A review of glacial geomorphology and chronology in northern Spain: timing and regional variability during the last glacial cycle.

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Abstract

In this paper we synthesize the research in glacial geomorphology and geochronology in northern Spain, with special attention to the evidence of local glacier maximum extent earlier than the global LGM of MIS 2 (18-21 ka BP). More accurate models of glacier evolution have been defined based on limnogeological, geochronological and geomorphological data. In the Pyrenees, OSL (Optically Stimulated Luminescence), surface exposure and radiocarbon dating techniques have identified end moraines and fluvial terraces corresponding to MIS 6 (about 170 ka) and even to MIS 8 (about 260 ka), and also established the timing of the last local glacial maxima as prior to global LGM (MIS 4, ca. 50- 70 ka). During the global LGM a smaller re-advance occurred but glaciers reached different extent in the Central and the Eastern Pyrenees. In NW Iberia, radiocarbon and OSL techniques point to local glacial maxima prior to ca 26 ka-38 ka and probably synchronous to 45 ka. Although some bias might have been introduced by the dating procedures, this review demonstrates that in both regions the local maximum extent occurred prior to the global LGM. The asynchronies between the glacial maxima chronologies in the different northern Spain mountain ranges suggest that local climate factors exert a strong control of mountain glaciers dynamics.

Key words: glacial geomorphology, geochronology, limnogeology, AMS, OSL, exposure ages, Iberian Peninsula

1. Introduction

The spatial distribution of ice masses and the timing of glacier advances and retreats are difficult to constrain in mountain areas, since remnants of early glaciations are usually eroded by later, more extensive ice advances (Elhers and Gibbard, 2007) and glacial landscapes and deposits are commonly formed during more than one cold stage (Hughes et al., 2006a, 2006b, 2010). However, when preservation is adequate, landforms and deposits can record information on past climate changes due to glacier sensitivity and rapid response to moisture and temperature fluctuations (Ivy-Ochs et al., 2008). Accurate landscape and glacial evolution models have been obtained with a multidisciplinary strategy including sedimentological surveys, geomorphological mapping, quantitative estimations of Equilibrium Line Altitudes -ELA's-, and absolute dating.

The study of glaciations in SW Europe started in late 19th century but absolute chronological models have only been proposed during the last decades when an increasing number of studies focused on the glacial geomorphology and timing of the last glacial cycle. Chronological evidence of local maximum glacier advances in SW Europe mountains prior to the global Last Glacial Maximum (LGM) —ca. 18-21 ka BP (Yokohama et al., 2000; Mix et al., 2001; Ehlers and Gibbard, 2007) — were already reported in one of the first reviews of world mountain glaciers (Gillespie and Molnar, 1995). Specific reviews on Mediterranean and SW European Mountains (García-Ruiz et al., 2003; Hughes and Woodward, 2008; García Ruiz et al., 2010), defined two chronological scenarios for the maximum glacier extent: i) local glacial maxima closely synchronous with the global LGM (ca. 18-21 ka BP) is supported by cosmogenic surface dating from central-eastern Spain, Pyrenees, Maritime Alps, and Turkey; ii) local glacial maxima occurring several thousand years earlier than MIS 2 global LGM (50-80 ka BP) is supported by radiocarbon, U-series and OSL dates obtained in the Cantabrian Mountains, Pyrenees, Italian Apennines and Pindus Mountains.

In this paper, we focus on the available information on timing and extent of past glaciations in the Pyrenees and the Cantabrian – Galician Mountains and investigate how the Iberian case studies fit with global reconstructions. The Pyrenees is a mountain range ca. 420 km long and 150 km width, trending E-W, and located at latitudes between 43º and 42º N (Fig.1a), with peaks higher than 3000 m a.s.l. (Aneto Peak, 3404 m; Possets Peak 3371m; Monte Perdido Peak, 3355 m). The Cantabrian Mountain Range, trending E-W, is the western extension of the Pyrenees reaching as far as Galicia Mountains (Fig. 1b). It is 480 km long and 65 to 120 km wide, and it has an asymmetric altitude distribution progressively descending from South, with peaks higher than 2,600 m (Torrecerredo

Peak, 2648 m a.s.l.) to the North on the Cantabrian Coast. Both chains show well-preserved glacial features, which have been described by different geomorphologists since the end of the 19th century (Penck, 1883; Fernández-Duro, 1879). Chronological studies started in the second half of the last century, leading to the establishment of different models of glacial evolution that have been progressively improved applying several dating techniques. In this paper we review the glacial features and timing of the Maximum Ice Extent (MIE) in both mountain areas focusing on selected sites with detailed geomorphological studies and robust chronology (including some new dates for the Cantabrian Mountain Range) and discuss the evidence supporting a local glacial maximum (GM) earlier than the global LGM for the last glacial cycle.

