glacial landforms during the subsequent deglaciation. Following the end of the Last Glacial Maximum (LGM) at 19-20 ka (Clark et al., 2009), most glaciated mountain regions underwent a rapid deglaciation only interrupted by brief periods of glacial growth. Glacial disappearance favoured paraglacial readjustment and intense periglacial and slope processes, which reshaped the already deglaciated valleys and cirques (Ballantyne, 2008). The current spatial distribution of glacial and periglacial features in the high elevations of most mid-latitude ice-free mountain ranges – such as the Pyrenees, where this research focuses on – results from the climatic shifts during Termination-1 (hereafter T-1), the period spanning from 20-19 to 11.7 ka (Denton et al., 2014).

At present, mid-latitude mountains such as those of the Mediterranean region are only sparsely glaciated, with small glaciers located in sheltered cirques (Hughes, 2014). However, there are several other mountain ranges located between 30 and 50° latitude North, such as the Alps or the Himalaya, which are still heavily glaciated and record recent dramatic glacial recession trends (Huss et al., 2021). An accurate comprehension of the climatic sensitivity of current glaciated mountain landscapes can help to understand the timing and magnitude of glacial change during the rapid climate shifts of T-1. The alternation of millennial-scale cold/warm periods during T-1 promoted a temporal pattern of glacial advances and retreats across the Northern Hemisphere (Allard et al., 2021; Rea et al., 2020). Ice core records from the Greenland interior show the following succession of climatic phases in the North Atlantic region during T-1: a cold climate regime during the Oldest Dryas (OD; GS-2.1a: 17.5-14.6 ka), followed by the general warm Bølling-Allerød (B-A) Interstadial (GI-1: 14.6-12.9 ka), with a subsequent intensification of cold conditions during the Younger Dryas (YD; GS-1: 12.9-11.7 ka) (Rasmussen et al., 2014).

In the Iberian mountains, recent advances in constraining the chronology of glacial activity during T-1 using Cosmic-Ray Exposure (CRE) dating have revealed contrasting spatio-temporal patterns (García-Ruiz et al., 2016b; Oliva et al., 2019; Palacios et al., 2017a). Whereas the maximum ice extent occurred asynchronously, glacial oscillations during T-1 show a more synchronous response: colder phases favoured glacial growth, whereas warm periods promoted glacial recession (Oliva et al., 2019) and the evolution from debris-free glaciers to rock glaciers such as those in the Iberian Range (García-Ruiz et al., 2020), Cantabrian Mountains (Rodríguez-Rodríguez et al., 2016) and Central (Palacios et al., 2017a, b) and Eastern Pyrenees (Andrés et al., 2018). To date, there is geomorphic evidence of glacial advances during the OD and YD, as well as of glacial retreat during the B-A and following the YD in the Cantabrian Mountains (Rodríguez-Rodríguez et al., 2017), Central Range (Carrasco et al., 2015; Palacios et al., 2012b), and Sierra Nevada (Gómez-Ortiz et al., 2012; Palacios et al., 2016). However, CRE ages require a careful interpretation and support of geomorphological criteria, as their uncertainty ranges may be greater than the time spanning the T-1 phases (Oliva et al., 2021a).

The Pyrenees were the most extensively glaciated mountain system during the maximum ice extent of the last glaciation in Iberia, hosting ~50% of the peninsula's glacial surface (Oliva et al., 2021b). The chronology of glacier fluctuations during T-1 seems to have followed a similar temporal pattern, although available data is principally derived from the range's southern slope (Andrés et al., 2018; Palacios et al., 2015b, 2017b, 2017a). Pyrenean glaciers readvanced to the valley bottoms during the OD, with ice tongues > 15 km long. During the B-A, by 15-14 ka, high temperatures favoured rapid deglaciation and glaciers disappeared from relatively low cirques shaped under peaks of ~2600 m a.s.l. (Oliva et al., 2021a). A subsequent readvance occurred during the YD, with smaller glaciers of only ~4 km length confined within the cirques of the highest massifs above 2200 m (Crest et al., 2017; García-Ruiz et al., 2016a; Jomelli et al., 2020; Palacios et al., 2017b, 2015a; Pallàs et al., 2006). Following both cold periods, glacial retreat triggered intense paraglacial dynamics at ~15-14 ka and ~11-10.5 ka, respectively, which led to the formation of rock glaciers and debris-covered glaciers, particularly in cirques below 2800 m (Andrés et al., 2018; Palacios et al., 2017b).

The timing of glacial oscillations is still poorly constrained on the northern slope of the Pyrenees, such as in the Upper Garonne Basin where this work focuses on. Recently, Reixach et al. (2021) have summarized the glacial evolution from 19 to 12 ka in the northern Pyrenees, reporting a spatial and temporal pattern for T-1 glacial oscillations similar to that on the southern slopes of the range. Jomelli et al. (2020) confirmed glacial growth during the YD in the Ariège Valley, where glaciers persisted in the highest cirques until the early Holocene. Another recent study that focused on the Bacivèr Cirque, near our study area, revealed a rapid deglaciation during the B-A, followed by glacial readvance and the formation of the highest moraines at the B-A/YD transition at ~12.9 ka; subsequently, intense paraglacial readjustment favoured the formation of debris-covered glaciers and rock glaciers inside these moraine systems (Oliva et al., 2021a).

In order to complement our understanding of the glacial history of the Pyrenees, this study uses new data to improve the deglaciation chronology of the Upper Garonne Basin, the largest glacial system in this mountain range, during the local last glacial maximum (ILGM; Fernandes et al., 2017). Glacial retreat in the Ruda Valley, where this work focuses on, left widespread geomorphic evidence of glacial and paraglacial origin that can be used to reconstruct the last stages of long-term deglaciation in the Central Pyrenees.

Our study, therefore, has the following objectives:

- To date the glacial landforms of the Upper Garonne Basin.
- To discuss the T-1 deglaciation in the Upper Garonne Basin and its geomorphological implications in the context of the known chronologies from the Pyrenees southern slope and from other Iberian and European mountain ranges.
- To examine the paleoclimatic and paleoenvironmental significance of the glacial response of the northern slope of the Pyrenees and compare it with the paleoclimatic evolution inferred from other records in the North Atlantic region.

## 2. Study area

The Pyrenees stretch E-W over 400 km across north Iberia, with the highest peaks (> 3000 m) in the central part of the range. This study focuses on the Ruda Valley, located between latitudes 42°40' N - 42°42' N and longitudes 0°57' E - 1°00' E (Fig. 1), that constitutes the headwaters of the Upper Garonne Basin. The area is included in the peripheral protection area of the National Park of Aigüestortes and Sant Maurici Lake.

### Figure 1

The Upper Garonne Valley shows the typical U-shaped cross-section of glacial valleys, and the Garonne River flows to the N-NW through the central axis of the Pyrenees. Several tributary rivers drain the glacial valleys descending from glacial cirques carved below peaks from 2800 to 3000 m. Glacial cirques are prevalingly NE-oriented, with floors generally between 2200 and 2400 m (Lopes et al., 2018).

The U-shaped Ruda Valley, drained by the Ruda River that runs 12 km northwards, constitutes the headwaters of the Upper Garonne Basin. Its catchment encompasses 34 km<sup>2</sup>, with an elevation range of ~1300 m between Saboredo Peak (2827 m) and the village of Baqueira (~1500 m). The Saboredo Cirque is located in the headwaters of the Ruda Valley; the cirque is N-exposed, extending over 5.7 km<sup>2</sup>, with its floor at ~2250-2400

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2 142 m. Saboredo (2827 m), Sendrosa (2702 m) and Locampo (2658 m) peaks are the highest elevations of the  
3 143 cirque, while its toe is located at ~2250 m.

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5 144 The present-day mean annual air temperature (MAAT) at Port de la Bonaigua station (2266 m) is 3 °C, with  
6 145 the regional 0 °C isotherm in the Central Pyrenees estimated to be at 2950 m (López-Moreno et al., 2019).  
7 146 Annual precipitation reaches 1227 mm and falls mostly as snow (average depth of 1 m) from October to May,  
8 147 with a snow duration of 7-8 months in the cirque floor. The combination of low temperatures and abundant  
9 148 snowfall determines the scarce vegetation cover above the treeline (2200-2300 m), where barren rocky terrain  
10 149 is widespread, together with small alpine meadows.

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13 150 The landscape of the Ruda Valley resulted from the interaction of tectonics and vertical incision, both during  
14 151 Quaternary glaciations and interglacial periods, the latter dominated by alluvial and mass-wasting processes  
15 152 (Fernandes et al., 2017). Glacial and periglacial landforms are preserved on different lithologies across the  
16 153 valley, composed of Carboniferous granodiorites and granites in the highest sections, with Cambrian-Devonian  
17 154 metamorphic (hornfels, marbles, slates, quartzites, schists) and sedimentary rocks (limestones, sandstones,  
18 155 conglomerates and lutites) in the lowest parts of the valley (ICGC, 2017; Quesada and Oliveira, 2019).  
19 155 However, despite abundant geomorphic evidence revealing that the Garonne Valley presented the longest  
20 156 glacier in the Pyrenees during the ILGM (ca. 80 km; Fernandes et al., 2017), the chronological framework of  
21 157 glacial activity in the entire Upper Garonne Basin is still unknown. The headwaters of the Ruda Valley include  
22 158 a rich sequence of moraine systems and polished bedrock surfaces – as well as relict rock glaciers and protalus  
23 159 lobes (Fernandes et al., 2018) – that can be used to improve the temporal resolution for the chronosequence of  
24 160 the final deglaciation of the upper Garonne Valley.  
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### 29 163 3. Methodology

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31 164 We reconstructed the last deglaciation chronology of the Ruda Valley in the Upper Garonne Basin by  
32 165 integrating geomorphological and chronological techniques. Field work was conducted during the summer of  
33 166 2019, when snow cover was limited to the concave and sheltered areas at the highest elevations (>2500 m),  
34 166 which enabled better identification of the glacial landforms, as well as sampling for CRE dating.  
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#### 36 168 3.1 Geomorphological mapping

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38 169 We produced a geomorphological map focusing on glacial and periglacial landforms based on recent general  
39 170 geomorphological maps from the area (Fernandes et al., 2017; Fernandes et al., 2021). Here we used the same  
40 171 approach based on the combination of aerial orthophotographs (0.25 m cell size) and high-resolution LIDAR  
41 171 digital elevation models (point density of 0.5-2 m<sup>-2</sup>) from the “Institut Cartogràfic i Geològic de Catalunya”  
42 172 (<http://www.icc.cat/appdownloads>). The symbology of geomorphological landforms followed Joly (1997).  
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#### 45 174 3.2 Field strategy and sampling

46 175 We collected 12 granite and granodiorite samples for CRE dating, using a hammer and chisel, from elevations  
47 176 spanning from 1862 to 2470 m (Table 1). To ensure that samples were intact since glacial retreat, we targeted  
48 176 at well-anchored moraine boulders and glacially polished bedrock that stood out within their surroundings,  
49 177 which minimizes the risk of having been buried under sediments (Fig. 2). To ensure the best cosmic-ray flux  
50 178 reception of the sampling sites, we restricted sample extraction to flat-topped and gentle surfaces, and sampled  
51 179 a thickness of up to 3-5 cm. Corrections for partial shielding due to the surrounding topography were  
52 180 implemented using the ArcGIS toolbox by Li (2018), which applies equations describing the effects of  
53 181 geometric shielding on the sampling sites and a model to predict the nuclide production rate at depth in sloped  
54 182 surfaces (Dunne et al. 1999); it required only a point shapefile of the sampling sites including strike/dip of the  
55 183 sampling surfaces and a digital elevation model of 5 m spatial resolution.  
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59 185 Table 1

60 186 Figure 2



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2 187 **3.3 Laboratory procedures and exposure age calculation**

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4 188 Samples were crushed and sieved (>200 g) at the Physical Geography Laboratory of the Universidad  
5 189 Complutense de Madrid (Spain) to the 125-500  $\mu\text{m}$  fraction, which was subsequently processed at the  
6 190 ‘Laboratoire National des Nucléides Cosmogéniques’ (LN<sub>2</sub>C) of the ‘Centre Européen de Recherche et  
7 191 d’Enseignement des Géosciences de l’Environnement’ (CEREGE; Aix-en-Provence, France). In accordance  
8 192 with the quartz-rich lithology of the samples, they were treated for the extraction of the *in-situ*-produced  
9 193 cosmogenic nuclide <sup>10</sup>Be.

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12 194 First, magnetic minerals were discarded through magnetic separation conducted in a “Frantz LB-1” magnetic  
13 195 separator. Once the non-magnetic fraction was isolated, it underwent several rounds of chemical attacks with  
14 196 a 1:2 mixture of concentrated hydrochloric (HCl) and hexafluorosilicic (H<sub>2</sub>SiF<sub>6</sub>, 2/3) acids to dissolve and  
15 197 discard the non-quartz minerals. The remaining minerals (i.e. non-dissolved feldspar minerals) were  
16 198 subsequently attacked with 4-6 successive partial dissolutions with HF to remove the atmospheric <sup>10</sup>Be and  
17 199 discard the remaining impurities. In order to be sure that the quartz was pure, samples were examined under a  
18 200 binocular microscope. As a result, ~20 g of pure quartz per sample was used to extract the <sup>10</sup>Be (Table 2). After  
19 201 adding ~150  $\mu\text{L}$  of an in-house manufactured (from a phenakite crystal) <sup>9</sup>Be carrier solution (spike,  
20 202 concentration:  $3025 \pm 9 \mu\text{g g}^{-1}$ ; Merchel et al., 2008), the purified quartz was dissolved with 48% concentrated  
21 203 HF (3.5 mL per g of quartz + 30 mL in excess). Following total dissolution, the resulting solution was  
22 204 evaporated until dryness, and the residue was recovered with HCl (7.1 molar). Be was then precipitated at  
23 205 pH=8 to beryllium hydroxide (Be(OH)<sub>2</sub>) by means of ammonia (NH<sub>3</sub>), and separated from other elements in  
24 206 resin columns: a Dowex 1x8 anionic exchange column to remove elements such as Fe, Mn and Ti, and a  
25 207 Dowex 50W x8 cationic exchange column to discard B and separate Be from Al (Merchel and Herpers, 1999).  
26 208 The final eluted Be was again precipitated, dried and oxidized to BeO at 700 °C. Finally, the targets for the  
27 209 accelerator mass spectrometer (AMS) measurements were prepared by mixing with niobium powder at an  
28 210 approximate 1:1 proportion and pressing the mixture into copper cathodes.

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31 211 The targets were analysed at the ‘Accelerator pour les Sciences de la Terre, Environnement et Risques’  
32 212 (ASTER) national AMS facility at CEREGE in order to measure the <sup>10</sup>Be/<sup>9</sup>Be ratio from which the <sup>10</sup>Be  
33 213 concentration was later inferred (Table 2). AMS measurements were calibrated against the in-house standard  
34 214 STD-11 with an assigned <sup>10</sup>Be/<sup>9</sup>Be ratio of  $(1.191 \pm 0.013) \times 10^{-11}$  (Braucher et al., 2015). Based on calibration  
35 215 with the previous ASTER AMS standard NIST-27900 (SRM4325), the new STD-11 is fully compatible with  
36 216 07KNSTD. The analytical 1 $\sigma$  uncertainties include uncertainties in the AMS counting statistics, an external  
37 217 0.5% AMS error (Arnold et al., 2010) and the uncertainty related to the chemical blank correction. The <sup>10</sup>Be  
38 218 half-life considered was  $(1.3870 \pm 0.0012) \times 10^6$  years (Chmeleff et al., 2010; Korschinek et al., 2010).

39 219 **Table 2**

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42 220 We calculated <sup>10</sup>Be exposure ages by using the CREP online calculator (Martin et al., 2017; available online  
43 221 at: <http://crep.cirp.cnr.fr/#/>), where we selected the following settings: LSD (Lifton-Sato-Dunai)  
44 222 elevation/latitude scaling scheme (Lifton et al., 2014), ERA40 atmospheric model (Uppala et al., 2005) and  
45 223 geomagnetic database based on the LSD framework (Lifton et al., 2014). These settings yielded a world-wide  
46 224 mean sea level high latitude (SLHL) <sup>10</sup>Be production rate of  $3.98 \pm 0.22 \text{ atoms g}^{-1} \text{ yr}^{-1}$ . Exposure ages of the  
47 225 samples with their 1 $\sigma$  full and analytical uncertainties are shown in Table 2. The uncertainties discussed  
48 226 throughout the text include analytical and production rate error unless otherwise stated.

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50  
51 227 In order to evaluate the potential impact of erosion on the exposure ages, we assumed a steady erosion rate (1  
52 228 mm ka<sup>-1</sup>) following a conservative maximum value in such a lithological setting according to André (2002).  
53 229 The impact of snow cover was explored assuming that snow currently remains in the area an annual average  
54 230 of 7.5 months at 2200-2300 m with a mean thickness of 100 cm – and assuming a snow density of 0.2 g cm<sup>-3</sup>  
55 231 (Styllas et al., 2018) – (Table 3), based on the available data from the Bonaigua weather station (Fig. 1)  
56 232 available since 1997-1998 (Servei Meteorològic de Catalunya; [http://www.igc.cat/web/ca/allaus\\_gruix\\_neu\\_v2.php?e=bonaigua&t=totes](http://www.igc.cat/web/ca/allaus_gruix_neu_v2.php?e=bonaigua&t=totes)). Erosion and snow corrections, as  
57 233

a whole, resulted in ages that were ~9% older (~1% erosion; ~8% snow cover). However, to enable comparisons with other areas, we use the non-corrected ages throughout the text. Moreover, both the thickness and duration of snow cover must have varied significantly since the cirque's deglaciation, given the alternation of cold/warm periods and their associated moisture supplies. Finally, the chi-squared test according to Ward and Wilson (1978) was applied to the samples belonging to each geomorphological unit established in the geomorphological map (Table 1) in order to detect potential exposure age outliers (i.e. "too old" or "too young"; Heyman et al., 2011). No outliers were detected (Table 3).

### Table 3

### 3.4 Glacier reconstruction and Equilibrium-Line Altitude (ELA) calculation

A three-dimensional paleoglacier reconstruction was performed for the different glacial phases using the 'GLaRe' ArcGIS toolbox devised by Pellitero et al. (2016). Former ice thickness was estimated by applying a perfect-plasticity physical-based rheological model along flowlines from the termini to the headwall (Van der Veen, 1999; Benn and Hulton, 2010). Such a toolbox requires only a flowline, a tentative paleoglacier geometry - approximated from lateral or frontal moraines - and a digital elevation model. Ice thickness was modelled by using a constant shear stress of 100 kPa (Paterson, 1994; Benn and Hulton, 2010). The effect of lateral side drag was corrected by integrating shape factors (F-factor) obtained from cross-sections (Schilling and Hollin, 1981).

ELAs were calculated by using the automatic toolbox 'ELA calculation' developed by Pellitero et al. (2015). The selected methods were the Accumulation Area Ratio (Porter, 1975; AARs:  $0.6 \pm 0.05$ ) and the Area Altitude Balance Ratio (AABR; Osmaston, 2005). We considered two balance ratios for the AABR method:  $1.75 \pm 0.71$  (global) and  $1.9 \pm 0.81$  (mid-latitude marine glaciers) (Rea, 2009). The average ELAs were used to discuss the paleoclimatic implications of the identified glacier advance/stillstand phases.