<Fig. 1.>

2. The glacial record in the Pyrenees

In the Iberian Peninsula, Pleistocene glaciers reached the greatest extent in the Pyrenees. The relief is dominated by large cirques separated by acute divides above 2000 m a.s.l., and U-shaped valleys developed by > 30 km-long ice tongues. Glacial landforms and till deposits occur close to the divides and end and lateral moraines, glacio-lacustrine deposits, glacial thresholds, over-excavated basins and *roches moutonnées*, among other glacial features appear in most valleys. Since the pioneering studies in the late 19th (Penck, 1883) and early 20th century (Panzer, 1926), one of the main controversies has been the number of glaciations responsible for the observed glacial and fluvial deposits. The discussion is still alive, as recent chronological data identify different phases of glacier advance and retreat ascribed to several cold stages (Vidal-Romani et al., 1999; Peña et al., 2004; Lewis et al., 2009; Delmas et al., 2011; García-Ruiz et al., 2011). Regardless of the dating methods, available dates have demonstrated the occurrence of a MIE some thousands of years before the global LGM (Jiménez-Sánchez and Farias, 2002; García-Ruiz et al., 2003; Peña et al., 2004; Jalut et al., 2010; Lewis et al., 2009; Moreno et al., 2010; Pallàs et al., 2010; Delmas et al., 2011; García-Ruiz et al., 2011; Rodríguez-Rodríguez et al., 2011; among others).

2.1. The history of glacial research in the Pyrenees

We use the same three stages (Pioneer, Mapping, and Advanced) defined by (Hughes et al., 2006a) in the Mediterranean mountains to describe the history of glacial research in the Pyrenees. The Pioneer Stage started in the second half of the 19th century by Penck (1883), who described that glaciers in the southern slope of the Pyrenees had their terminal area at higher altitudes than those on the northern slope (800-1000 m and 400-600 m, respectively), and identified the main moraine landforms in the Aragón and Ara valleys. Panzer (1926) studied the frontal complex in the Aragón River Valley and suggested the presence of two different glacial stages attributed to the Riss and Würm glaciations. Panzer arrived to such conclusion because of the (apparent) connection between two main end moraines and two fluvial terrace levels (60 and 20 m). Llopis Lladó (1947), Fontboté (1948) and Nussbaum (1949) also postulated two glaciations in the Aragón and Gállego valleys. Barrère (1963) was the first to argue that only one glaciation was present in the Central Pyrenees (Riss), whereas the Würm would correspond to moraines located relatively close to the headwaters.

The Mapping Stage started with the excellent geomorphological maps produced by Barrère (1971), who gave a synthetic perspective of the location of the main glacial deposits and the extent of the glaciers in the Aragón, Gállego and Ara basins. Fieldwork and geomorphological mapping advanced the knowledge of the Pyrenean glaciers during the last decades of the 20th century: Martí-Bono (1973), Serrat et al. (1983), Martínez de Pisón (1989), Vidal-Bardán and Sánchez-Carpintero (1990), García Ruiz et al. (1992), Serrano and Martínez de Pisón (1994), Martí-Bono (1996), Serrano (1998), García Ruiz et al. (2000), and García-Ruiz and Martí-Bono (2002); Serrano et al. (2002), and García-Ruiz and Martí-Bono (2011). Finally, in an advanced stage, the classification of till deposits according to their shape, position, pedological development and relationships with fluvial terraces, detailed geomorphological surveys and absolute age control allowed definition of several glacial stages in both the Central and Eastern Pyrenees:

1. A pre-Würm stage with till remnants in a tributary of the Gállego Valley about 8 km downstream from the main ice tongue (Martí-Bono, 1996; Serrano, 1998).

2. The MIE, generally represented by two or, more commonly, three lateral moraines, in the Aragón-Subordán (García-Ruiz and Martí-Bono, 2011), Gállego (Barrère, 1971; Serrano, 1998; Peña et al., 2004), Ara (Serrano and Martínez de Pisón, 1994; García-Ruiz and Martí-Bono, 2002) and Ésera valleys (Martínez de Pisón, 1989; Bordonau, 1992; García-Ruiz et al., 1992). Most of ice- or moraine-dammed glaciolacustrine deposits coincided in time with the MIE, when tributary streams with high discharges developed small lakes in their final stretch, lately filled with a complex of fluvial, torrential and glaciolacustrine sediments (Serrat et al., 1983; Bordonau, 1992; Sancho et al., 2011)

3. A number of frontal, lateral moraines and till deposits representing ice tongues of progressively smaller size, including an important advance coinciding with the LGM. They have been studied in detail in the Escarra Valley, a tributary of the Gállego glacier (Martínez de Pisón and Serrano, 1998; García-Ruiz et al., 2003), the Aragón Subordán Valley (García-Ruiz and Martí-Bono, 2011), and the Ésera Valley (Bordonau, 1992). Some of the deposits show a clear separation between the main glacier and its tributaries; others represent a new, short and limited advance during the Late Glacial, and finally, others represent a minor advance usually attributed to Younger Dryas.