## 4. Results

The Ruda Valley preserves glacial and periglacial landforms indicative of glacial evolution during the last deglaciation in the Upper Garonne Basin. The chronological framework of this evolution is unveiled by a dataset of 12  $^{10}\text{Be}$  CRE ages obtained from glacial landforms.

### 4.1 Geomorphological evidence

The wide range of glacial features from the highest sectors to the Ruda Valley floor shows evidence that the entire valley was heavily glaciated during the ILGM and recorded several advances or stillstands within the long-term deglaciation trend. The Ruda Valley is divided into two main geomorphological units (Fig. 3):

- (i) The valley floor and mountain slopes follow the typical U-shaped morphology of a glacial valley. The steep slopes, descending from ~2200 to 1800 m, connect the lateral cirques with the valley floor. The hillslopes of the valley start at the upper edge of hanging glacial cirques that define the glacial trimline, marking the uppermost extent of recent glaciations. Hillslopes are concave to rectilinear 300-400 m-long sections steeper than  $30^\circ$  that are mostly colonized by forest (i.e. *Pinus nigra*). The upper part of the valley is defined by two staggered, overdeepened basins filled with peatlands at average elevations of 2150 and 2070 m. At their margins, two moraine systems occur at 2190 and 2080 m. Two samples were collected from boulders from the external frontal moraine (ARAN-38 and ARAN-37) and the other two were obtained from the internal moraine unit: one from a boulder on the moraine ridge (ARAN-27), and the other from a polished surface with glacial striae (ARAN-28) and moraine boulders (Fig. 4). The lowest part of the valley floor, with a relatively flat slope characteristic of U-shaped glacial valleys and descending from ~1800 to 1500 m, is colonized by alpine meadows and coniferous forests. Isolated *roches moutonnées* are found in the highest section of the valley bottom where granites prevail ( $> \sim 1800$  m), while they are absent in lower areas, where slates and marbles are the dominant lithology. Two samples were collected from polished surfaces at 1900 and 1860 m (ARAN-39 and ARAN-40).

(ii) The highest peaks of the Ruda Valley surrounding the Saboredo Cirque range from 2700-2800 m (Sendrosa and Saboredo peaks) to 2525 m in the mountain divide that may have functioned as a glacial diffluence pass between the Saboredo (Garonne basin) and the Ratera cirques (Noguera Pallaresa basin) (Fig. 3). Saboredo is a complex cirque (Barr and Spagnolo, 2015) that includes several minor hollows carved out on the marginal flanks of the peaks. They are separated by truncated spurs with polished surfaces, where one sample (ARAN-31) was collected to examine the timing of glacial thinning at the cirque level. The highest moraines in the Ruda Valley are located at ~2400-2500 m, at the foot of the highest peaks. A prominent cirque moraine at an altitude of 2470 m is located on the eastern slope of Sendrosa Peak; here, we collected two samples (ARAN-33 and ARAN-32). Above the upper moraines, fed by abundant debris supply from the cirque rock walls, rock glaciers and protalus lobes are found. The cirque floors extend between 2200 and 2400 m and preserve features indicative of intense glacial erosion (e.g. overdeepened basins or glacial thresholds) and glacial deposits (e.g. moraines or erratic boulders). In the Saboredo Cirque, glacial abrasion and periglacial processes shaped a large amphitheatre with several overdeepened basins. They are surrounded by polished surfaces and occupied by peatlands and lakes. Three samples were collected from polished surfaces from 2360 to 2270 m (Fig. 3) (ARAN-36; ARAN-35; ARAN-34; Fig. 2; Table 1). Glacial retreat left scattered boulders and till across the area.

Figure 3

Figure 4

#### 4.2 Geochronological data

The 12 CRE samples collected from the Ruda Valley returned a chronology spanning from  $\sim 15 \pm 0.9$  to  $11.8 \pm 0.7$  ka, with all ages being consistent with the geomorphological distribution of the dated surfaces (Fig. 4).

Polished surfaces from the highest parts of the Ruda Valley located ~200-250 m below the cirques provided CRE ages of  $15.0 \pm 0.9$  (ARAN-40) and  $13.8 \pm 0.8$  ka (ARAN-39, Fig. 4A) (average:  $14.4 \pm 1.2$  ka). A similar age was obtained from a sample collected from a truncated spur dividing the Sendrosa and Saboredo cirques, which yielded  $14.8 \pm 0.9$  ka (ARAN-31). A sample collected from a polished surface at the edge of the Saboredo cirque floor (Fig. 3) reported an age of  $14.0 \pm 0.8$  ka (ARAN-34, Fig. 4D).

Moraine boulders from the external moraine located in the lower overdeepened basin yielded ages of  $13.3 \pm 0.8$  (ARAN-38) and  $13.7 \pm 1.0$  ka (ARAN-37) (average:  $13.5 \pm 0.9$  ka). A boulder from the internal moraine ridge located 110 m above the former moraine provided an age of  $13.1 \pm 0.8$  ka (ARAN-27); this age was compared with a sample from a polished surface which correlates with the internal moraine (Fig. 4C) that returned an age of  $13.0 \pm 0.8$  ka (ARAN-28), which agrees with the latter.

Two samples from polished surfaces of the Saboredo Cirque above 2300 m yielded ages of  $12.8 \pm 0.8$  (ARAN-35) and  $12.7 \pm 0.8$  ka (ARAN-36) ( $12.7 \pm 0.8$  ka). The highest moraines are located inside the highest cirque hollows, such as the Sendrosa hollow. Here, two moraine boulders yielded ages of  $11.8 \pm 0.7$  (ARAN-33) and  $13.3 \pm 1.0$  ka (ARAN-32) ( $12.6 \pm 1.3$  ka).

#### 4.3 Reconstructed paleoglaciers and ELAs

The ELAs of the three reconstructed paleoglacier extents were calculated for the Ruda catchment. The paleoglacier existing at  $13.5 \pm 0.9$  ka deposited a moraine system at 2080 m, with an average ELA located at 2461 m (Table 4). The paleoglacier at  $13.0 \pm 0.8$  ka formed a moraine at 2190 m, with an average ELA at 2505 m. Finally, the Sendrosa Cirque moraine located at 2470 m was dated at  $12.6 \pm 1.3$  ka, with an ELA of 2571 m (Table 4).

Table 4

### 5. Discussion

#### 5.1 Interpretation of the geomorphological observations and CRE results



The geomorphological analysis supported by  $^{10}\text{Be}$  dating of glacial landforms show the spatial and temporal patterns of the deglaciation between the B-A and the early YD in the Ruda Valley (Fig. 5). Although the CRE dataset is consistent with sequence of glacial landforms, uncertainty remains in assigning ages either to the late B-A or to the early YD.

The ages of samples from polished surfaces at the valley at 1860-1900 m (ARAN-40 and ARAN-39;  $14.4 \pm 1.2$  ka) suggest that the beginning of glacial retreat in the Ruda Valley took place during the B-A. As glaciers retreated, they became confined within their respective cirques, as revealed by the age of  $14.8 \pm 0.9$  ka (ARAN-31) of a polished surface on the ridge dividing the Sendrosa and Saboredo cirques (Fig. 3). The sample from an outcrop on the lower edge of the cirque (ARAN-34;  $14.0 \pm 0.8$  ka) indicates that ice covered the entire Saboredo cirque floor until  $\sim 14$  ka.

The sampled boulders of the external moraine system (ARAN-38 and ARAN-37) showed statistical coherence based on the chi-squared test following Ward and Wilson (1978) and suggested a glacier advance or stillstand at  $13.5 \pm 0.9$  ka. The age of  $13.0 \pm 0.8$  ka from a moraine boulder (ARAN-27) and an adjacent polished surface (ARAN-28) of the internal moraine system indicate another short advance or stillstand, which is also consistent with the geochronological sequence. These results from staggered moraine systems at  $\sim 2080$  and  $\sim 2190$  m provided chronologically consistent mean ages within the B-A. But is based on their distinct relative position that they have been ascribed to two different stages. Nevertheless, during this period, glaciers flowed from the Saboredo and Sendrosa cirques and generated moraines on the lower overdeepened basins.

According to the ages of samples from the highest surfaces of the Saboredo Cirque floor at 2320-2360 m (ARAN-35 and ARAN-36;  $12.7 \pm 0.8$  ka), the entire cirque floor was deglaciated at the onset of the YD. These ages are consistent with those of the samples from the highest dated moraine cirques in the Sendrosa Cirque at 2460-2470 m (ARAN-32 and ARAN-33), which was also abandoned by the ice by  $12.6 \pm 1.3$  ka. No statistical inconsistency was detected in the exposure ages from this moraine.

## Figure 5

### 5.2 The chronology of deglaciation

During the ILGM, the Garonne glacier flowed  $\sim 80$  km reaching the Loures-Barousse-Barbazan basin, where it formed a terminal moraine system (Fernandes et al., 2017; Stange et al., 2014). Until the onset of deglaciation, 300-400 m thick ice sat in the Ruda Valley, as shown by the highest lateral moraines between the Ruda and Aiguamòg valleys. This ice thickness must have favoured the occurrence of warm-based glaciers, thus implying basal sliding evidenced by striations on the bedrock surface. Hence, no nuclide inheritance from previous exposure periods is expected at the sampled surfaces.

Samples from polished surfaces from the bottom of the Ruda Valley at 1860-1900 m indicate that by  $14.4 \pm 1.2$  ka, i.e. during the B-A, the glacier was retreating towards the glacial cirque, and was already disconnected from the main Ruda glacier. Consequently, one third of the Ruda Valley was already deglaciated at that time, indicating that by the early B-A the Garonne glacier represented only  $\sim 5\%$  of the length that it had during the ILGM (Fernandes et al., 2017).

As glaciers retreated and thinned during the B-A, ice masses became individualized within their respective cirques, and the valley glaciers flowing from Sendrosa and Saboredo cirques merged only in the lowest overdeepened basins (Fig. 3). As conditions were less favourable, glaciers receded to higher elevations and the Sendrosa and Saboredo paleoglaciers became isolated in their respective basins by  $14.8 \pm 0.9$  ka (Fig. 3). The ice covered the entire Saboredo cirque floor until  $\sim 14$  ka, as revealed by the sample collected from an outcrop on the edge of the cirque ( $14.0 \pm 0.8$  ka), indicating that glacial recession in the Ruda Valley was underway at  $\sim 15$ -14 ka. Similarly, a rapid deglaciation was also recorded at  $\sim 15$ -14 ka in the nearby Bacivèr Cirque, which favoured its complete deglaciation (Fig. 1; Oliva et al., 2021a).

The strong clustering of ages at between  $\sim 13.5$  and  $12.7$  ka confirms the dynamic environmental response of the Ruda palaeoglaciers to the abrupt climate transition between the B-A and the early YD. However, our CRE

dataset cannot assess the magnitude and patterns of glacial recession during the early B-A in the Ruda Valley, or whether the entire cirque was deglaciated and glaciers reformed and subsequently advanced. Our results suggest, however, that the millennial-scale trend towards glacier recession during the B-A was interrupted by two glacial advances or stillstands with moraine development at  $13.5 \pm 0.9$  and  $13.0 \pm 0.8$  ka (Fig. 5). Samples from moraines located at average elevations of  $\sim 2080$  and  $\sim 2190$  m, respectively, indicated the existence of valley glaciers 3.5 and 2.5 km long during the middle of the B-A and during the transition from the B-A to the YD. Indeed, the reconstructed glacial surface for both bases was of 4.4 and 3.8 km<sup>2</sup>, respectively, with a maximum ice thickness of 300 and 200 m for each of these phases (Fig. 6a, b).

### Figure 6

Evidence of glacial recession during the middle of the B-A was also recorded on polished surfaces from the neighbouring Bacivèr Cirque at  $13.5 \pm 0.8$  ka (Oliva et al., 2021a). Local topographical factors, such as the shape of the cirque floor (convex versus concave areas) determined the pattern of glacial flow of thin ice masses during the glacial advances/stillstands that occurred during the B-A. Glaciers flowed downwards from the cirque through concave areas across the western slope, favouring the development of the moraine arcs and crests in lower overdeepened basins (Fig. 3). It is likely that the topography played a major role in determining the process and timing of the deglaciation in both cirques: the accumulation area above 2400 m is larger in the Saboredo Cirque (552 ha) with respect to the Bacivèr Cirque (244 ha) and its prevailing aspect is also more favourable for ice persistence (N vs W).

The Saboredo Cirque floor became ice free at the beginning of the YD, as revealed by the samples from polished surfaces of the cirque floor at  $\sim 2350$  m (average age =  $12.7 \pm 0.8$  ka). The evidence of glacial striations on the rock surface make unlikely the possibility of nuclide inheritance at the sampling sites. Glaciers might thus also have persisted during the early YD in the highest hollows of the Saboredo Cirque, as was the case in the northern cirque of Sendrosa Peak. Here, the cirque moraine at 2470 m, formed by a 0.4 km-long glacier (with a surface of 0.7 km<sup>2</sup> and a thickness of 90 m; Fig. 6c), was abandoned at  $12.6 \pm 1.3$  ka. Inside the highest glacial cirques of the Ruda Valley, there are numerous relict rock glaciers and protalus lobes indicative of the past occurrence of permafrost during their formation (Fernandes et al., 2018). As in most of the Pyrenees, these are probably glacier-derived permafrost features associated with the paraglacial phase, when glacial shrinking favoured wall debuttrassing, a high debris supply and the subsequent burial of the residual ice masses (Andrés et al., 2018; Oliva et al., 2016).

In short, by the early B-A the Garonne Paleoglacier had receded  $\sim 95\%$  of its  $\sim 80$  km length of the ILGM (Fernandes et al., 2017; Stange et al., 2014). The Ruda Paleoglacier receded 4 km in only  $\sim 1.7$  ka ( $\sim 14.4$  to  $\sim 12.7$  ka) from an elevation of  $\sim 1800$  to  $>2400$ - $2500$  m. This trend towards glacial shrinking before the final ice disappearance was interrupted by three phases of glacier advance or stillstands at  $13.5 \pm 0.9$ ,  $13.0 \pm 0.8$  and  $12.6 \pm 1.3$  ka, which generated moraines at  $\sim 2080$ , 2190 and 2470 m, respectively.

### 5.3 Late Quaternary glacial dynamics in the Central Pyrenees and Iberian Peninsula in the context of European glacial evolution

The scientific knowledge on the chronology of the last deglaciation in the Pyrenees has experienced substantial advances over the last years in parallel to progress on glacial chronological constraints in other European and Iberian mountain ranges (García-Ruiz et al., 2010; Palacios and García-Ruiz, 2015). There is increasing evidence that the timing of glacial oscillations during T-1 across Europe followed a similar sequence (Palacios et al., 2021): a significant glacier advance took place during the OD followed by a rapid retreat during the B-A and a subsequent more limited glacier advance during the YD (Oliva et al., 2019).

Dated evidence of glacial responses in the Pyrenees to Northern Hemisphere cold events during T-1 have mostly focused on the southern slope (Oliva et al., 2019), although there is an increasing body of new data from the northern slope of the range that confirms glacial advances and stillstands there during the OD and YD, and glacial shrinking during the B-A (Reixach et al., 2021). The elevation of the peaks as well as the topography of the cirques played a major role in the duration and intensity of the glaciation during T-1 in the

high massifs of the Pyrenees. Cirques surrounded by peaks above 2800-2900 m likely included glaciers during most of T-1 until the onset of the Holocene (see e.g. Jomelli et al., 2020; Pallàs et al., 2006; Reixach et al., 2021), while those situated at the foot of lower peaks (i.e. 2500-2600 m) were completely deglaciated during the B-A (Oliva et al., 2021a; Palacios et al., 2017b). Following ice retreat, the slopes were affected by intense readjustments (large, catastrophic rock slope failures, rock slope deformations, rockfalls, and rockslides), particularly in areas where the local lithostructure was prone to instability (Fernandes et al., 2020).

In the Central Pyrenees, a period of glacial growth occurred during the OD after the massive deglaciation that took place following the LGM. Moraines dated at 17-15 ka have been reported in the highest valleys of the main massifs, such as in the Ariège (Delmas et al., 2011; Jomelli et al., 2020; Reixach et al., 2021), Ésera (Crest et al., 2017), Gállego (Palacios et al., 2015a), Malniu (Andrés et al., 2018; Palacios et al., 2015b; Pallàs et al., 2010), Têt (Delmas et al., 2008; Tomkins et al., 2018), Arànsér (Andrés et al., 2018; Palacios et al., 2015b), and Bassiès (Crest et al., 2017). During that time, glaciers flowed downslope from the headwaters of the highest valleys, with valley glaciers often exceeding 10-15 km (Oliva et al., 2019). Consequently, it is likely that, by that time, most of the Ruda Valley was still glaciated, with the ice tongue probably merging with the main Garonne Paleoglacier that was also fed by neighbouring glacier tributaries, such as those descending from the Beret area (Oliva et al., 2021a). Some of the lateral moraine remnants distributed on the left side of the Garonne Valley, ca. 20 km from the highest parts of the Saboredo Cirque, formerly attributed to the OD (Fernandes et al., 2021; Fernandes et al., 2017) may be also correlated with some of the highest moraines and erratic boulders distributed in the Ruda Valley (Fig. 3). Other Iberian mountain ranges include geomorphic evidence of glacial advances during the OD (Palacios et al., 2017a), such as in the Monasterio Valley in the Cantabrian Mountains (Serrano et al., 2016), Peña Negra Valley in the Iberian Range (García-Ruiz et al., 2020), Peñalara massif in the Central Range (Palacios et al., 2012a), San Juan Valley in the Sierra Nevada (Palacios et al., 2016), En Garcia Valley in the Eastern Pyrenees (Reixach et al., 2021), and Gállego and Ésera valleys in the Central Pyrenees (Crest et al., 2017; Palacios et al., 2017b).

No moraine remnants have been found in the lowest and central sections of the Ruda Valley, and the prevailing lithology of slates and marbles in this area is not favourable for the preservation of glacially abraded surfaces. Indeed, the neighbouring Bacivèr Cirque was completely deglaciated also by ~15-14 ka (Oliva et al., 2021a). Thus, our data cannot confirm whether small glaciers persisted in the highest parts of the Ruda Valley during the B-A, although CRE ages support climatic conditions during the early B-A unfavourable for the maintenance of the OD ice masses (Fig. 7). In other valleys of the Pyrenees, glaciers also receded rapidly at ~14.6-14 ka as shown by CRE ages from polished surfaces and scattered erratic boulders in the Ariège ( $14.4 \pm 1.1$ ,  $14.1 \pm 0.7$ ,  $14.0 \pm 0.8$  ka; Delmas et al., 2011), Gállego ( $14.6 \pm 2.3$  ka; Palacios et al., 2015a) and Ésera valleys ( $14.3 \pm 0.5$  ka; Crest et al., 2017). Locally, some massifs of the Pyrenees were fully deglaciated during the B-A, with small glaciers persisting only in the highest cirques (Delmas, 2015). A similar pattern was detected in other Iberian ranges, such as in the Sierra Nevada (Palacios et al., 2016), Cantabrian Mountains (Rodríguez-Rodríguez et al., 2017) and the Iberian Range (García-Ruiz et al., 2020), where ice masses shrank significantly by ~15-14 ka. Glacial retreat in the Iberian Mountains was triggered by the sudden temperature increase of 3-5 °C at the onset of the B-A in western Europe, as inferred from climate models (Renssen and Isarin, 2001). In parallel, significant shifts occurred in the Atlantic Meridional Overturning Circulation (Obase and Abe-Ouchi, 2019). In north Iberia, a maximum increase of the mean annual temperature of 7.5°C at the onset of GI-1 has been estimated in northern Iberia, based on the Ostolo cave speleothem isotopic record (Bernal-Wormull et al., 2021). Marine records from the western Iberian margin also pointed to a 5-6°C of warming at the onset of the B-A (Martrat et al., 2007). These warmer conditions also favoured the recession of glaciers across most European mountain ranges, such as in the Alps at 15.9-14.3 cal ka BP (Ivy-Ochs, 2015), Tatra Mountains at 14.8-14.2 ka (Zasadni et al., 2020), Taurus Mountains at  $14.6 \pm 2.8$  ka (Ciner and Sarikaya, 2015), as well as the Scandinavian Ice Sheet, which recorded a shrinking trend starting at 14.6 cal ka BP (Mangerud et al., 2013).