2.2. Selected sites of Pyrenees

The main features of the southern slope Pyrenean glaciers have been already described several decades ago, but the chronology has remained unsolved until recent times. Several glacial stages were defined even at the beginning of the 20^{th} century, and then refined in the 1950's and 1960's. The main terminal basins (Aragón, Gállego and Querol) were identified early, with detailed descriptions of the end and lateral moraines and their relationships with the fluvial terraces. After the 1980's, absolute dating techniques were applied to those records, first conventional ^{14}C, and later ^{14}C AMS, OSL techniques and surface exposure ages in sediments, blocks and polished surfaces.

Mardones and Jalut (1983) obtained some ^{14}C dates in the French Pyrenees from a lacustrine sequence in the moraine complex of Lourdes (Pau Valley); the lacustrine sequence was dated up to

38,400 uncal. yr BP. They extrapolated a basal age of 45,000 years and attributed an age between 50,000 and 70,000 years to the moraines that enclosed the lacustrine basin. This was the first time that an asynchroneity between the MIE and the global LGM was reported for the Pyrenees and even for Europe. Few years later, other studies obtained ^{14}C dates older than 30,000 yr for glacier-related sediments in both the French Central Massif (Etlicher and De Goer de Hervé (1988), and the Vosgues (Seret et al., 1990). In the Spanish Pyrenees, Montserrat (1992) dated with radiocarbon the bottom of the glaciolacustrine deposit of Tramacastilla, Gállego Valley, at 29,400 ± 600 yr BP. Lateral moraines located about 100 m above the lake indicated that the MIE would have occurred thousands of years before (García-Ruiz et al., 2003). Nevertheless, other authors argued that the radiocarbon dates from Mardones and Jalut (1983) were affected by the hard-water effect or by mixture of re-worked organic matter (Turner and Hannon, 1988; Pallàs et al., 2006) and the controversy started in the late 1980s.

Some areas located in the headwaters of the Gállego, Aragón, Cinca and Ara Rivers have been selected as examples that are representative of the geomorphological and geochronological glacial record of the Pyrenees (Fig. 1a).

2.2.1. The Gállego River Valley

Recent studies on glacial chronology in the Gállego Valley have focused on both the terminal basin (the Senegüé basin) and the headwaters. The Senegüé terminal basin developed within the Eocene Flysch and marl formations of the Inner Pyrenean Depression; the lowest moraine remnants occur at about 800 m a.s.l. When the ice tongue reached this basin, it was 400 m thick. The lateral moraines, represented by two or three ridges, indicate a relatively rapid thinning towards the terminal area. Two till outcrops and other remnants of glacial deposits, as scattered sub-rounded blocks, occur close to the village of Aurín, near Sabiñánigo. Barrère (1963) had pointed out that no evidence of more than one glaciation could be found in the Gállego Valley, because the end moraines in the Senegüé basin only connected with a 20 m fluvial terrace, whereas the 60 m fluvial terrace and pediment were disconnected from the glacial deposits. Recent studies from Peña et al. (2004) have contributed to clarify the problem. Using the OSL in sandy levels interbedded with till, Peña et al. (2004) dated the Aurín end moraine at 85 ± 5 ka. This age, however seems too old and “out of the stratigraphic context” because the proglacial fluvial terrace that starts immediately downstrean from this till was dated at 69 ± 8 ka and the same terrace level was dated at 66 ± 4 ka about 20 km downstream. Considering all the available ages, the age of the end moraine of Aurín is likely between 66 and 69 ka (MIS 4); six kilometres upstream from the Aurín area, the Gállego glacier deposited a big, arched and transverse end moraine, called the Senegüé moraine, during another cold stage at 35 ± 3 and 36 ± 2 ka (Peña et al., 2004; Lewis et al., 2009).

Remains of older glaciations have been recently dated in the Gállego valley. A fluvioglacial terrace at Sabiñánigo, disconnected from any till deposit, has been dated with OSL at 155 and 156 ka (MIS 6). Glacial deposits ascribed to MIS 6 have also been recognized and dated in Greece and the Balkans (Hughes et al., 2006b, 2010), well correlated with fluvial terraces for the Middle and Late Pleistocene in Greece (Woodward et al., 2008).

The evolution of the Gállego glacier has also been studied in the headwaters. The El Portalet peat bog is located in an over carved glacial cirque in a relatively low relief area close to the main Pyrenean divide at 1802 m a.s.l., surrounded by 2100-2300 m a.s.l. peaks. When the ice melted, a moraine-dammed lake developed at the bottom of the cirque. Sedimentation in post-glacial times resulted in the infilling of the lake and the development of the actual peat bog in late Holocene times. González-Sampériz et al. (2006) dated with ^{14}C AMS the base of the 8 m long lacustrine sequence deposited after the glacier retreat, at 32,183 – 33,773 cal. yr BP. This was somewhat surprising, since it would mean that the headwater was totally or partially deglaciated very early, and that by that time the Gállego glacier would be mainly fed from the Aguas Limpias glacier, a main tributary whose headwater is surrounded by peaks higher than 3,000 m. The lacustrine sequence shows a hiatus dated as 22,953 ± 120 cal. yr BP and interpreted as evidence for a glacier re-advance coinciding with the global LGM. Another nearby peatbog, known as Formigal peat bog, developed in an ephemeral lake basin originated by a deep-seated landslide that temporarily blocked the course of the Gállego River (García-Ruiz et al., 2003) has also yielded a similar basal age (24,070 ± 170 cal. yr BP). Individual organic matter fragments and concentrated pollen were used in both sequences for radiocarbon dating to minimize the hard-water effect and contamination from old carbon sources. The chronology for both sequences indicates that the Gállego glacier was spatially restricted to headwater areas during the global LGM of MIS 2.

2.2.2. The Aragón River Valley

The terminal basin of the Aragón Valley (the Villanúa basin) is one of the best-studied in the Pyrenees. The Aragón glacier ended in a large basin in the Eocene Flysch Sector, with lateral moraines progressively at lower altitudes, indicating a thinning of the ice tongue due to the enlargement of the valley and the increasing temperatures at lower altitudes. Three lateral moraines flank both sides of the basin, damming small lakes in the tributaries (Fig. 2.). Up to six frontal moraines appear in the valley floor within a distance of 3 km. Their age and glacial phase has been a matter of debate for almost a century. Two of the six arcs are larger (M1 and M2) than the rest (m1, m2, m3, m4). The outermost arc (M1) forms an impressive hill transverse to the valley and it seems to connect with the 60 m fluvial terrace, as pointed out by Panzer (1926) and Llopis-Lladó (1947). M2 links with the 20 m fluvial terrace. The inner arcs (m2, m3, m4) connect with the lower fluvial terrace level (7-8 m). Barrère (1963) considered that the M1 moraine did not connect with the 60 m terrace, but it was leaning against the pre-existing scarp of the terrace, and rejected the idea of a link between the two sedimentary bodies. The morphometric characteristics of clasts in the 60 m terrace (flatness and roundness indices) led to Höllermann (1971) and Martí-Bono (1973) to conclude that the terrace was fluvioglacial and deposited during a period of glacier progression. Therefore, the two larger moraines in the terminal basin of the Aragón glacier would correspond to only one glaciation, and the 60 m terrace was formed during an older glaciation. Vidal-Bardán and Sánchez-Carpintero (1990) based on a soil development analyses in fluvial terraces and tills also suggested there was no connection between M1 and the 60 m terrace.

García-Ruiz et al. (2011) proposed three main stages using OSL dating on quartz grains from fluvial sand levels (Fig. 2): (i) The M1 moraine was deposited at 171 ± 22 ka (MIS 6); (ii) the M2 moraine was deposited at 68 ± 7 ka (MIS 4), synchronous to the 20 m fluvial terrace connected to the moraine; the m2 moraine was dated at 51 ± 4 ka and would also correspond to this stage; and (iii) the 60 m fluvial terrace was dated at 263 ± 21 ka, so it was deposited during a previous glacial stage (MIS 8). In such a case, no stratigraphical connection exists between M1 and the 60 m terrace. Nevertheless, it is important to note that the sample from the 60 m terrace was taken in a tributary of the Aragón River (The Aragón Subordán River), 23 km downstream of the ice front.

The Aragón Valley glacial chronology provides a new perspective of the Late Pleistocene glaciations in the Pyrenees: it solves the problem of the number of glaciations represented in the Villanúa basin, and confirms the occurrence of the MIE several thousand years before the global LGM.