1  
2 463 However, the B-A was a climatically unstable period characterised by an alternation of warm and cold  
3 464 temperatures (Rasmussen et al., 2014). The gradual decrease of temperatures from the early to the late B-A  
4 465 was depicted by periods of enhanced cold conditions in the high latitudes of the Northern Hemisphere, which  
5 466 also favoured periods of glacial growth in the Pyrenees (Table 5). Indeed, CRE results indicate the formation  
6 467 of moraines in the Ruda Valley at  $13.5 \pm 0.9$  and  $13.0 \pm 0.8$  ka (Fig. 7). These CRE ages are consistent with  
7 468 the chronostratigraphic sequence but must be taken as approximate due to their uncertainty ranges and the low  
8 469 number of dated samples at each moraine. At that time, small ice tongues  $\sim 3$  km long flowed downslope from  
9 470 the main Saboredo Cirque as well as from other neighbouring glacial hollows to overdeepened basins, forming  
10 471 moraines at 2080 and 2190 m, respectively. Whereas evidence of phases of glacier advance or stillstands during  
11 472 the transition from the late B-A to the early YD were already found in other cirques in the Central Pyrenees,  
12 473 such as in the Bacivèr Cirque (Oliva et al., 2021a), and in the Cuerpo de Hombre Valley (Sierra de Gredos,  
13 474 Central Range), with moraine stabilization at 13.1 ka (Carrasco et al., 2015), this is the first study that reports  
14 475 geochronological evidence of a phase with glacial development in the second half of the B-A after the long-  
15 476 term cooling initiated at  $\sim 14.5$  ka (Fig. 7; Table 5). A few studies in Europe have also found evidence of glacier  
16 477 advances or stillstands during the middle B-A, such as in the Tatra Mountains where moraines stabilized at  
17 478  $13.4 \pm 0.5$  ka (Engel et al., 2017). By contrast, geomorphic evidence of a period of glacial expansion during  
18 479 the transition between the B-A to the early YD was detected in several regions, such as in the  
19 480 Hardangerfjorden-Herdla area, western Norway, where a till deposit suggested a glacier advance of the  
20 481 Scandinavian Ice sheet at 13.5-13.0 ka (Mangerud et al., 2016), and in the West Highland and Mull ice fields  
21 482 of the Scottish Highlands, where 18  $^{14}\text{C}$  dates from organic fragments found in basal tills revealed a glacial  
22 483 readvance from the end of the B-A or earliest YD (Bromley et al., 2018), although these results are considered  
23 484 controversial (Small and Fabel, 2016).

#### 29 485 Table 5

30  
31 486 Based on the position of these two moraine ridges, the ELA in the Ruda Valley glacier must have been located  
32 487 at 2461 and 2505 m during the glacial advances of the second part of the B-A and the late B-A or early YD,  
33 488 respectively. Considering that the regional ELA in the Pyrenees currently lies at approximately 3100 m  
34 489 (Jomelli et al., 2020; René, 2011), the ELA must have been located 639-595 m lower than present-day  
35 490 conditions. Assuming an average lapse rate of  $0.65 \text{ }^{\circ}\text{C } 100 \text{ m}^{-1}$  and no change in precipitation, we infer that  
36 491 summer temperatures must have been 4.2 and 3.9  $^{\circ}\text{C}$  lower than present-day during these two phases (Table  
37 492 4).

#### 40 493 Figure 7

41  
42 494 The temperature decrease during the early YD favoured the persistence of glaciers inside the highest hollows  
43 495 in the Ruda Valley, depositing moraines at  $\sim 2400$ -2500 m, which were later abandoned by the ice at  $12.6 \pm$   
44 496  $1.3$  ka (Fig 7). The existence of glaciers at high altitudes of the Saboredo Cirque is also confirmed by polished  
45 497 surfaces at elevations of 2200-2300 m that were dated at  $12.7 \pm 0.8$  ka. In the neighbouring Bacivèr Cirque,  
46 498 moraines at 2400 m were abandoned by the ice at 12.8 ka (Oliva et al., 2021a). The elevation difference  
47 499 between the ELA during the YD (2571 m) in the Ruda Valley and present-day values indicates that summer  
48 500 temperatures were  $\sim 3.4 \text{ }^{\circ}\text{C}$  lower during that cold phase (Table 4), which is very similar to the  $\sim 3.0 \text{ }^{\circ}\text{C}$  inferred  
49 501 for the Bacivèr Cirque (Oliva et al., 2021a). Indeed, topoclimatic factors must have played a key role  
50 502 maintaining small glaciers during the YD at the foot of steep cirque walls, where shading and snow  
51 503 accumulation allowed the longer persistence of reduced ice masses (Boston and Lukas, 2019). In the Pyrenees,  
52 504 YD glacial advances have been detected in cirques of the Central (Delmas, 2009; Delmas et al., 2008; García-  
53 505 Ruiz et al., 2016a) and Eastern sectors (Crest et al., 2017; Jomelli et al., 2020; Pallàs et al., 2010; Reixach et  
54 506 al., 2021), where 2-3.5 km-long glaciers formed at 12.9-11.2 ka. Based on chronostratigraphic data from a  
55 507 variety of sources, evidence of YD glacial advances was also recorded in the highest cirques of other Iberian  
56 508 massifs, although no direct ages are yet available (García-Ruiz et al., 2016a). This is true for the NW ranges  
57 509 (Cowton et al., 2009) and the Cantabrian Mountains (Pellitero et al., 2019; Serrano et al., 2015), where  
58 510 moraines above 1800 and 2000 m, respectively, were associated with this cold phase. In the Sierra Nevada,



1  
2 511 YD moraines dated at 13-12 ka are located in the highest cirques facing east from 2350 (north face) to 2800 m  
3 512 (south face) (Oliva et al., 2014; Palacios et al., 2016). These glacial advances were driven by a temperature  
4 513 cooling quantified at 2-4 °C in western Europe that was triggered by a slowdown of the overturning circulation  
5 514 in the North Atlantic (Renssen et al., 2018). In the Pyrenees, the YD was the coldest period of the last  
6 515 deglaciation recorded by a decrease of  $^{18}\text{O}$  in the Ostolo and Seso speleothems that might reflect the  
7 516 temperature decrease of  $\sim 5^\circ\text{C}$  (Bartolomé et al., 2015; Bernal-Wormull et al., 2021). Colder conditions also  
8 517 favoured glacial growth in other mountainous regions of Europe, with widespread advances across the Alps  
9 518 between 13.5 and 12 ka (Ivy-Ochs, 2015; Ivy-Ochs et al., 2009), in the Tatra Mountains where moraines  
10 519 stabilized at  $11.9 \pm 0.5$  ka (Engel et al., 2017) and in Turkey where they formed at  $12.6 \pm 2.3$  ka (Ciner and  
11 520 Sarikaya, 2015).

12  
13 521 The final deglaciation of the cirques favoured paraglacial dynamics and the formation of the numerous rock  
14 522 glaciers existing today in the Saboredo and Sendrosa cirques, as well as in the lateral cirques distributed along  
15 523 the Ruda Valley (Fig. 3). These features are currently inactive under the present-day climate regime (Fernandes  
16 524 et al., 2018), with abundant lichen coverage and forest colonization, particularly on the lowest ridges. In the  
17 525 Pyrenees, as well as in other Iberian ranges, cirque wall readjustment following deglaciation triggered different  
18 526 landforms, such as debris-covered glaciers or rock glaciers, that remained active until the early Holocene  
19 527 (Oliva et al., 2019). This is the case of the Bacivèr cirque, where the debris-covered glacier that formed inside  
20 528 the highest YD moraines stabilized at  $7.2 \pm 0.5$  ka (Oliva et al., 2021a), or of several other massifs in the  
21 529 Central and Eastern Pyrenees where rock glaciers were active until well into the Holocene (Andrés et al., 2018;  
22 530 García-Ruiz et al., 2016a). A similar pattern was also detected in other Iberian massifs, such as in the Iberian  
23 531 Range (Aumaître et al., 2017) or Sierra Nevada (Palacios et al., 2016), where these paraglacial landforms  
24 532 formed immediately after the deglaciation of the cirques, and remained active until the complete melting of  
25 533 their relict ice during the Holocene Thermal Maximum.  
26  
27

## 28 534 6. Conclusions

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30  
31 535 The Central Pyrenees were extensively glaciated during Quaternary glacial phases and were only slightly  
32 536 transformed by periglacial, nival and slope processes during subsequent interglacial periods such as the  
33 537 Holocene. As a result, the landscape in the Central Pyrenees includes a wide range of geomorphological  
34 538 features that constitutes a prominent natural heritage that contributed to the designation of most protected areas  
35 539 in this mountain range. This is the case of the Ruda Valley, located on the periphery of the National Park of  
36 540 Aigüestortes and Sant Maurici Lake. The valley is one of the most scenic glacial valleys of the Pyrenees, and  
37 541 includes a wide variety of glacial and periglacial landforms, particularly above 1800 m. However, in contrast  
38 542 to the southern slope of the Central Pyrenees where the timing of the glacial evolution has been widely studied,  
39 543 on the northern side there are still temporal and spatial gaps in the chronology of glacial activity. In this work,  
40 544 we determined the magnitude and timing of T-1 glacial oscillations in the Upper Garonne Basin by producing  
41 545 12 new CRE ages of glacial landforms (moraine boulders and glacially polished surfaces) that provide evidence  
42 546 of the spatio-temporal pattern of glacial oscillations from the B-A to the YD.  
43

44  
45 547 The Upper Garonne Basin hosted the largest glacial system of the Pyrenees during the last glacial cycle thanks  
46 548 to its Atlantic-influenced climate and the high altitude of the peaks in the central part of this range. The long-  
47 549 term post-LGM warming favoured glacial recession, particularly during the early B-A, and by  $14.4 \pm 1.2$  ka  
48 550 the Ruda Paleoglacier was already disconnected from the main Garonne Paleoglacier. At that time, the glacier  
49 551 front was located at 1860-1900 m, and some areas of the Saboredo Cirque were already ice-free as a result of  
50 552 ice thinning. The B-A was also a period of great climatic variability, and the small glaciers in the Ruda Valley  
51 553 showed rapid response times with formation of moraines during the coldest B-A phases in north Iberia. Phases  
52 554 of glacier advance or stillstands promoted the development of moraines at  $13.5 \pm 0.9$  ka, as well as during the  
53 555 transition between the late B-A and the onset of the YD at  $13.0 \pm 0.8$  ka. The YD probably favoured the  
54 556 persistence of small glaciers during an early stage of this cold phase; there is evidence of glacial retreat at the  
55 557 uppermost part of the cirque floor at  $12.7 \pm 0.7$  ka ( $\sim 2300$ -2350 m) and moraine abandonment of the highest



1  
2 558 cirque moraines (2470 m) at  $12.6 \pm 1.3$  ka. Subsequently, glacial disappearance must have promoted  
3 559 paraglacial dynamics and the formation of the (currently relict) rock glaciers existing within these moraines.  
4

5 560 Whereas the impact of the B-A deglaciation on the current landscape of the Pyrenean cirques was already  
6 561 known, the sequence of glacial advances and retreats during the B-A and YD presented in this study provides  
7 562 one of the most accurate deglacial histories in southern Europe. This opens new perspectives on the  
8 563 paleoclimatic evolution in the Iberian Peninsula, in particular, and the Mediterranean region in general. The  
9 564 period spanning from the B-A to the YD has been shown to be a major driver of transformation of the mountain  
10 565 landscapes of the Pyrenees and a key phase to better frame Holocene environmental dynamics in the highest  
11 566 lands of this mountain range. Future studies with new chronological data from other Pyrenean valleys and  
12 567 Iberian mountains will help to better understand T-1 glacier advances and retreats as well as the climate  
13 568 mechanisms behind millennial-scale glacial oscillations in southern Europe.

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## 37 585 7. References

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## List of figures

Figure 1. Geographical setting of the Ruda Valley. (A) Location in Europe. (B) Location in the Pyrenees with the distribution of the studies mentioned in the discussion (green stars). (C) Limits and topography of the valley with the distribution and view direction of the photos in Figure 2.

Figure 2. Examples of the main geomorphological landforms of glacial origin in the Ruda Valley, together with the location and typology of the dated samples: (A) terminal moraine damming a paleolake at 2080 m; (B) N-S perspective of the Saboredo Cirque floor showing several lakes distributed in staggered overdeepened basins; (C) polished bedrock surface including striations that reveal the former glacial flow; and (D) *roches moutonnées* located in the forested valley bottom of the valley at 1860 m.

Figure 3. (A) Geomorphological map of the study area, with the location of the CRE samples; and (B) enlargement of the highest parts of the cirque including the location of the  $^{10}\text{Be}$  ages.

Figure 4. Examples of sampled moraines and polished bedrock surfaces, together with CRE ages (ka).

Figure 5. Reconstruction of the glacial extent during the different time stages: (A) Extensive glaciers occupying the entire Upper Garonne Basin during the ILGM based on the reconstruction of [Fernandes et al. \(2017\)](#); (B) Glacial extent during the mid B-A at ~13.5 ka in the Ruda Valley; (C) Glacier flowing from the Saboredo Cirque during the transition between the late B-A and the early YD at ~13 ka; and (D) Small glaciers confined within the highest hollows next to the Sendrosa (D) and Saboredo (E) peaks during the early YD at ~12.7/12.6 ka.

Figure 6. Schematic evolution of the glacial phases recorded in the Ruda Valley during the last glacial cycle: (A) extensive and thick glaciers inundated the valley and only few nunataks protruded the ice during the ILGM; (B) warm temperatures in the early B-A favoured glacial recession, and subsequently the lowest areas of the valley as well as some rock outcrops of the cirque became exposed; (C) glacial advances/stillstands formed moraines during the B-A and at the transition to the YD; (D) small glaciers persisted during the early YD in the highest north-facing cirques; and (E) subsequent temperature increases favoured complete deglaciation and triggered paraglacial dynamics, which promoted the formation of rock glaciers in the recently deglaciated ice-free cirques.

Figure 7. Normalized probability distribution functions (PDF) of surface exposure ages vs. temperature evolution from the OD to the early Holocene based on the  $\delta^{18}\text{O}$  record from the NGRIP ice core from Greenland and the  $\delta^{18}\text{O}$  record from the speleothem of the Ostolo cave in north Iberia (time periods are defined after [Rasmussen et al. \(2014\)](#)). The plots of the units result from the sum of the individual PDF of the samples belonging to them; cold (warm) phases are represented by the blue (red) bands.

1 **List of tables**

2 **Table 1. Geographic location of samples, topographic shielding factor and sample thickness.**

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Sample name	Landform	Latitude (DD)	Longitude (DD)	Elevation (m a.s.l.) <sup>a</sup>	Topographic shielding factor	Thickness (cm)	Lithology	Size of the boulders largest axis (m)
<i>Valley polished surfaces</i>								
ARAN-40	Polished surface	42.6416	0.9713	1862	0.9630	3.7	Granodiorite	
ARAN-39	Polished surface	42.6393	0.9701	1900	0.9640	5.0	Granodiorite	
<i>Valley external moraines</i>								
ARAN-38	Moraine boulder	42.6296	0.9624	2079	0.9759	4.2	Granite	3.5
ARAN-37	Moraine boulder	42.6299	0.9620	2075	0.9752	3.0	Granite	3
<i>Valley internal moraines</i>								
ARAN-28	Polished surface	42.6263	0.9593	2188	1.0000	3.5	Granite	
ARAN-27	Moraine boulder	42.6257	0.9599	2184	0.9852	5.0	Granite	4
<i>Cirque polished surfaces</i>								
ARAN-36	Polished surface	42.6111	0.9681	2360	0.9806	3.5	Granite	
ARAN-35	Polished surface	42.6152	0.9630	2316	0.9875	3.2	Granite	
ARAN-34	Polished surface	42.6224	0.9596	2266	0.9901	5.0	Granite	
ARAN-31	Polished surface	42.6274	0.9536	2378	1.0000	3.0	Granite	
<i>Cirque moraines</i>								
ARAN-33	Moraine boulder	42.6275	0.9487	2470	0.9841	3.0	Granite	5
ARAN-32	Moraine boulder	42.6277	0.9480	2467	0.9789	4.0	Granite	6

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32 <sup>a</sup> Elevations are derived from the 5 m Digital Elevation Model of the Spanish "Instituto Geográfico Nacional" and are subjected to a vertical accuracy of ±5 m.

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**Table 2. AMS analytical data and calculated exposure ages.  $^{10}\text{Be}/^{9}\text{Be}$  ratios were inferred from measurements at the ASTER AMS facility. Chemistry blank correction was done by subtracting the number of  $^{10}\text{Be}$  atoms in the blank from that in the sample. Individual ages are shown with their full uncertainties (including analytical AMS uncertainty and production rate uncertainty) and analytical uncertainty only within brackets. Arithmetic mean ages are given with their full uncertainties (including standard deviation and production rate uncertainty) and standard deviations only in brackets.**

<b><math>^{10}\text{Be}</math> samples analytical AMS data</b>								
Sample name	Quartz weight (g)	Mass of carrier ( $^9\text{Be}$ mg)	ASTER AMS cathode number	$^{10}\text{Be}/^9\text{Be}$ ( $10^{-14}$ )	Blank correction (%)	$[^{10}\text{Be}]$ ( $10^4$ atoms $\text{g}^{-1}$ )	Age (ka)	Mean age (ka)
<i>Valley polished surfaces</i>								
ARAN-40	20.44	0.46	CHBL	$16.34 \pm 0.56$	2.11	$23.85 \pm 0.83$	$15.0 \pm 0.9$ (0.5)	$14.4 \pm 1.2$ (0.8)
ARAN-39	21.32	0.45	CHBK	$15.92 \pm 0.49$	2.18	$22.15 \pm 0.70$	$13.8 \pm 0.8$ (0.4)	
<i>Valley external moraines</i>								
ARAN-38	21.11	0.46	CHBJ	$17.56 \pm 0.54$	1.95	$24.99 \pm 0.79$	$13.3 \pm 0.8$ (0.4)	$13.5 \pm 0.9$ (0.2)
ARAN-37	21.38	0.46	CHBI	$18.08 \pm 0.87$	1.09	$25.74 \pm 1.25$	$13.7 \pm 1.0$ (0.6)	
<i>Valley internal moraines</i>								
ARAN-28	20.97	0.46	CHBB	$18.79 \pm 0.62$	1.05	$27.13 \pm 0.91$	$13.0 \pm 0.8$ (0.4)	$13.0 \pm 0.8$ (0.1)
ARAN-27	20.89	0.45	CHBA	$18.64 \pm 0.58$	1.85	$26.61 \pm 0.84$	$13.1 \pm 0.8$ (0.4)	
<i>Cirque polished surfaces</i>								
ARAN-36	21.51	0.46	CHBH	$21.24 \pm 0.65$	1.62	$29.53 \pm 0.92$	$12.7 \pm 0.8$ (0.4)	$12.7 \pm 0.8$ (0.0)
ARAN-35	21.43	0.47	CHBG	$20.18 \pm 0.62$	1.65	$29.05 \pm 0.91$	$12.8 \pm 0.8$ (0.4)	
ARAN-34	20.85	0.46	CHBF	$21.13 \pm 0.66$	1.63	$30.35 \pm 0.96$	$14.0 \pm 0.8$ (0.4)	
ARAN-31	21.40	0.44	CHBC	$26.54 \pm 0.72$	1.35	$35.74 \pm 0.98$	$14.8 \pm 0.9$ (0.4)	
<i>Cirque moraines</i>								
ARAN-33	20.26	0.45	CHBE	$20.30 \pm 0.68$	1.70	$29.92 \pm 1.02$	$11.8 \pm 0.7$ (0.4)	$12.6 \pm 1.3$ (1.0)
ARAN-32	21.63	0.45	CHBD	$24.19 \pm 1.26$	0.85	$33.15 \pm 1.75$	$13.3 \pm 1.0$ (0.7)	
<b>Chemistry blank details</b>								
Blank name	Processed with	mass of carrier ( $^9\text{Be}$ mg)	ASTER AMS cathode number	$^{10}\text{Be}/^9\text{Be}$ ( $10^{-14}$ )	$[^{10}\text{Be}]$ ( $10^4$ atoms)			
BK-1	ARAN-27, 31, 33, 34, 35, 36, 38, 39, 40	0.46	CHAT	$0.34 \pm 0.04$	$10.30 \pm 1.24$	-	-	
BK-2	ARAN-32	0.46	CHBM	$0.20 \pm 0.03$	$6.13 \pm 0.80$	-	-	
BK-3	ARAN-28, 37	0.46	IGHI	$0.20 \pm 0.03$	$5.98 \pm 0.93$	-	-	



**Table 3. Exposure ages according to erosion and snow cover corrections.**

Sample name	Exposure ages (arithmetic mean, in ka)			
	No correction	Erosion correction	Snow correction	Erosion + snow correction
<i>Valley - polished surfaces</i>	$14.4 \pm 1.2$	$14.6 \pm 1.2$	$15.5 \pm 1.3$	$15.7 \pm 1.3$
ARAN-40	$15.0 \pm 0.9 (0.5)$	$15.2 \pm 0.9 (0.5)$	$16.1 \pm 1.0 (0.5)$	$16.4 \pm 1.0 (0.6)$
ARAN-39	$13.8 \pm 0.8 (0.4)$	$14.0 \pm 0.8 (0.4)$	$14.8 \pm 0.9 (0.4)$	$15.0 \pm 0.9 (0.5)$
<i>Valley - external moraines</i>	$13.5 \pm 0.9$	$13.7 \pm 0.9$	$14.5 \pm 1.0$	$14.7 \pm 1.0$
ARAN-38	$13.3 \pm 0.8 (0.4)$	$13.5 \pm 0.8 (0.4)$	$14.4 \pm 0.9 (0.4)$	$14.6 \pm 0.9 (0.4)$
ARAN-37	$13.7 \pm 1.0 (0.6)$	$13.8 \pm 1.0 (0.7)$	$14.7 \pm 1.0 (0.7)$	$14.9 \pm 1.0 (0.7)$
<i>Valley - internal moraines</i>	$13.0 \pm 0.8$	$13.2 \pm 0.8$	$14.1 \pm 0.9$	$14.2 \pm 0.9$
ARAN-28	$13.0 \pm 0.8 (0.4)$	$13.1 \pm 0.8 (0.4)$	$14.0 \pm 0.9 (0.5)$	$14.2 \pm 0.9 (0.5)$
ARAN-27	$13.1 \pm 0.8 (0.4)$	$13.3 \pm 0.8 (0.4)$	$14.1 \pm 0.8 (0.4)$	$14.3 \pm 0.9 (0.4)$
<i>Cirque - polished surfaces</i>	$12.7 \pm 0.7$	$12.9 \pm 0.8$	$13.7 \pm 0.8$	$13.9 \pm 0.8$
ARAN-36	$12.7 \pm 0.8 (0.4)$	$12.9 \pm 0.8 (0.4)$	$13.7 \pm 0.8 (0.4)$	$13.9 \pm 0.8 (0.4)$
ARAN-35	$12.8 \pm 0.8 (0.4)$	$12.9 \pm 0.8 (0.4)$	$13.8 \pm 0.8 (0.4)$	$14.0 \pm 0.8 (0.4)$
ARAN-34	$14.0 \pm 0.8 (0.4)$	$14.2 \pm 0.9 (0.4)$	$15.1 \pm 0.9 (0.5)$	$15.3 \pm 0.9 (0.5)$
ARAN-31	$14.8 \pm 0.9 (0.4)$	$15.1 \pm 0.9 (0.4)$	$16.0 \pm 0.9 (0.4)$	$16.2 \pm 0.9 (0.4)$
<i>Cirque - moraines</i>	$12.6 \pm 1.3$	$12.7 \pm 1.4$	$13.5 \pm 1.4$	$13.7 \pm 1.5$
ARAN-33	$11.8 \pm 0.7 (0.4)$	$12.0 \pm 0.8 (0.4)$	$12.7 \pm 0.8 (0.4)$	$12.9 \pm 0.8 (0.4)$
ARAN-32	$13.3 \pm 1.0 (0.7)$	$13.5 \pm 1.0 (0.7)$	$14.3 \pm 1.0 (0.7)$	$14.5 \pm 1.0 (0.7)$

**Table 4. Reconstructed ELAs (m a.s.l.) for the middle, late B-A and YD moraines using the AAR and AABR methods.**

Phase	Glacier	AAR (0.6±0.2)	AABR (1.9±0.81)	AABR (1.75±0.71)	Average ELA
Valley external moraines (middle B-A; GI-1c)	Ruda	2481 ± 20	2449 -30/+20	2454 -30/+15	2461
Valley internal moraines (transition B-A/YD; GI-1b)	Ruda	2515 -15/+10	2498 -25/+15	2503 -25/+20	2505
Cirque moraines (early YD)	Sendrosa	2568 -10/+5	2571 -15/+5	2576 ± 10	2571

**Table 5. Glacial response in the Ruda Valley during the B-A Interstitial according to the reconstructed climate in Greenland and in Iberian Peninsula.**

Phase (ages in ka)	Climate trend in Greenland based on ice core records (Rasmussen et al., 2014)	Glacier response in the Ruda Valley	Climate in the Iberian Peninsula based on other independent proxies
GI-1e (14.7-14.1)	Abrupt warming (3-5°C)	Rapid recession at ~15-14 ka	Gradual warming in the northern Iberia at 15.4-13.4 cal ka BP (Moreno et al., 2010). Warming conditions in the Pyrenees at 14.7-14.0 cal ka BP (González-Sampérez et al., 2006).
GI-1d (14.1-14.0)	Short cold reversal (1-2°C)		
GI-1c (14.0-13.3)	Significant warming at the beginning (2-4°C) followed by a gradual cooling at (~2°C). Intense cold conditions at 13.7-13.6 ka	Glacier advance or stillstand with moraine formation at 13.5 ka	Colder conditions starting at 13.5 cal ka BP in northern Iberia (Moreno et al., 2010). Cold climate in the Pyrenees at 14-13.4 cal ka BP (González-Sampérez et al., 2006).
GI-1b (13.3-13.1)	Short cold reversal (1-2°C)	-	Short periods of colder and drier conditions (Moreno et al., 2010). Expansion of cold-climate forest species in the Pyrenees at 13.4-13.1 cal ka BP (González-Sampérez et al., 2006).
GI-1a (13.1-12.9)	Early warming followed by a general cooling	Glacier advance or stillstand with moraine formation at 13.0 ka	Cold and dry conditions in the Pyrenees starting at 13.1 cal ka BP (González-Sampérez et al., 2006).

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2 1 **Glacial oscillations during the Bølling-Allerød Interstadial-Younger Dryas**  
3 2 **transition in the Ruda Valley, Central Pyrenees**  
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## Abstract

The Upper Garonne Basin included the largest glacial system in the Pyrenees during the last glacial cycle. Within the long-term glacial retreat during Termination-1 (T-1), glacier fluctuations left geomorphic evidence in the area. However, the chronology of T-1 glacial oscillations on the northern slope of the Central Pyrenees is still poorly constrained. Here, we introduce new geomorphological observations and a 12-sample dataset of  $^{10}\text{Be}$  Cosmic-Ray Exposure (CRE) ages from the Ruda Valley. This U-shaped valley, surrounded by peaks exceeding 2800 m a.s.l., includes a sequence of moraines and polished surfaces that enabled to reconstruct the chronology of the last deglaciation. Following the maximum ice extent, warmer conditions prevailing at ~15-14 ka, during the Bølling-Allerød (B-A) Interstadial, favoured glacial retreat in the Ruda Valley. Within the B-A, glaciers experienced two phases of advance/stillstand with moraine formation at 13.5 and 13.0 ka. During the early YD, glacial retreat exposed the highest surfaces of the Saboredo Cirque (~2300-2350 m) at 12.7 ka. Small glaciers persisted only inside the highest cirques (~2470 m), such as in Sendrosa Cirque, with moraines stabilizing at 12.6 ka. The results of this work present the most complete chronology for Pyrenean glacial oscillations from the B-A to the YD.

**Key words:** Central Pyrenees, Bølling-Allerød, Younger Dryas, Cosmic-Ray Exposure dating, deglaciation

## 1. Introduction

The present-day landscape of the highest mid-latitude mountain environments is dominated by the effects of glacial dynamics that prevailed during the last glacial cycle and of post-glacial processes that reshaped glacial landforms during the subsequent deglaciation. Following the end of the Last Glacial Maximum (LGM) at 19-20 ka (Clark et al., 2009), most glaciated mountain regions underwent a rapid deglaciation only interrupted by brief periods of glacial growth. Glacial disappearance favoured paraglacial readjustment and intense periglacial and slope processes, which reshaped the already deglaciated valleys and cirques (Ballantyne, 2008). The current spatial distribution of glacial and periglacial features in the high elevations of most mid-latitude ice-free mountain ranges – such as the Pyrenees, where this research focuses on – results from the climatic shifts during Termination-1 (hereafter T-1), the period spanning from 20-19 to 11.7 ka (Denton et al., 2014).

At present, mid-latitude mountains such as those of the Mediterranean region are only sparsely glaciated, with small glaciers located in sheltered cirques (Hughes, 2014). However, there are several other mountain ranges located between 30 and 50° latitude North, such as the Alps or the Himalaya, which are still heavily glaciated and record recent dramatic glacial recession trends (Huss et al., 2021). An accurate comprehension of the climatic sensitivity of current glaciated mountain landscapes can help to understand the timing and magnitude of glacial change during the rapid climate shifts of T-1. The alternation of millennial-scale cold/warm periods during T-1 promoted a temporal pattern of glacial advances and retreats across the Northern Hemisphere (Allard et al., 2021; Rea et al., 2020). Ice core records from the Greenland interior show the following succession of climatic phases in the North Atlantic region during T-1: a cold climate regime during the Oldest Dryas (OD; GS-2.1a: 17.5-14.6 ka), followed by the general warm Bølling-Allerød (B-A) Interstadial (GI-1: 14.6-12.9 ka), with a subsequent intensification of cold conditions during the Younger Dryas (YD; GS-1: 12.9-11.7 ka) (Rasmussen et al., 2014).

In the Iberian mountains, recent advances in constraining the chronology of glacial activity during T-1 using Cosmic-Ray Exposure (CRE) dating have revealed contrasting spatio-temporal patterns (García-Ruiz et al., 2016b; Oliva et al., 2019; Palacios et al., 2017a). Whereas the maximum ice extent occurred asynchronously, glacial oscillations during T-1 show a more synchronous response: colder phases favoured glacial growth, whereas warm periods promoted glacial recession (Oliva et al., 2019) and the evolution from debris-free glaciers to rock glaciers such as those in the Iberian Range (García-Ruiz et al., 2020), Cantabrian Mountains (Rodríguez-Rodríguez et al., 2016) and Central (Palacios et al., 2017a, b) and Eastern Pyrenees (Andrés et al., 2018). To date, there is geomorphic evidence of glacial advances during the OD and YD, as well as of glacial retreat during the B-A and following the YD in the Cantabrian Mountains (Rodríguez-Rodríguez et al., 2017), Central Range (Carrasco et al., 2015; Palacios et al., 2012b), and Sierra Nevada (Gómez-Ortiz et al., 2012; Palacios et al., 2016). However, CRE ages require a careful interpretation and support of geomorphological criteria, as their uncertainty ranges may be greater than the time spanning the T-1 phases (Oliva et al., 2021a).

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2 93 The Pyrenees were the most extensively glaciated mountain system during the maximum ice extent of the last  
3 94 glaciation in Iberia, hosting ~50% of the peninsula's glacial surface (Oliva et al., 2021b). The chronology of  
4 95 glacier fluctuations during T-1 seems to have followed a similar temporal pattern, although available data is  
5 96 principally derived from the range's southern slope (Andrés et al., 2018; Palacios et al., 2015b, 2017b, 2017a).  
6 97 Pyrenean glaciers readvanced to the valley bottoms during the OD, with ice tongues > 15 km long. During the  
7 98 B-A, by 15-14 ka, high temperatures favoured rapid deglaciation and glaciers disappeared from relatively low  
8 99 cirques shaped under peaks of ~2600 m a.s.l. (Oliva et al., 2021a). A subsequent readvance occurred during  
9 100 the YD, with smaller glaciers of only ~4 km length confined within the cirques of the highest massifs above  
10 101 2200 m (Crest et al., 2017; García-Ruiz et al., 2016a; Jomelli et al., 2020; Palacios et al., 2017b, 2015a; Pallàs  
11 102 et al., 2006). Following both cold periods, glacial retreat triggered intense paraglacial dynamics at ~15-14 ka  
12 103 and ~11-10.5 ka, respectively, which led to the formation of rock glaciers and debris-covered glaciers,  
13 104 particularly in cirques below 2800 m (Andrés et al., 2018; Palacios et al., 2017b).

15 105 The timing of glacial oscillations is still poorly constrained on the northern slope of the Pyrenees, such as in  
16 106 the Upper Garonne Basin where this work focuses on. Recently, Reixach et al. (2021) have summarized the  
17 107 glacial evolution from 19 to 12 ka in the northern Pyrenees, reporting a spatial and temporal pattern for T-1  
18 108 glacial oscillations similar to that on the southern slopes of the range. Jomelli et al. (2020) confirmed glacial  
19 109 growth during the YD in the Ariège Valley, where glaciers persisted in the highest cirques until the early  
20 110 Holocene. Another recent study that focused on the Bacivèr Cirque, near our study area, revealed a rapid  
21 111 deglaciation during the B-A, followed by glacial readvance and the formation of the highest moraines at the  
22 112 B-A/YD transition at ~12.9 ka; subsequently, intense paraglacial readjustment favoured the formation of  
23 113 debris-covered glaciers and rock glaciers inside these moraine systems (Oliva et al., 2021a).

25 114 In order to complement our understanding of the glacial history of the Pyrenees, this study uses new data to  
26 115 improve the deglaciation chronology of the Upper Garonne Basin, the largest glacial system in this mountain  
27 116 range, during the local last glacial maximum (ILGM; Fernandes et al., 2017). Glacial retreat in the Ruda Valley,  
28 117 where this work focuses on, left widespread geomorphic evidence of glacial and paraglacial origin that can be  
29 118 used to reconstruct the last stages of long-term deglaciation in the Central Pyrenees.

31 119 Our study, therefore, has the following objectives:

- 32 120 - To date the glacial landforms of the Upper Garonne Basin.
- 33 121 - To discuss the T-1 deglaciation in the Upper Garonne Basin and its geomorphological implications in  
34 122 the context of the known chronologies from the Pyrenees southern slope and from other Iberian and  
35 123 European mountain ranges.
- 36 124 - To examine the paleoclimatic and paleoenvironmental significance of the glacial response of the  
37 125 northern slope of the Pyrenees and compare it with the paleoclimatic evolution inferred from other  
38 126 records in the North Atlantic region.

## 42 128 2. Study area

43 129 The Pyrenees stretch E-W over 400 km across north Iberia, with the highest peaks (> 3000 m) in the central  
44 130 part of the range. This study focuses on the Ruda Valley, located between latitudes 42°40' N - 42°42' N and  
45 131 longitudes 0°57' E - 1°00' E (Fig. 1), that constitutes the headwaters of the Upper Garonne Basin. The area is  
46 132 included in the peripheral protection area of the National Park of Aigüestortes and Sant Maurici Lake.

### 49 133 Figure 1

50  
51 134 The Upper Garonne Valley shows the typical U-shaped cross-section of glacial valleys, and the Garonne River  
52 135 flows to the N-NW through the central axis of the Pyrenees. Several tributary rivers drain the glacial valleys  
53 136 descending from glacial cirques carved below peaks from 2800 to 3000 m. Glacial cirques are prevalingly  
54 137 NE-oriented, with floors generally between 2200 and 2400 m (Lopes et al., 2018).

56 138 The U-shaped Ruda Valley, drained by the Ruda River that runs 12 km northwards, constitutes the headwaters  
57 139 of the Upper Garonne Basin. Its catchment encompasses 34 km<sup>2</sup>, with an elevation range of ~1300 m between  
58 140 Saboredo Peak (2827 m) and the village of Baqueira (~1500 m). The Saboredo Cirque is located in the  
59 141 headwaters of the Ruda Valley; the cirque is N-exposed, extending over 5.7 km<sup>2</sup>, with its floor at ~2250-2400



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2 142 m. Saboredo (2827 m), Sendrosa (2702 m) and Locampo (2658 m) peaks are the highest elevations of the  
3 143 cirque, while its toe is located at ~2250 m.

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5 144 The present-day mean annual air temperature (MAAT) at Port de la Bonaigua station (2266 m) is 3 °C, with  
6 145 the regional 0 °C isotherm in the Central Pyrenees estimated to be at 2950 m (López-Moreno et al., 2019).  
7 146 Annual precipitation reaches 1227 mm and falls mostly as snow (average depth of 1 m) from October to May,  
8 147 with a snow duration of 7-8 months in the cirque floor. The combination of low temperatures and abundant  
9 148 snowfall determines the scarce vegetation cover above the treeline (2200-2300 m), where barren rocky terrain  
10 149 is widespread, together with small alpine meadows.

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13 150 The landscape of the Ruda Valley resulted from the interaction of tectonics and vertical incision, both during  
14 151 Quaternary glaciations and interglacial periods, the latter dominated by alluvial and mass-wasting processes  
15 152 (Fernandes et al., 2017). Glacial and periglacial landforms are preserved on different lithologies across the  
16 153 valley, composed of Carboniferous granodiorites and granites in the highest sections, with Cambrian-Devonian  
17 154 metamorphic (hornfels, marbles, slates, quartzites, schists) and sedimentary rocks (limestones, sandstones,  
18 155 conglomerates and lutites) in the lowest parts of the valley (ICGC, 2017; Quesada and Oliveira, 2019).  
19 155 However, despite abundant geomorphic evidence revealing that the Garonne Valley presented the longest  
20 156 glacier in the Pyrenees during the ILGM (ca. 80 km; Fernandes et al., 2017), the chronological framework of  
21 157 glacial activity in the entire Upper Garonne Basin is still unknown. The headwaters of the Ruda Valley include  
22 158 a rich sequence of moraine systems and polished bedrock surfaces – as well as relict rock glaciers and protalus  
23 159 lobes (Fernandes et al., 2018) – that can be used to improve the temporal resolution for the chronosequence of  
24 160 the final deglaciation of the upper Garonne Valley.  
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### 29 163 3. Methodology

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31 164 We reconstructed the last deglaciation chronology of the Ruda Valley in the Upper Garonne Basin by  
32 165 integrating geomorphological and chronological techniques. Field work was conducted during the summer of  
33 166 2019, when snow cover was limited to the concave and sheltered areas at the highest elevations (>2500 m),  
34 166 which enabled better identification of the glacial landforms, as well as sampling for CRE dating.  
35 167

#### 36 168 3.1 Geomorphological mapping

37 168  
38 169 We produced a geomorphological map focusing on glacial and periglacial landforms based on recent general  
39 170 geomorphological maps from the area (Fernandes et al., 2017; Fernandes et al., 2021). Here we used the same  
40 171 approach based on the combination of aerial orthophotographs (0.25 m cell size) and high-resolution LIDAR  
41 171 digital elevation models (point density of 0.5-2 m<sup>-2</sup>) from the “Institut Cartogràfic i Geològic de Catalunya”  
42 172 (<http://www.icc.cat/appdownloads>). The symbology of geomorphological landforms followed Joly (1997).  
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#### 45 174 3.2 Field strategy and sampling

46 175 We collected 12 granite and granodiorite samples for CRE dating, using a hammer and chisel, from elevations  
47 176 spanning from 1862 to 2470 m (Table 1). To ensure that samples were intact since glacial retreat, we targeted  
48 176 at well-anchored moraine boulders and glacially polished bedrock that stood out within their surroundings,  
49 177 which minimizes the risk of having been buried under sediments (Fig. 2). To ensure the best cosmic-ray flux  
50 178 reception of the sampling sites, we restricted sample extraction to flat-topped and gentle surfaces, and sampled  
51 179 a thickness of up to 3-5 cm. Corrections for partial shielding due to the surrounding topography were  
52 180 implemented using the ArcGIS toolbox by Li (2018), which applies equations describing the effects of  
53 181 geometric shielding on the sampling sites and a model to predict the nuclide production rate at depth in sloped  
54 182 surfaces (Dunne et al. 1999); it required only a point shapefile of the sampling sites including strike/dip of the  
55 183 sampling surfaces and a digital elevation model of 5 m spatial resolution.  
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59 185 Table 1

60 186 Figure 2

### 3.3 Laboratory procedures and exposure age calculation

Samples were crushed and sieved (>200 g) at the Physical Geography Laboratory of the Universidad Complutense de Madrid (Spain) to the 125-500  $\mu\text{m}$  fraction, which was subsequently processed at the ‘Laboratoire National des Nucléides Cosmogéniques’ (LN<sub>2</sub>C) of the ‘Centre Européen de Recherche et d’Enseignement des Géosciences de l’Environnement’ (CEREGE; Aix-en-Provence, France). In accordance with the quartz-rich lithology of the samples, they were treated for the extraction of the *in-situ*-produced cosmogenic nuclide <sup>10</sup>Be.

First, magnetic minerals were discarded through magnetic separation conducted in a “Frantz LB-1” magnetic separator. Once the non-magnetic fraction was isolated, it underwent several rounds of chemical attacks with a 1:2 mixture of concentrated hydrochloric (HCl) and hexafluorosilicic (H<sub>2</sub>SiF<sub>6</sub>, 2/3) acids to dissolve and discard the non-quartz minerals. The remaining minerals (i.e. non-dissolved feldspar minerals) were subsequently attacked with 4-6 successive partial dissolutions with HF to remove the atmospheric <sup>10</sup>Be and discard the remaining impurities. In order to be sure that the quartz was pure, samples were examined under a binocular microscope. As a result, ~20 g of pure quartz per sample was used to extract the <sup>10</sup>Be (Table 2). After adding ~150  $\mu\text{L}$  of an in-house manufactured (from a phenakite crystal) <sup>9</sup>Be carrier solution (spike, concentration:  $3025 \pm 9 \mu\text{g g}^{-1}$ ; Merchel et al., 2008), the purified quartz was dissolved with 48% concentrated HF (3.5 mL per g of quartz + 30 mL in excess). Following total dissolution, the resulting solution was evaporated until dryness, and the residue was recovered with HCl (7.1 molar). Be was then precipitated at pH=8 to beryllium hydroxide (Be(OH)<sub>2</sub>) by means of ammonia (NH<sub>3</sub>), and separated from other elements in resin columns: a Dowex 1x8 anionic exchange column to remove elements such as Fe, Mn and Ti, and a Dowex 50W x8 cationic exchange column to discard B and separate Be from Al (Merchel and Herpers, 1999). The final eluted Be was again precipitated, dried and oxidized to BeO at 700 °C. Finally, the targets for the accelerator mass spectrometer (AMS) measurements were prepared by mixing with niobium powder at an approximate 1:1 proportion and pressing the mixture into copper cathodes.

The targets were analysed at the ‘Accelerator pour les Sciences de la Terre, Environnement et Risques’ (ASTER) national AMS facility at CEREGE in order to measure the <sup>10</sup>Be/<sup>9</sup>Be ratio from which the <sup>10</sup>Be concentration was later inferred (Table 2). AMS measurements were calibrated against the in-house standard STD-11 with an assigned <sup>10</sup>Be/<sup>9</sup>Be ratio of  $(1.191 \pm 0.013) \times 10^{-11}$  (Braucher et al., 2015). Based on calibration with the previous ASTER AMS standard NIST-27900 (SRM4325), the new STD-11 is fully compatible with 07KNSTD. The analytical 1 $\sigma$  uncertainties include uncertainties in the AMS counting statistics, an external 0.5% AMS error (Arnold et al., 2010) and the uncertainty related to the chemical blank correction. The <sup>10</sup>Be half-life considered was  $(1.3870 \pm 0.0012) \times 10^6$  years (Chmeleff et al., 2010; Korschinek et al., 2010).

#### Table 2

We calculated <sup>10</sup>Be exposure ages by using the CREP online calculator (Martin et al., 2017; available online at: <http://crep.cirp.cnr.fr/#/>), where we selected the following settings: LSD (Lifton-Sato-Dunai) elevation/latitude scaling scheme (Lifton et al., 2014), ERA40 atmospheric model (Uppala et al., 2005) and geomagnetic database based on the LSD framework (Lifton et al., 2014). These settings yielded a world-wide mean sea level high latitude (SLHL) <sup>10</sup>Be production rate of  $3.98 \pm 0.22 \text{ atoms g}^{-1} \text{ yr}^{-1}$ . Exposure ages of the samples with their 1 $\sigma$  full and analytical uncertainties are shown in Table 2. The uncertainties discussed throughout the text include analytical and production rate error unless otherwise stated.

In order to evaluate the potential impact of erosion on the exposure ages, we assumed a steady erosion rate (1 mm ka<sup>-1</sup>) following a conservative maximum value in such a lithological setting according to André (2002). The impact of snow cover was explored assuming that snow currently remains in the area an annual average of 7.5 months at 2200-2300 m with a mean thickness of 100 cm – and assuming a snow density of 0.2 g cm<sup>-3</sup> (Styllas et al., 2018) – (Table 3), based on the available data from the Bonaigua weather station (Fig. 1) available since 1997-1998 (Servei Meteorològic de Catalunya; [http://www.igc.cat/web/ca/allaus\\_gruix\\_neu\\_v2.php?e=bonaigua&t=totes](http://www.igc.cat/web/ca/allaus_gruix_neu_v2.php?e=bonaigua&t=totes)). Erosion and snow corrections, as

a whole, resulted in ages that were ~9% older (~1% erosion; ~8% snow cover). However, to enable comparisons with other areas, we use the non-corrected ages throughout the text. Moreover, both the thickness and duration of snow cover must have varied significantly since the cirque's deglaciation, given the alternation of cold/warm periods and their associated moisture supplies. Finally, the chi-squared test according to Ward and Wilson (1978) was applied to the samples belonging to each geomorphological unit established in the geomorphological map (Table 1) in order to detect potential exposure age outliers (i.e. "too old" or "too young"; Heyman et al., 2011). No outliers were detected (Table 3).

### Table 3

## 3.4 Glacier reconstruction and Equilibrium-Line Altitude (ELA) calculation

A three-dimensional paleoglacier reconstruction was performed for the different glacial phases using the 'GLaRe' ArcGIS toolbox devised by Pellitero et al. (2016). Former ice thickness was estimated by applying a perfect-plasticity physical-based rheological model along flowlines from the termini to the headwall (Van der Veen, 1999; Benn and Hulton, 2010). Such a toolbox requires only a flowline, a tentative paleoglacier geometry - approximated from lateral or frontal moraines - and a digital elevation model. Ice thickness was modelled by using a constant shear stress of 100 kPa (Paterson, 1994; Benn and Hulton, 2010). The effect of lateral side drag was corrected by integrating shape factors (F-factor) obtained from cross-sections (Schilling and Hollin, 1981).

ELAs were calculated by using the automatic toolbox 'ELA calculation' developed by Pellitero et al. (2015). The selected methods were the Accumulation Area Ratio (Porter, 1975; AARs:  $0.6 \pm 0.05$ ) and the Area Altitude Balance Ratio (AABR; Osmaston, 2005). We considered two balance ratios for the AABR method:  $1.75 \pm 0.71$  (global) and  $1.9 \pm 0.81$  (mid-latitude marine glaciers) (Rea, 2009). The average ELAs were used to discuss the paleoclimatic implications of the identified glacier advance/stillstand phases.

## 4. Results

The Ruda Valley preserves glacial and periglacial landforms indicative of glacial evolution during the last deglaciation in the Upper Garonne Basin. The chronological framework of this evolution is unveiled by a dataset of 12  $^{10}\text{Be}$  CRE ages obtained from glacial landforms.

### 4.1 Geomorphological evidence

The wide range of glacial features from the highest sectors to the Ruda Valley floor shows evidence that the entire valley was heavily glaciated during the ILGM and recorded several advances or stillstands within the long-term deglaciation trend. The Ruda Valley is divided into two main geomorphological units (Fig. 3):

- (i) The valley floor and mountain slopes follow the typical U-shaped morphology of a glacial valley. The steep slopes, descending from ~2200 to 1800 m, connect the lateral cirques with the valley floor. The hillslopes of the valley start at the upper edge of hanging glacial cirques that define the glacial trimline, marking the uppermost extent of recent glaciations. Hillslopes are concave to rectilinear 300-400 m-long sections steeper than  $30^\circ$  that are mostly colonized by forest (i.e. *Pinus nigra*). The upper part of the valley is defined by two staggered, overdeepened basins filled with peatlands at average elevations of 2150 and 2070 m. At their margins, two moraine systems occur at 2190 and 2080 m. Two samples were collected from boulders from the external frontal moraine (ARAN-38 and ARAN-37) and the other two were obtained from the internal moraine unit: one from a boulder on the moraine ridge (ARAN-27), and the other from a polished surface with glacial striae (ARAN-28) and moraine boulders (Fig. 4). The lowest part of the valley floor, with a relatively flat slope characteristic of U-shaped glacial valleys and descending from ~1800 to 1500 m, is colonized by alpine meadows and coniferous forests. Isolated *roches moutonnées* are found in the highest section of the valley bottom where granites prevail ( $> \sim 1800$  m), while they are absent in lower areas, where slates and marbles are the dominant lithology. Two samples were collected from polished surfaces at 1900 and 1860 m (ARAN-39 and ARAN-40).

(ii) The highest peaks of the Ruda Valley surrounding the Saboredo Cirque range from 2700-2800 m (Sendrosa and Saboredo peaks) to 2525 m in the mountain divide that may have functioned as a glacial diffluence pass between the Saboredo (Garonne basin) and the Ratera cirques (Noguera Pallaresa basin) (Fig. 3). Saboredo is a complex cirque (Barr and Spagnolo, 2015) that includes several minor hollows carved out on the marginal flanks of the peaks. They are separated by truncated spurs with polished surfaces, where one sample (ARAN-31) was collected to examine the timing of glacial thinning at the cirque level. The highest moraines in the Ruda Valley are located at ~2400-2500 m, at the foot of the highest peaks. A prominent cirque moraine at an altitude of 2470 m is located on the eastern slope of Sendrosa Peak; here, we collected two samples (ARAN-33 and ARAN-32). Above the upper moraines, fed by abundant debris supply from the cirque rock walls, rock glaciers and protalus lobes are found. The cirque floors extend between 2200 and 2400 m and preserve features indicative of intense glacial erosion (e.g. overdeepened basins or glacial thresholds) and glacial deposits (e.g. moraines or erratic boulders). In the Saboredo Cirque, glacial abrasion and periglacial processes shaped a large amphitheatre with several overdeepened basins. They are surrounded by polished surfaces and occupied by peatlands and lakes. Three samples were collected from polished surfaces from 2360 to 2270 m (Fig. 3) (ARAN-36; ARAN-35; ARAN-34; Fig. 2; Table 1). Glacial retreat left scattered boulders and till across the area.

Figure 3

Figure 4

#### 4.2 Geochronological data

The 12 CRE samples collected from the Ruda Valley returned a chronology spanning from  $\sim 15 \pm 0.9$  to  $11.8 \pm 0.7$  ka, with all ages being consistent with the geomorphological distribution of the dated surfaces (Fig. 4).

Polished surfaces from the highest parts of the Ruda Valley located ~200-250 m below the cirques provided CRE ages of  $15.0 \pm 0.9$  (ARAN-40) and  $13.8 \pm 0.8$  ka (ARAN-39, Fig. 4A) (average:  $14.4 \pm 1.2$  ka). A similar age was obtained from a sample collected from a truncated spur dividing the Sendrosa and Saboredo cirques, which yielded  $14.8 \pm 0.9$  ka (ARAN-31). A sample collected from a polished surface at the edge of the Saboredo cirque floor (Fig. 3) reported an age of  $14.0 \pm 0.8$  ka (ARAN-34, Fig. 4D).

Moraine boulders from the external moraine located in the lower overdeepened basin yielded ages of  $13.3 \pm 0.8$  (ARAN-38) and  $13.7 \pm 1.0$  ka (ARAN-37) (average:  $13.5 \pm 0.9$  ka). A boulder from the internal moraine ridge located 110 m above the former moraine provided an age of  $13.1 \pm 0.8$  ka (ARAN-27); this age was compared with a sample from a polished surface which correlates with the internal moraine (Fig. 4C) that returned an age of  $13.0 \pm 0.8$  ka (ARAN-28), which agrees with the latter.

Two samples from polished surfaces of the Saboredo Cirque above 2300 m yielded ages of  $12.8 \pm 0.8$  (ARAN-35) and  $12.7 \pm 0.8$  ka (ARAN-36) ( $12.7 \pm 0.8$  ka). The highest moraines are located inside the highest cirque hollows, such as the Sendrosa hollow. Here, two moraine boulders yielded ages of  $11.8 \pm 0.7$  (ARAN-33) and  $13.3 \pm 1.0$  ka (ARAN-32) ( $12.6 \pm 1.3$  ka).

#### 4.3 Reconstructed paleoglaciers and ELAs

The ELAs of the three reconstructed paleoglacier extents were calculated for the Ruda catchment. The paleoglacier existing at  $13.5 \pm 0.9$  ka deposited a moraine system at 2080 m, with an average ELA located at 2461 m (Table 4). The paleoglacier at  $13.0 \pm 0.8$  ka formed a moraine at 2190 m, with an average ELA at 2505 m. Finally, the Sendrosa Cirque moraine located at 2470 m was dated at  $12.6 \pm 1.3$  ka, with an ELA of 2571 m (Table 4).

Table 4

### 5. Discussion

#### 5.1 Interpretation of the geomorphological observations and CRE results



The geomorphological analysis supported by  $^{10}\text{Be}$  dating of glacial landforms show the spatial and temporal patterns of the deglaciation between the B-A and the early YD in the Ruda Valley (Fig. 5). Although the CRE dataset is consistent with sequence of glacial landforms, uncertainty remains in assigning ages either to the late B-A or to the early YD.

The ages of samples from polished surfaces at the valley at 1860-1900 m (ARAN-40 and ARAN-39;  $14.4 \pm 1.2$  ka) suggest that the beginning of glacial retreat in the Ruda Valley took place during the B-A. As glaciers retreated, they became confined within their respective cirques, as revealed by the age of  $14.8 \pm 0.9$  ka (ARAN-31) of a polished surface on the ridge dividing the Sendrosa and Saboredo cirques (Fig. 3). The sample from an outcrop on the lower edge of the cirque (ARAN-34;  $14.0 \pm 0.8$  ka) indicates that ice covered the entire Saboredo cirque floor until  $\sim 14$  ka.

The sampled boulders of the external moraine system (ARAN-38 and ARAN-37) showed statistical coherence based on the chi-squared test following Ward and Wilson (1978) and suggested a glacier advance or stillstand at  $13.5 \pm 0.9$  ka. The age of  $13.0 \pm 0.8$  ka from a moraine boulder (ARAN-27) and an adjacent polished surface (ARAN-28) of the internal moraine system indicate another short advance or stillstand, which is also consistent with the geochronological sequence. These results from staggered moraine systems at  $\sim 2080$  and  $\sim 2190$  m provided chronologically consistent mean ages within the B-A. But is based on their distinct relative position that they have been ascribed to two different stages. Nevertheless, during this period, glaciers flowed from the Saboredo and Sendrosa cirques and generated moraines on the lower overdeepened basins.

According to the ages of samples from the highest surfaces of the Saboredo Cirque floor at 2320-2360 m (ARAN-35 and ARAN-36;  $12.7 \pm 0.8$  ka), the entire cirque floor was deglaciated at the onset of the YD. These ages are consistent with those of the samples from the highest dated moraine cirques in the Sendrosa Cirque at 2460-2470 m (ARAN-32 and ARAN-33), which was also abandoned by the ice by  $12.6 \pm 1.3$  ka. No statistical inconsistency was detected in the exposure ages from this moraine.

## Figure 5

### 5.2 The chronology of deglaciation

During the ILGM, the Garonne glacier flowed  $\sim 80$  km reaching the Loures-Barousse-Barbazan basin, where it formed a terminal moraine system (Fernandes et al., 2017; Stange et al., 2014). Until the onset of deglaciation, 300-400 m thick ice sat in the Ruda Valley, as shown by the highest lateral moraines between the Ruda and Aiguamòg valleys. This ice thickness must have favoured the occurrence of warm-based glaciers, thus implying basal sliding evidenced by striations on the bedrock surface. Hence, no nuclide inheritance from previous exposure periods is expected at the sampled surfaces.

Samples from polished surfaces from the bottom of the Ruda Valley at 1860-1900 m indicate that by  $14.4 \pm 1.2$  ka, i.e. during the B-A, the glacier was retreating towards the glacial cirque, and was already disconnected from the main Ruda glacier. Consequently, one third of the Ruda Valley was already deglaciated at that time, indicating that by the early B-A the Garonne glacier represented only  $\sim 5\%$  of the length that it had during the ILGM (Fernandes et al., 2017).

As glaciers retreated and thinned during the B-A, ice masses became individualized within their respective cirques, and the valley glaciers flowing from Sendrosa and Saboredo cirques merged only in the lowest overdeepened basins (Fig. 3). As conditions were less favourable, glaciers receded to higher elevations and the Sendrosa and Saboredo paleoglaciers became isolated in their respective basins by  $14.8 \pm 0.9$  ka (Fig. 3). The ice covered the entire Saboredo cirque floor until  $\sim 14$  ka, as revealed by the sample collected from an outcrop on the edge of the cirque ( $14.0 \pm 0.8$  ka), indicating that glacial recession in the Ruda Valley was underway at  $\sim 15$ -14 ka. Similarly, a rapid deglaciation was also recorded at  $\sim 15$ -14 ka in the nearby Bacivèr Cirque, which favoured its complete deglaciation (Fig. 1; Oliva et al., 2021a).

The strong clustering of ages at between  $\sim 13.5$  and  $12.7$  ka confirms the dynamic environmental response of the Ruda palaeoglaciers to the abrupt climate transition between the B-A and the early YD. However, our CRE

dataset cannot assess the magnitude and patterns of glacial recession during the early B-A in the Ruda Valley, or whether the entire cirque was deglaciated and glaciers reformed and subsequently advanced. Our results suggest, however, that the millennial-scale trend towards glacier recession during the B-A was interrupted by two glacial advances or stillstands with moraine development at  $13.5 \pm 0.9$  and  $13.0 \pm 0.8$  ka (Fig. 5). Samples from moraines located at average elevations of  $\sim 2080$  and  $\sim 2190$  m, respectively, indicated the existence of valley glaciers 3.5 and 2.5 km long during the middle of the B-A and during the transition from the B-A to the YD. Indeed, the reconstructed glacial surface for both bases was of 4.4 and 3.8 km<sup>2</sup>, respectively, with a maximum ice thickness of 300 and 200 m for each of these phases (Fig. 6a, b).

### Figure 6

Evidence of glacial recession during the middle of the B-A was also recorded on polished surfaces from the neighbouring Bacivèr Cirque at  $13.5 \pm 0.8$  ka (Oliva et al., 2021a). Local topographical factors, such as the shape of the cirque floor (convex versus concave areas) determined the pattern of glacial flow of thin ice masses during the glacial advances/stillstands that occurred during the B-A. Glaciers flowed downwards from the cirque through concave areas across the western slope, favouring the development of the moraine arcs and crests in lower overdeepened basins (Fig. 3). It is likely that the topography played a major role in determining the process and timing of the deglaciation in both cirques: the accumulation area above 2400 m is larger in the Saboredo Cirque (552 ha) with respect to the Bacivèr Cirque (244 ha) and its prevailing aspect is also more favourable for ice persistence (N vs W).

The Saboredo Cirque floor became ice free at the beginning of the YD, as revealed by the samples from polished surfaces of the cirque floor at  $\sim 2350$  m (average age =  $12.7 \pm 0.8$  ka). The evidence of glacial striations on the rock surface make unlikely the possibility of nuclide inheritance at the sampling sites. Glaciers might thus also have persisted during the early YD in the highest hollows of the Saboredo Cirque, as was the case in the northern cirque of Sendrosa Peak. Here, the cirque moraine at 2470 m, formed by a 0.4 km-long glacier (with a surface of 0.7 km<sup>2</sup> and a thickness of 90 m; Fig. 6c), was abandoned at  $12.6 \pm 1.3$  ka. Inside the highest glacial cirques of the Ruda Valley, there are numerous relict rock glaciers and protalus lobes indicative of the past occurrence of permafrost during their formation (Fernandes et al., 2018). As in most of the Pyrenees, these are probably glacier-derived permafrost features associated with the paraglacial phase, when glacial shrinking favoured wall debuttrressing, a high debris supply and the subsequent burial of the residual ice masses (Andrés et al., 2018; Oliva et al., 2016).

In short, by the early B-A the Garonne Paleoglacier had receded  $\sim 95\%$  of its  $\sim 80$  km length of the ILGM (Fernandes et al., 2017; Stange et al., 2014). The Ruda Paleoglacier receded 4 km in only  $\sim 1.7$  ka ( $\sim 14.4$  to  $\sim 12.7$  ka) from an elevation of  $\sim 1800$  to  $>2400$ - $2500$  m. This trend towards glacial shrinking before the final ice disappearance was interrupted by three phases of glacier advance or stillstands at  $13.5 \pm 0.9$ ,  $13.0 \pm 0.8$  and  $12.6 \pm 1.3$  ka, which generated moraines at  $\sim 2080$ , 2190 and 2470 m, respectively.

### 5.3 Late Quaternary glacial dynamics in the Central Pyrenees and Iberian Peninsula in the context of European glacial evolution

The scientific knowledge on the chronology of the last deglaciation in the Pyrenees has experienced substantial advances over the last years in parallel to progress on glacial chronological constraints in other European and Iberian mountain ranges (García-Ruiz et al., 2010; Palacios and García-Ruiz, 2015). There is increasing evidence that the timing of glacial oscillations during T-1 across Europe followed a similar sequence (Palacios et al., 2021): a significant glacier advance took place during the OD followed by a rapid retreat during the B-A and a subsequent more limited glacier advance during the YD (Oliva et al., 2019).

Dated evidence of glacial responses in the Pyrenees to Northern Hemisphere cold events during T-1 have mostly focused on the southern slope (Oliva et al., 2019), although there is an increasing body of new data from the northern slope of the range that confirms glacial advances and stillstands there during the OD and YD, and glacial shrinking during the B-A (Reixach et al., 2021). The elevation of the peaks as well as the topography of the cirques played a major role in the duration and intensity of the glaciation during T-1 in the



high massifs of the Pyrenees. Cirques surrounded by peaks above 2800-2900 m likely included glaciers during most of T-1 until the onset of the Holocene (see e.g. Jomelli et al., 2020; Pallàs et al., 2006; Reixach et al., 2021), while those situated at the foot of lower peaks (i.e. 2500-2600 m) were completely deglaciated during the B-A (Oliva et al., 2021a; Palacios et al., 2017b). Following ice retreat, the slopes were affected by intense readjustments (large, catastrophic rock slope failures, rock slope deformations, rockfalls, and rockslides), particularly in areas where the local lithostructure was prone to instability (Fernandes et al., 2020).

In the Central Pyrenees, a period of glacial growth occurred during the OD after the massive deglaciation that took place following the LGM. Moraines dated at 17-15 ka have been reported in the highest valleys of the main massifs, such as in the Ariège (Delmas et al., 2011; Jomelli et al., 2020; Reixach et al., 2021), Ésera (Crest et al., 2017), Gállego (Palacios et al., 2015a), Malniu (Andrés et al., 2018; Palacios et al., 2015b; Pallàs et al., 2010), Têt (Delmas et al., 2008; Tomkins et al., 2018), Arànsér (Andrés et al., 2018; Palacios et al., 2015b), and Bassiès (Crest et al., 2017). During that time, glaciers flowed downslope from the headwaters of the highest valleys, with valley glaciers often exceeding 10-15 km (Oliva et al., 2019). Consequently, it is likely that, by that time, most of the Ruda Valley was still glaciated, with the ice tongue probably merging with the main Garonne Paleoglacier that was also fed by neighbouring glacier tributaries, such as those descending from the Beret area (Oliva et al., 2021a). Some of the lateral moraine remnants distributed on the left side of the Garonne Valley, ca. 20 km from the highest parts of the Saboredo Cirque, formerly attributed to the OD (Fernandes et al., 2021; Fernandes et al., 2017) may be also correlated with some of the highest moraines and erratic boulders distributed in the Ruda Valley (Fig. 3). Other Iberian mountain ranges include geomorphic evidence of glacial advances during the OD (Palacios et al., 2017a), such as in the Monasterio Valley in the Cantabrian Mountains (Serrano et al., 2016), Peña Negra Valley in the Iberian Range (García-Ruiz et al., 2020), Peñalara massif in the Central Range (Palacios et al., 2012a), San Juan Valley in the Sierra Nevada (Palacios et al., 2016), En Garcia Valley in the Eastern Pyrenees (Reixach et al., 2021), and Gállego and Ésera valleys in the Central Pyrenees (Crest et al., 2017; Palacios et al., 2017b).

No moraine remnants have been found in the lowest and central sections of the Ruda Valley, and the prevailing lithology of slates and marbles in this area is not favourable for the preservation of glacially abraded surfaces. Indeed, the neighbouring Bacivèr Cirque was completely deglaciated also by ~15-14 ka (Oliva et al., 2021a). Thus, our data cannot confirm whether small glaciers persisted in the highest parts of the Ruda Valley during the B-A, although CRE ages support climatic conditions during the early B-A unfavourable for the maintenance of the OD ice masses (Fig. 7). In other valleys of the Pyrenees, glaciers also receded rapidly at ~14.6-14 ka as shown by CRE ages from polished surfaces and scattered erratic boulders in the Ariège ( $14.4 \pm 1.1$ ,  $14.1 \pm 0.7$ ,  $14.0 \pm 0.8$  ka; Delmas et al., 2011), Gállego ( $14.6 \pm 2.3$  ka; Palacios et al., 2015a) and Ésera valleys ( $14.3 \pm 0.5$  ka; Crest et al., 2017). Locally, some massifs of the Pyrenees were fully deglaciated during the B-A, with small glaciers persisting only in the highest cirques (Delmas, 2015). A similar pattern was detected in other Iberian ranges, such as in the Sierra Nevada (Palacios et al., 2016), Cantabrian Mountains (Rodríguez-Rodríguez et al., 2017) and the Iberian Range (García-Ruiz et al., 2020), where ice masses shrank significantly by ~15-14 ka. Glacial retreat in the Iberian Mountains was triggered by the sudden temperature increase of 3-5 °C at the onset of the B-A in western Europe, as inferred from climate models (Renssen and Isarin, 2001). In parallel, significant shifts occurred in the Atlantic Meridional Overturning Circulation (Obase and Abe-Ouchi, 2019). In north Iberia, a maximum increase of the mean annual temperature of 7.5°C at the onset of GI-1 has been estimated in northern Iberia, based on the Ostolo cave speleothem isotopic record (Bernal-Wormull et al., 2021). Marine records from the western Iberian margin also pointed to a 5-6°C of warming at the onset of the B-A (Martrat et al., 2007). These warmer conditions also favoured the recession of glaciers across most European mountain ranges, such as in the Alps at 15.9-14.3 cal ka BP (Ivy-Ochs, 2015), Tatra Mountains at 14.8-14.2 ka (Zasadni et al., 2020), Taurus Mountains at  $14.6 \pm 2.8$  ka (Ciner and Sarikaya, 2015), as well as the Scandinavian Ice Sheet, which recorded a shrinking trend starting at 14.6 cal ka BP (Mangerud et al., 2013).

1  
2 463 However, the B-A was a climatically unstable period characterised by an alternation of warm and cold  
3 464 temperatures (Rasmussen et al., 2014). The gradual decrease of temperatures from the early to the late B-A  
4 465 was depicted by periods of enhanced cold conditions in the high latitudes of the Northern Hemisphere, which  
5 466 also favoured periods of glacial growth in the Pyrenees (Table 5). Indeed, CRE results indicate the formation  
6 467 of moraines in the Ruda Valley at  $13.5 \pm 0.9$  and  $13.0 \pm 0.8$  ka (Fig. 7). These CRE ages are consistent with  
7 468 the chronostratigraphic sequence but must be taken as approximate due to their uncertainty ranges and the low  
8 469 number of dated samples at each moraine. At that time, small ice tongues  $\sim 3$  km long flowed downslope from  
9 470 the main Saboredo Cirque as well as from other neighbouring glacial hollows to overdeepened basins, forming  
10 471 moraines at 2080 and 2190 m, respectively. Whereas evidence of phases of glacier advance or stillstands during  
11 472 the transition from the late B-A to the early YD were already found in other cirques in the Central Pyrenees,  
12 473 such as in the Bacivèr Cirque (Oliva et al., 2021a), and in the Cuerpo de Hombre Valley (Sierra de Gredos,  
13 474 Central Range), with moraine stabilization at 13.1 ka (Carrasco et al., 2015), this is the first study that reports  
14 475 geochronological evidence of a phase with glacial development in the second half of the B-A after the long-  
15 476 term cooling initiated at  $\sim 14.5$  ka (Fig. 7; Table 5). A few studies in Europe have also found evidence of glacier  
16 477 advances or stillstands during the middle B-A, such as in the Tatra Mountains where moraines stabilized at  
17 478  $13.4 \pm 0.5$  ka (Engel et al., 2017). By contrast, geomorphic evidence of a period of glacial expansion during  
18 479 the transition between the B-A to the early YD was detected in several regions, such as in the  
19 480 Hardangerfjorden-Herdla area, western Norway, where a till deposit suggested a glacier advance of the  
20 481 Scandinavian Ice sheet at 13.5-13.0 ka (Mangerud et al., 2016), and in the West Highland and Mull ice fields  
21 482 of the Scottish Highlands, where 18  $^{14}\text{C}$  dates from organic fragments found in basal tills revealed a glacial  
22 483 readvance from the end of the B-A or earliest YD (Bromley et al., 2018), although these results are considered  
23 484 controversial (Small and Fabel, 2016).

#### 29 485 Table 5

30  
31 486 Based on the position of these two moraine ridges, the ELA in the Ruda Valley glacier must have been located  
32 487 at 2461 and 2505 m during the glacial advances of the second part of the B-A and the late B-A or early YD,  
33 488 respectively. Considering that the regional ELA in the Pyrenees currently lies at approximately 3100 m  
34 489 (Jomelli et al., 2020; René, 2011), the ELA must have been located 639-595 m lower than present-day  
35 490 conditions. Assuming an average lapse rate of  $0.65\text{ }^{\circ}\text{C } 100\text{ m}^{-1}$  and no change in precipitation, we infer that  
36 491 summer temperatures must have been 4.2 and 3.9  $^{\circ}\text{C}$  lower than present-day during these two phases (Table  
37 492 4).

#### 40 493 Figure 7

41  
42 494 The temperature decrease during the early YD favoured the persistence of glaciers inside the highest hollows  
43 495 in the Ruda Valley, depositing moraines at  $\sim 2400$ -2500 m, which were later abandoned by the ice at  $12.6 \pm$   
44 496  $1.3$  ka (Fig 7). The existence of glaciers at high altitudes of the Saboredo Cirque is also confirmed by polished  
45 497 surfaces at elevations of 2200-2300 m that were dated at  $12.7 \pm 0.8$  ka. In the neighbouring Bacivèr Cirque,  
46 498 moraines at 2400 m were abandoned by the ice at 12.8 ka (Oliva et al., 2021a). The elevation difference  
47 499 between the ELA during the YD (2571 m) in the Ruda Valley and present-day values indicates that summer  
48 500 temperatures were  $\sim 3.4\text{ }^{\circ}\text{C}$  lower during that cold phase (Table 4), which is very similar to the  $\sim 3.0\text{ }^{\circ}\text{C}$  inferred  
49 501 for the Bacivèr Cirque (Oliva et al., 2021a). Indeed, topoclimatic factors must have played a key role  
50 502 maintaining small glaciers during the YD at the foot of steep cirque walls, where shading and snow  
51 503 accumulation allowed the longer persistence of reduced ice masses (Boston and Lukas, 2019). In the Pyrenees,  
52 504 YD glacial advances have been detected in cirques of the Central (Delmas, 2009; Delmas et al., 2008; García-  
53 505 Ruiz et al., 2016a) and Eastern sectors (Crest et al., 2017; Jomelli et al., 2020; Pallàs et al., 2010; Reixach et  
54 506 al., 2021), where 2-3.5 km-long glaciers formed at 12.9-11.2 ka. Based on chronostratigraphic data from a  
55 507 variety of sources, evidence of YD glacial advances was also recorded in the highest cirques of other Iberian  
56 508 massifs, although no direct ages are yet available (García-Ruiz et al., 2016a). This is true for the NW ranges  
57 509 (Cowton et al., 2009) and the Cantabrian Mountains (Pellitero et al., 2019; Serrano et al., 2015), where  
58 510 moraines above 1800 and 2000 m, respectively, were associated with this cold phase. In the Sierra Nevada,

1  
2 511 YD moraines dated at 13-12 ka are located in the highest cirques facing east from 2350 (north face) to 2800 m  
3 512 (south face) (Oliva et al., 2014; Palacios et al., 2016). These glacial advances were driven by a temperature  
4 513 cooling quantified at 2-4 °C in western Europe that was triggered by a slowdown of the overturning circulation  
5 514 in the North Atlantic (Renssen et al., 2018). In the Pyrenees, the YD was the coldest period of the last  
6 515 deglaciation recorded by a decrease of  $^{18}\text{O}$  in the Ostolo and Seso speleothems that might reflect the  
7 516 temperature decrease of  $\sim 5^\circ\text{C}$  (Bartolomé et al., 2015; Bernal-Wormull et al., 2021). Colder conditions also  
8 517 favoured glacial growth in other mountainous regions of Europe, with widespread advances across the Alps  
9 518 between 13.5 and 12 ka (Ivy-Ochs, 2015; Ivy-Ochs et al., 2009), in the Tatra Mountains where moraines  
10 519 stabilized at  $11.9 \pm 0.5$  ka (Engel et al., 2017) and in Turkey where they formed at  $12.6 \pm 2.3$  ka (Ciner and  
11 520 Sarikaya, 2015).

12  
13 521 The final deglaciation of the cirques favoured paraglacial dynamics and the formation of the numerous rock  
14 522 glaciers existing today in the Saboredo and Sendrosa cirques, as well as in the lateral cirques distributed along  
15 523 the Ruda Valley (Fig. 3). These features are currently inactive under the present-day climate regime (Fernandes  
16 524 et al., 2018), with abundant lichen coverage and forest colonization, particularly on the lowest ridges. In the  
17 525 Pyrenees, as well as in other Iberian ranges, cirque wall readjustment following deglaciation triggered different  
18 526 landforms, such as debris-covered glaciers or rock glaciers, that remained active until the early Holocene  
19 527 (Oliva et al., 2019). This is the case of the Bacivèr cirque, where the debris-covered glacier that formed inside  
20 528 the highest YD moraines stabilized at  $7.2 \pm 0.5$  ka (Oliva et al., 2021a), or of several other massifs in the  
21 529 Central and Eastern Pyrenees where rock glaciers were active until well into the Holocene (Andrés et al., 2018;  
22 530 García-Ruiz et al., 2016a). A similar pattern was also detected in other Iberian massifs, such as in the Iberian  
23 531 Range (Aumaître et al., 2017) or Sierra Nevada (Palacios et al., 2016), where these paraglacial landforms  
24 532 formed immediately after the deglaciation of the cirques, and remained active until the complete melting of  
25 533 their relict ice during the Holocene Thermal Maximum.  
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27

## 28 534 6. Conclusions

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30  
31 535 The Central Pyrenees were extensively glaciated during Quaternary glacial phases and were only slightly  
32 536 transformed by periglacial, nival and slope processes during subsequent interglacial periods such as the  
33 537 Holocene. As a result, the landscape in the Central Pyrenees includes a wide range of geomorphological  
34 538 features that constitutes a prominent natural heritage that contributed to the designation of most protected areas  
35 539 in this mountain range. This is the case of the Ruda Valley, located on the periphery of the National Park of  
36 540 Aigüestortes and Sant Maurici Lake. The valley is one of the most scenic glacial valleys of the Pyrenees, and  
37 541 includes a wide variety of glacial and periglacial landforms, particularly above 1800 m. However, in contrast  
38 542 to the southern slope of the Central Pyrenees where the timing of the glacial evolution has been widely studied,  
39 543 on the northern side there are still temporal and spatial gaps in the chronology of glacial activity. In this work,  
40 544 we determined the magnitude and timing of T-1 glacial oscillations in the Upper Garonne Basin by producing  
41 545 12 new CRE ages of glacial landforms (moraine boulders and glacially polished surfaces) that provide evidence  
42 546 of the spatio-temporal pattern of glacial oscillations from the B-A to the YD.  
43

44  
45 547 The Upper Garonne Basin hosted the largest glacial system of the Pyrenees during the last glacial cycle thanks  
46 548 to its Atlantic-influenced climate and the high altitude of the peaks in the central part of this range. The long-  
47 549 term post-LGM warming favoured glacial recession, particularly during the early B-A, and by  $14.4 \pm 1.2$  ka  
48 550 the Ruda Paleoglacier was already disconnected from the main Garonne Paleoglacier. At that time, the glacier  
49 551 front was located at 1860-1900 m, and some areas of the Saboredo Cirque were already ice-free as a result of  
50 552 ice thinning. The B-A was also a period of great climatic variability, and the small glaciers in the Ruda Valley  
51 553 showed rapid response times with formation of moraines during the coldest B-A phases in north Iberia. Phases  
52 554 of glacier advance or stillstands promoted the development of moraines at  $13.5 \pm 0.9$  ka, as well as during the  
53 555 transition between the late B-A and the onset of the YD at  $13.0 \pm 0.8$  ka. The YD probably favoured the  
54 556 persistence of small glaciers during an early stage of this cold phase; there is evidence of glacial retreat at the  
55 557 uppermost part of the cirque floor at  $12.7 \pm 0.7$  ka ( $\sim 2300$ -2350 m) and moraine abandonment of the highest

1  
2 558 cirque moraines (2470 m) at  $12.6 \pm 1.3$  ka. Subsequently, glacial disappearance must have promoted  
3 559 paraglacial dynamics and the formation of the (currently relict) rock glaciers existing within these moraines.  
4

5 560 Whereas the impact of the B-A deglaciation on the current landscape of the Pyrenean cirques was already  
6 561 known, the sequence of glacial advances and retreats during the B-A and YD presented in this study provides  
7 562 one of the most accurate deglacial histories in southern Europe. This opens new perspectives on the  
8 563 paleoclimatic evolution in the Iberian Peninsula, in particular, and the Mediterranean region in general. The  
9 564 period spanning from the B-A to the YD has been shown to be a major driver of transformation of the mountain  
10 565 landscapes of the Pyrenees and a key phase to better frame Holocene environmental dynamics in the highest  
11 566 lands of this mountain range. Future studies with new chronological data from other Pyrenean valleys and  
12 567 Iberian mountains will help to better understand T-1 glacier advances and retreats as well as the climate  
13 568 mechanisms behind millennial-scale glacial oscillations in southern Europe.

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2 884 **List of figures**

3  
4 885 Figure 1. Geographical setting of the Ruda Valley. (A) Location in Europe. (B) Location in the Pyrenees with  
5 886 the distribution of the studies mentioned in the discussion (green stars). (C) Limits and topography of the valley  
6 887 with the distribution and view direction of the photos in Figure 2.

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8 888 Figure 2. Examples of the main geomorphological landforms of glacial origin in the Ruda Valley, together  
9 889 with the location and typology of the dated samples: (A) terminal moraine damming a paleolake at 2080 m;  
10 890 (B) N-S perspective of the Saboredo Cirque floor showing several lakes distributed in staggered overdeepened  
11 891 basins; (C) polished bedrock surface including striations that reveal the former glacial flow; and (D) *roches*  
12 892 *moutonnées* located in the forested valley bottom of the valley at 1860 m.

14 893 Figure 3. (A) Geomorphological map of the study area, with the location of the CRE samples; and (B)  
15 894 enlargement of the highest parts of the cirque including the location of the  $^{10}\text{Be}$  ages.

17 895 Figure 4. Examples of sampled moraines and polished bedrock surfaces, together with CRE ages (ka).

19 896 Figure 5. Reconstruction of the glacial extent during the different time stages: (A) Extensive glaciers occupying  
20 897 the entire Upper Garonne Basin during the ILGM based on the reconstruction of [Fernandes et al. \(2017\)](#); (B)  
21 898 Glacial extent during the mid B-A at ~13.5 ka in the Ruda Valley; (C) Glacier flowing from the Saboredo  
22 899 Cirque during the transition between the late B-A and the early YD at ~13 ka; and (D) Small glaciers confined  
23 900 within the highest hollows next to the Sendrosa (D) and Saboredo (E) peaks during the early YD at ~12.7/12.6  
25 901 ka.

26 902 Figure 6. Schematic evolution of the glacial phases recorded in the Ruda Valley during the last glacial cycle:  
27 903 (A) extensive and thick glaciers inundated the valley and only few nunataks protruded the ice during the ILGM;  
29 904 (B) warm temperatures in the early B-A favoured glacial recession, and subsequently the lowest areas of the  
30 905 valley as well as some rock outcrops of the cirque became exposed; (C) glacial advances/stillstands formed  
31 906 moraines during the B-A and at the transition to the YD; (D) small glaciers persisted during the early YD in  
32 907 the highest north-facing cirques; and (E) subsequent temperature increases favoured complete deglaciation and  
34 908 triggered paraglacial dynamics, which promoted the formation of rock glaciers in the recently deglaciated ice-  
35 909 free cirques.

36 910 Figure 7. Normalized probability distribution functions (PDF) of surface exposure ages vs. temperature  
38 911 evolution from the OD to the early Holocene based on the  $\delta^{18}\text{O}$  record from the NGRIP ice core from  
39 912 Greenland and the  $\delta^{18}\text{O}$  record from the speleothem of the Ostolo cave in north Iberia (time periods are defined  
40 913 after [Rasmussen et al. \(2014\)](#)). The plots of the units result from the sum of the individual PDF of the samples  
41 914 belonging to them; cold (warm) phases are represented by the blue (red) bands.

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1 **List of tables**

2 **Table 1. Geographic location of samples, topographic shielding factor and sample thickness.**

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Sample name	Landform	Latitude (DD)	Longitude (DD)	Elevation (m a.s.l.) <sup>a</sup>	Topographic shielding factor	Thickness (cm)	Lithology	Size of the boulders largest axis (m)
<i>Valley polished surfaces</i>								
ARAN-40	Polished surface	42.6416	0.9713	1862	0.9630	3.7	Granodiorite	
ARAN-39	Polished surface	42.6393	0.9701	1900	0.9640	5.0	Granodiorite	
<i>Valley external moraines</i>								
ARAN-38	Moraine boulder	42.6296	0.9624	2079	0.9759	4.2	Granite	3.5
ARAN-37	Moraine boulder	42.6299	0.9620	2075	0.9752	3.0	Granite	3
<i>Valley internal moraines</i>								
ARAN-28	Polished surface	42.6263	0.9593	2188	1.0000	3.5	Granite	
ARAN-27	Moraine boulder	42.6257	0.9599	2184	0.9852	5.0	Granite	4
<i>Cirque polished surfaces</i>								
ARAN-36	Polished surface	42.6111	0.9681	2360	0.9806	3.5	Granite	
ARAN-35	Polished surface	42.6152	0.9630	2316	0.9875	3.2	Granite	
ARAN-34	Polished surface	42.6224	0.9596	2266	0.9901	5.0	Granite	
ARAN-31	Polished surface	42.6274	0.9536	2378	1.0000	3.0	Granite	
<i>Cirque moraines</i>								
ARAN-33	Moraine boulder	42.6275	0.9487	2470	0.9841	3.0	Granite	5
ARAN-32	Moraine boulder	42.6277	0.9480	2467	0.9789	4.0	Granite	6

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32 <sup>a</sup> Elevations are derived from the 5 m Digital Elevation Model of the Spanish "Instituto Geográfico Nacional" and are subjected to a vertical accuracy of  $\pm 5$  m.

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**Table 2. AMS analytical data and calculated exposure ages.  $^{10}\text{Be}/^9\text{Be}$  ratios were inferred from measurements at the ASTER AMS facility. Chemistry blank correction was done by subtracting the number of  $^{10}\text{Be}$  atoms in the blank from that in the sample. Individual ages are shown with their full uncertainties (including analytical AMS uncertainty and production rate uncertainty) and analytical uncertainty only within brackets. Arithmetic mean ages are given with their full uncertainties (including standard deviation and production rate uncertainty) and standard deviations only in brackets.**

<b><math>^{10}\text{Be}</math> samples analytical AMS data</b>								
Sample name	Quartz weight (g)	Mass of carrier ( $^9\text{Be}$ mg)	ASTER AMS cathode number	$^{10}\text{Be}/^9\text{Be}$ ( $10^{-14}$ )	Blank correction (%)	$[^{10}\text{Be}]$ ( $10^4$ atoms $\text{g}^{-1}$ )	Age (ka)	Mean age (ka)
<i>Valley polished surfaces</i>								
ARAN-40	20.44	0.46	CHBL	$16.34 \pm 0.56$	2.11	$23.85 \pm 0.83$	$15.0 \pm 0.9$ (0.5)	$14.4 \pm 1.2$ (0.8)
ARAN-39	21.32	0.45	CHBK	$15.92 \pm 0.49$	2.18	$22.15 \pm 0.70$	$13.8 \pm 0.8$ (0.4)	
<i>Valley external moraines</i>								
ARAN-38	21.11	0.46	CHBJ	$17.56 \pm 0.54$	1.95	$24.99 \pm 0.79$	$13.3 \pm 0.8$ (0.4)	$13.5 \pm 0.9$ (0.2)
ARAN-37	21.38	0.46	CHBI	$18.08 \pm 0.87$	1.09	$25.74 \pm 1.25$	$13.7 \pm 1.0$ (0.6)	
<i>Valley internal moraines</i>								
ARAN-28	20.97	0.46	CHBB	$18.79 \pm 0.62$	1.05	$27.13 \pm 0.91$	$13.0 \pm 0.8$ (0.4)	$13.0 \pm 0.8$ (0.1)
ARAN-27	20.89	0.45	CHBA	$18.64 \pm 0.58$	1.85	$26.61 \pm 0.84$	$13.1 \pm 0.8$ (0.4)	
<i>Cirque polished surfaces</i>								
ARAN-36	21.51	0.46	CHBH	$21.24 \pm 0.65$	1.62	$29.53 \pm 0.92$	$12.7 \pm 0.8$ (0.4)	$12.7 \pm 0.8$ (0.0)
ARAN-35	21.43	0.47	CHBG	$20.18 \pm 0.62$	1.65	$29.05 \pm 0.91$	$12.8 \pm 0.8$ (0.4)	
ARAN-34	20.85	0.46	CHBF	$21.13 \pm 0.66$	1.63	$30.35 \pm 0.96$	$14.0 \pm 0.8$ (0.4)	
ARAN-31	21.40	0.44	CHBC	$26.54 \pm 0.72$	1.35	$35.74 \pm 0.98$	$14.8 \pm 0.9$ (0.4)	
<i>Cirque moraines</i>								
ARAN-33	20.26	0.45	CHBE	$20.30 \pm 0.68$	1.70	$29.92 \pm 1.02$	$11.8 \pm 0.7$ (0.4)	$12.6 \pm 1.3$ (1.0)
ARAN-32	21.63	0.45	CHBD	$24.19 \pm 1.26$	0.85	$33.15 \pm 1.75$	$13.3 \pm 1.0$ (0.7)	
<b>Chemistry blank details</b>								
Blank name	Processed with	mass of carrier ( $^9\text{Be}$ mg)	ASTER AMS cathode number	$^{10}\text{Be}/^9\text{Be}$ ( $10^{-14}$ )	$[^{10}\text{Be}]$ ( $10^4$ atoms)			
BK-1	ARAN-27, 31, 33, 34, 35, 36, 38, 39, 40	0.46	CHAT	$0.34 \pm 0.04$	$10.30 \pm 1.24$	-	-	
BK-2	ARAN-32	0.46	CHBM	$0.20 \pm 0.03$	$6.13 \pm 0.80$	-	-	
BK-3	ARAN-28, 37	0.46	IGHI	$0.20 \pm 0.03$	$5.98 \pm 0.93$	-	-	

**Table 3. Exposure ages according to erosion and snow cover corrections.**

Sample name	Exposure ages (arithmetic mean, in ka)			
	No correction	Erosion correction	Snow correction	Erosion + snow correction
<i>Valley - polished surfaces</i>	$14.4 \pm 1.2$	$14.6 \pm 1.2$	$15.5 \pm 1.3$	$15.7 \pm 1.3$
ARAN-40	$15.0 \pm 0.9$ (0.5)	$15.2 \pm 0.9$ (0.5)	$16.1 \pm 1.0$ (0.5)	$16.4 \pm 1.0$ (0.6)
ARAN-39	$13.8 \pm 0.8$ (0.4)	$14.0 \pm 0.8$ (0.4)	$14.8 \pm 0.9$ (0.4)	$15.0 \pm 0.9$ (0.5)
<i>Valley - external moraines</i>	$13.5 \pm 0.9$	$13.7 \pm 0.9$	$14.5 \pm 1.0$	$14.7 \pm 1.0$
ARAN-38	$13.3 \pm 0.8$ (0.4)	$13.5 \pm 0.8$ (0.4)	$14.4 \pm 0.9$ (0.4)	$14.6 \pm 0.9$ (0.4)
ARAN-37	$13.7 \pm 1.0$ (0.6)	$13.8 \pm 1.0$ (0.7)	$14.7 \pm 1.0$ (0.7)	$14.9 \pm 1.0$ (0.7)
<i>Valley - internal moraines</i>	$13.0 \pm 0.8$	$13.2 \pm 0.8$	$14.1 \pm 0.9$	$14.2 \pm 0.9$
ARAN-28	$13.0 \pm 0.8$ (0.4)	$13.1 \pm 0.8$ (0.4)	$14.0 \pm 0.9$ (0.5)	$14.2 \pm 0.9$ (0.5)
ARAN-27	$13.1 \pm 0.8$ (0.4)	$13.3 \pm 0.8$ (0.4)	$14.1 \pm 0.8$ (0.4)	$14.3 \pm 0.9$ (0.4)
<i>Cirque - polished surfaces</i>	$12.7 \pm 0.7$	$12.9 \pm 0.8$	$13.7 \pm 0.8$	$13.9 \pm 0.8$
ARAN-36	$12.7 \pm 0.8$ (0.4)	$12.9 \pm 0.8$ (0.4)	$13.7 \pm 0.8$ (0.4)	$13.9 \pm 0.8$ (0.4)
ARAN-35	$12.8 \pm 0.8$ (0.4)	$12.9 \pm 0.8$ (0.4)	$13.8 \pm 0.8$ (0.4)	$14.0 \pm 0.8$ (0.4)
ARAN-34	$14.0 \pm 0.8$ (0.4)	$14.2 \pm 0.9$ (0.4)	$15.1 \pm 0.9$ (0.5)	$15.3 \pm 0.9$ (0.5)
ARAN-31	$14.8 \pm 0.9$ (0.4)	$15.1 \pm 0.9$ (0.4)	$16.0 \pm 0.9$ (0.4)	$16.2 \pm 0.9$ (0.4)
<i>Cirque - moraines</i>	$12.6 \pm 1.3$	$12.7 \pm 1.4$	$13.5 \pm 1.4$	$13.7 \pm 1.5$
ARAN-33	$11.8 \pm 0.7$ (0.4)	$12.0 \pm 0.8$ (0.4)	$12.7 \pm 0.8$ (0.4)	$12.9 \pm 0.8$ (0.4)
ARAN-32	$13.3 \pm 1.0$ (0.7)	$13.5 \pm 1.0$ (0.7)	$14.3 \pm 1.0$ (0.7)	$14.5 \pm 1.0$ (0.7)

**Table 4. Reconstructed ELAs (m a.s.l.) for the middle, late B-A and YD moraines using the AAR and AABR methods.**

Phase	Glacier	AAR (0.6±0.2)	AABR (1.9±0.81)	AABR (1.75±0.71)	Average ELA
Valley external moraines (middle B-A; GI-1c)	Ruda	2481 ± 20	2449 -30/+20	2454 -30/+15	2461
Valley internal moraines (transition B-A/YD; GI-1b)	Ruda	2515 -15/+10	2498 -25/+15	2503 -25/+20	2505
Cirque moraines (early YD)	Sendrosa	2568 -10/+5	2571 -15/+5	2576 ± 10	2571

**Table 5. Glacial response in the Ruda Valley during the B-A Interstidal according to the reconstructed climate in Greenland and in Iberian Peninsula.**

Phase (ages in ka)	Climate trend in Greenland based on ice core records (Rasmussen et al., 2014)	Glacier response in the Ruda Valley	Climate in the Iberian Peninsula based on other independent proxies
GI-1e (14.7-14.1)	Abrupt warming (3-5°C)	Rapid recession at ~15-14 ka	Gradual warming in the northern Iberia at 15.4-13.4 cal ka BP (Moreno et al., 2010). Warming conditions in the Pyrenees at 14.7-14.0 cal ka BP (González-Sampérez et al., 2006).
GI-1d (14.1-14.0)	Short cold reversal (1-2°C)		
GI-1c (14.0-13.3)	Significant warming at the beginning (2-4°C) followed by a gradual cooling at (~2°C). Intense cold conditions at 13.7-13.6 ka	Glacier advance or stillstand with moraine formation at 13.5 ka	Colder conditions starting at 13.5 cal ka BP in northern Iberia (Moreno et al., 2010). Cold climate in the Pyrenees at 14-13.4 cal ka BP (González-Sampérez et al., 2006).
GI-1b (13.3-13.1)	Short cold reversal (1-2°C)	-	Short periods of colder and drier conditions (Moreno et al., 2010). Expansion of cold-climate forest species in the Pyrenees at 13.4-13.1 cal ka BP (González-Sampérez et al., 2006).
GI-1a (13.1-12.9)	Early warming followed by a general cooling	Glacier advance or stillstand with moraine formation at 13.0 ka	Cold and dry conditions in the Pyrenees starting at 13.1 cal ka BP (González-Sampérez et al., 2006).

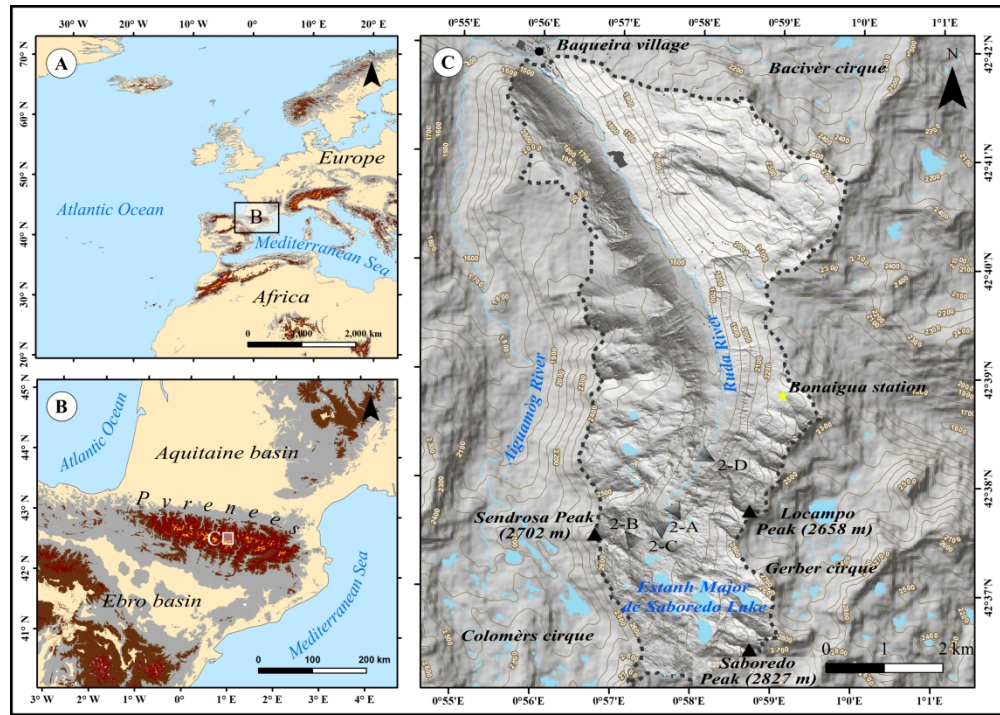


Figure 1. Geographical setting of the Ruda Valley. (A) Location in Europe. (B) Location in the Pyrenees with the distribution of the studies mentioned in the discussion (green stars). (C) Limits and topography of the valley with the distribution and view direction of the photos in Figure 2.

2476x1757mm (39 x 39 DPI)

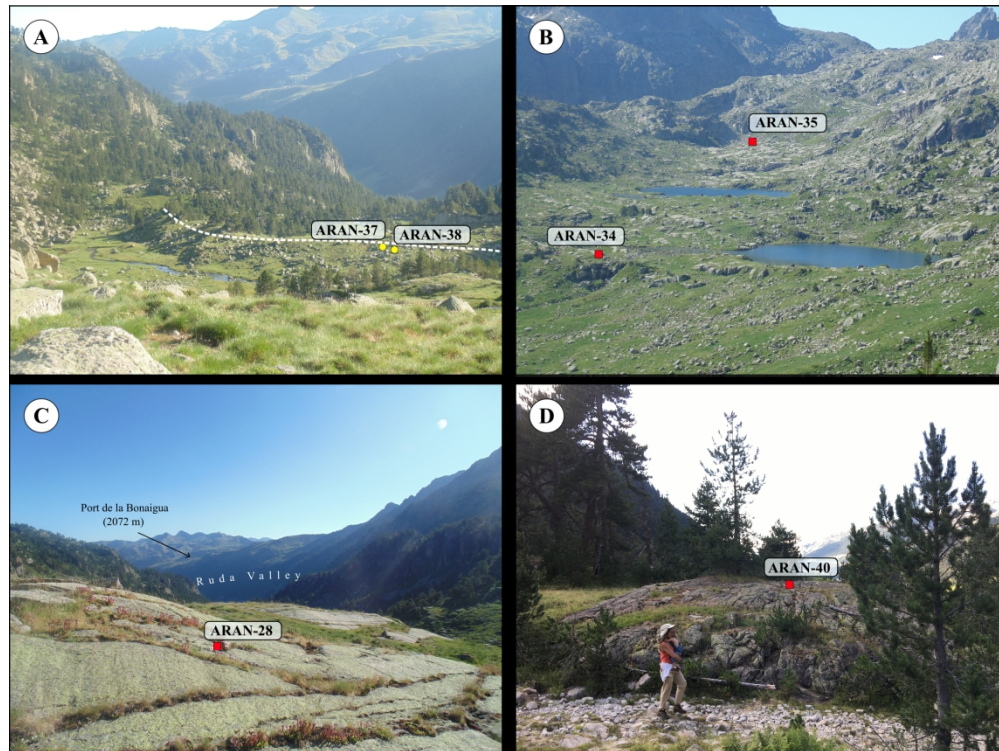
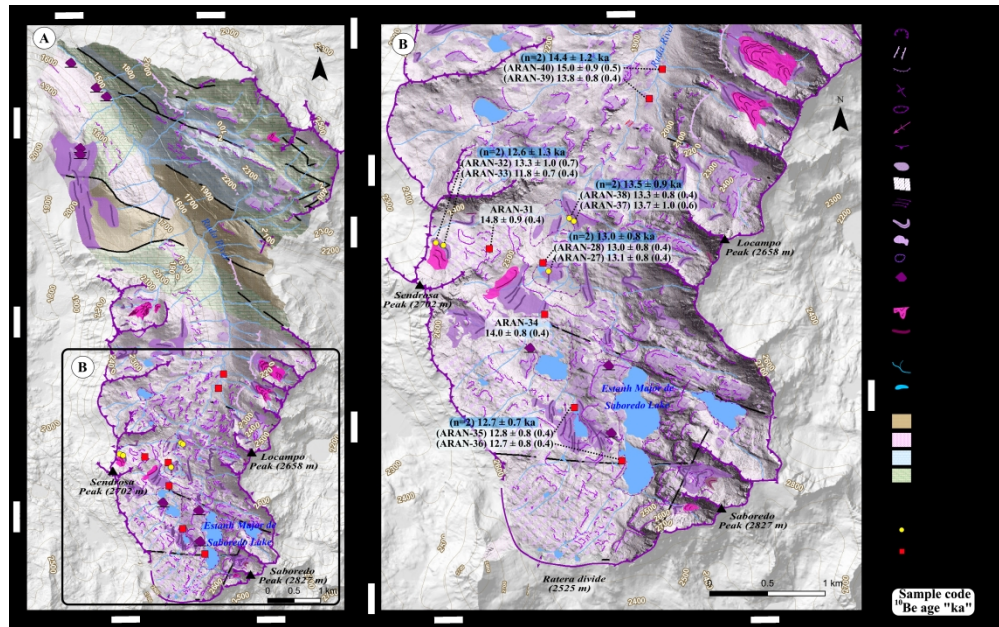


Figure 2. Examples of the main geomorphological landforms of glacial origin in the Ruda Valley, together with the location and typology of the dated samples: (A) terminal moraine damming a paleolake at 2080 m; (B) N-S perspective of the Saboredo Cirque floor showing several lakes distributed in staggered overdeepened basins; (C) polished bedrock surface including striations that reveal the former glacial flow; and (D) roches moutonnées located in the forested valley bottom of the valley at 1860 m.





3331x2073mm (39 x 39 DPI)

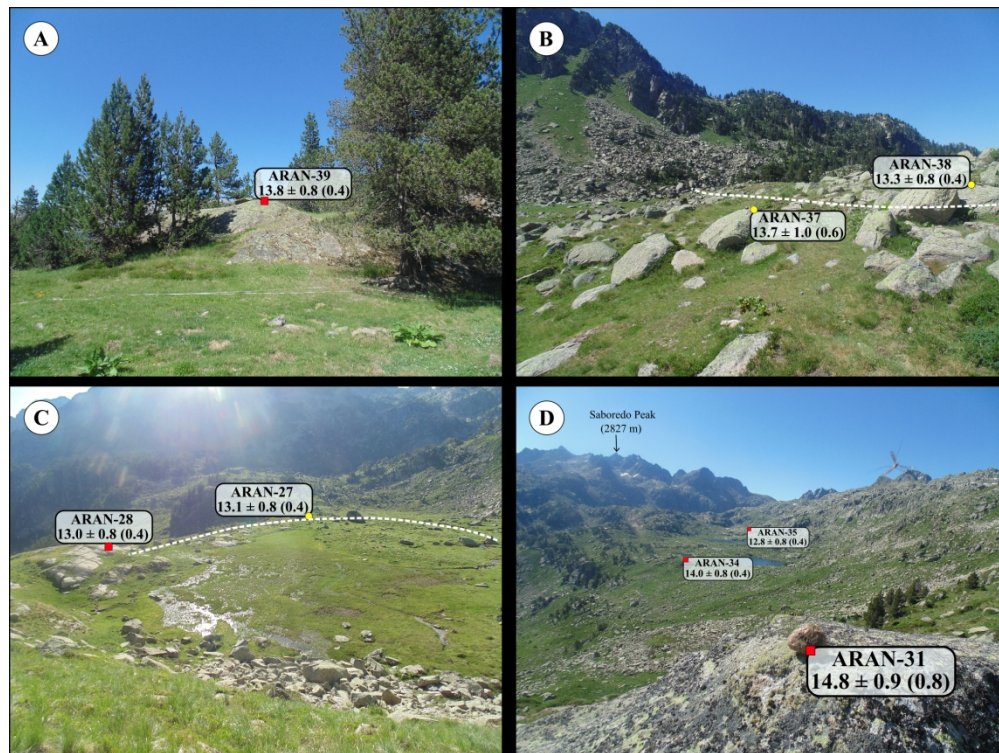


Figure 4. Examples of sampled moraines and polished bedrock surfaces, together with CRE ages (ka).

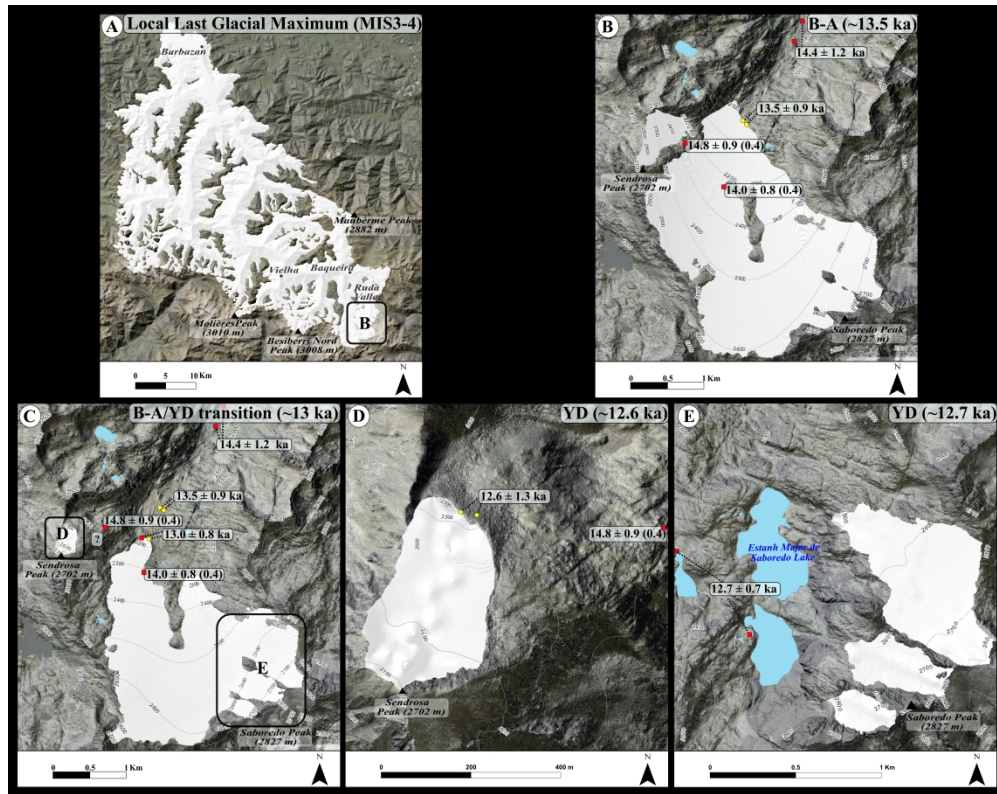


Figure 5. Reconstruction of the glacial extent during the different time stages: (A) Extensive glaciers occupying the entire Upper Garonne Basin during the ILGM based on the reconstruction of Fernandes et al. (2017); (B) Glacial extent during the mid B-A at  $\sim 13.5$  ka in the Ruda Valley; (C) Glacier flowing from the Saboredo Cirque during the transition between the late B-A and the early YD at  $\sim 13$  ka; and (D) Small glaciers confined within the highest hollows next to the Sendrosa (D) and Saboredo (E) peaks during the early YD at  $\sim 12.7/12.6$  ka.



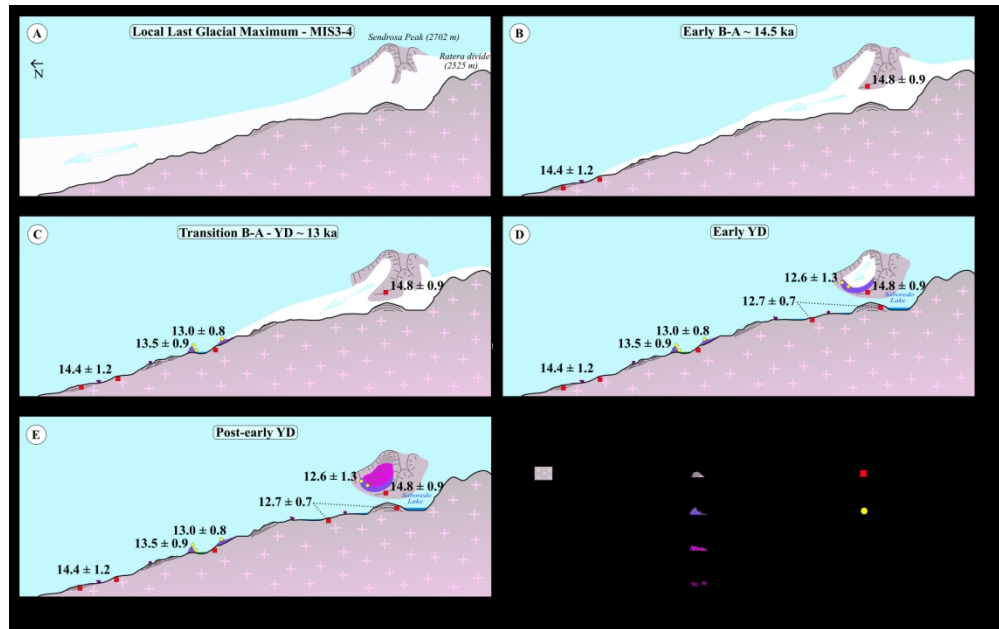


Figure 6. Schematic evolution of the glacial phases recorded in the Ruda Valley during the last glacial cycle: (A) extensive and thick glaciers inundated the valley and only few nunataks protruded the ice during the ILGM; (B) warm temperatures in the early B-A favoured glacial recession, and subsequently the lowest areas of the valley as well as some rock outcrops of the cirque became exposed; (C) glacial advances/standstills formed moraines during the B-A and at the transition to the YD; (D) small glaciers persisted during the early YD in the highest north-facing cirques; and (E) subsequent temperature increases favoured complete deglaciation and triggered paraglacial dynamics, which promoted the formation of rock glaciers in the recently deglaciated ice-free cirques.

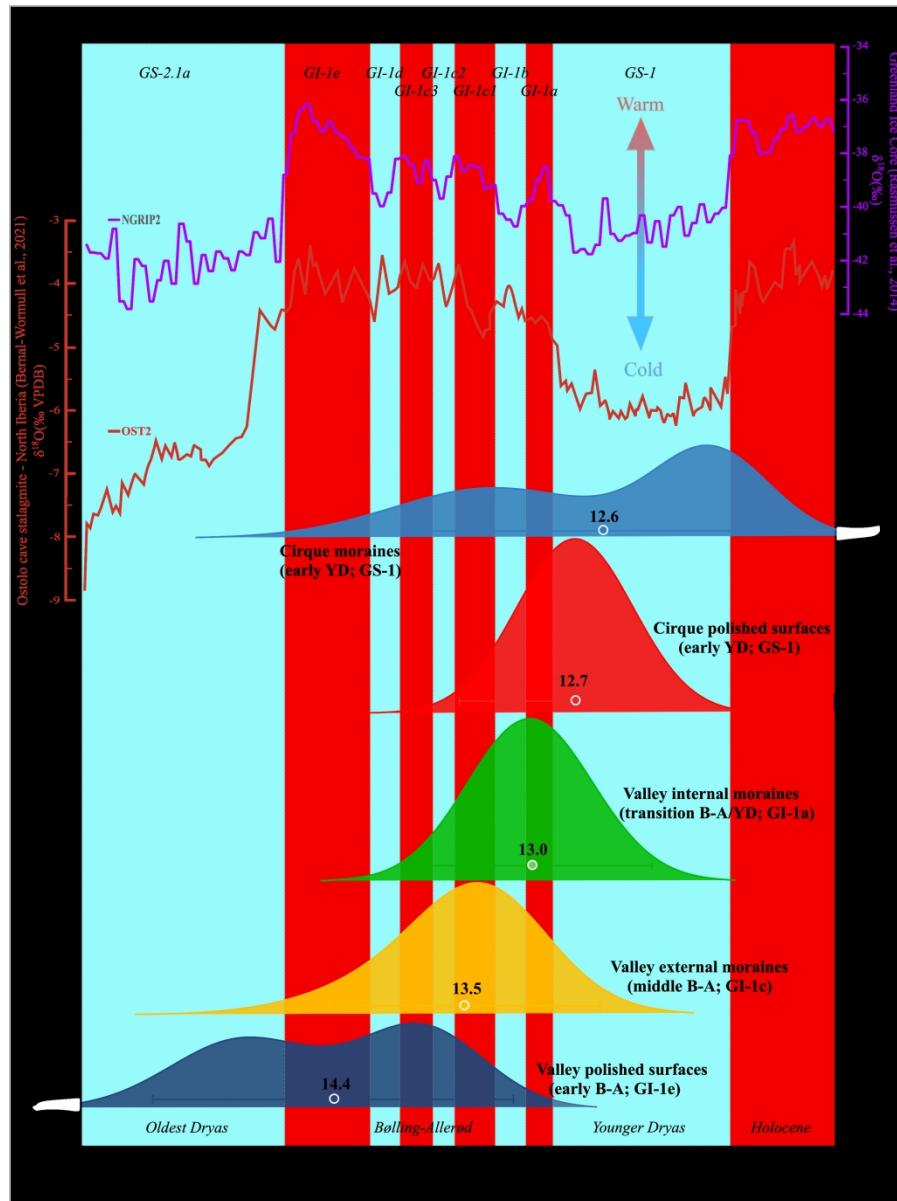


Figure 7. Normalized probability distribution functions (PDF) of surface exposure ages vs. temperature evolution from the OD to the early Holocene based on the  $\delta^{18}\text{O}$  record from the NGRIP ice core from Greenland and the  $\delta^{18}\text{O}$  record from the speleothem of the Ostolo cave in north Iberia (time periods are defined after Rasmussen et al. (2014)). The plots of the units result from the sum of the individual PDF of the samples belonging to them; cold (warm) phases are represented by the blue (red) bands.

670x895mm (118 x 118 DPI)