<Fig. 2. >

2.2. The Central and Eastern Pyrenees valleys

The headwaters of the Cinca River are in the northern slopes of the Monte Perdido Massif, at more than 3,000 m a.s.l., and close to some of the last glacier remains in the Pyrenees. During the Late Pleistocene, a large glacial tongue flowed in the upper Cinca Valley, reaching the confluence with the Cinqueta Valley, at about 780 m a.s.l. This end moraine connected with a fluvial terrace dated with OSL at 62.7 ± 3.9 ka (Lewis et al., 2009). This age is similar to a related fluvial terrace in the Cinca River, dated in nine different locations, with an OSL averaged age of ca. 64.4 ka. This age of maximum glacier extent is similar to the moraine M2 in the Villanúa basin (Aragón Valley). Loess deposits in the Cinca Valley dated about 20 ± 3 ka (OSL) also suggest that during the global LGM the climate was colder than before (Lewis et al., 2009). Another evidence for colder climate during the global LGM is the development of stratified screes in the same Cinca Valley (Devotas Canyon), with a basal age of 22,800 ± 200 cal. yr BP (García-Ruiz et al., 2001).

The Ara Valley, between the Aragón and the Cinca valleys, also has provided evidence of an early MIE in the Pyrenees. The Ara Valley developed a thick ice tongue whose lateral moraines dammed most of the fluvial tributary valleys before its terminal area close to Sarvisé. The largest ice-dammed valley was Sorrosal, where the Linás de Broto lake was progressively infilled with torrential and glaciolacustrine sediments. The lateral moraine that dammed the lake and the glaciolacustrine deposit yielded similar OSL ages (49±8 ka for the moraine and between 49±11 and 55±9 ka for the lake sediments, Sancho et al., 2011).

The results in the western and central Pyrenees using two different dating techniques (^{14}C AMS and OSL) confirmed the asynchroneity between the MIE in the southern Pyrenean glaciers and the global LGM. Recently, the use of cosmogenic surface dating methods in the Central and Eastern Pyrenees re-opened the discussion. In the Upper Ribagorzana Valley Pallàs et al. (2006) analyzed 25 erosive surfaces and granodiorite blocks and dated the end moraine at 21.3± 4.4 ka ^{10}Be, and the inner moraines between 16 and 11 ka ^{10}Be. In the Têt Valley (northern side of the Eastern Pyrenees) Delmas et al. (2008) reported similar extents for the ice tongues during MIS 2, 3, 4 and 5. In the Malniu Valley (tributary of the Querol Valley, in the southern face of the Eastern Pyrenees) Pallàs et al. (2010) reported the occurrence of the MIE at 49.2 ± 1.3 ka ^{10}Be, with a similar glacial advance during MIS 2. However, recently in the Ariège Valley, Delmas et al. (2011, 2012) dated the MIE at 79.9 ± 14.3 ka ^{10}Be. In addition, in the Andorra valleys, Turu i Michels et al. (2011) dated the MIE at 59 ± 1.18 ka ^{21}Ne. These surface exposure ages are in agreement with glacier chronologies derived from OSL and ^{14}C AMS techniques that proposed a MIE prior to global LGM.

3. The glacial record in NW Iberia

3.1. The history of glacial research in NW Spain

The glacial record of NW Iberia has been studied since the end of the 19th century, starting with pioneering contributions on glacial geomorphology in the Trevinca-Sanabria area (Fernández Duro, 1879) that included the first qualitative descriptions, sketches, and hypothesis about glacial evolution (Pioneer Stage).

However it was not until the last two decades of the 20th century, when a reactivation of the interest on glacial geomorphology of NW Iberia took place, initiating the Mapping Stage. These new studies focused on the qualitative and quantitative description of glacial deposits and erosion landforms. Some of them reported the first ELA altitude estimations and ice flow patterns, based on fieldwork, photointerpretation, and the development of geomorphological maps, as well as morphometric analyses (Arenillas and Alonso, 1981; Flor and Baylón, 1989; Suárez-Rodríguez 1990; Castañón and Frochoso, 1992a, 1992b; Alonso, 1993, 1998). Using geomorphological models initially established in the Pyrenees (Vilaplana, 1983; Bordonau et al., 1992) the MIE ice geometry in the area was reconstructed and retreat and stabilization stages were proposed for the last glacial cycle although field observations did not rule out the occurrence of deposits from previous glaciations (Jiménez-Sánchez, 1996; Menéndez-Duarte and Marquínez, 1996). The last remnants of glacial ice in the Cantabrian Mountains were described (González and Alonso, 1994) and attributed to the Little Ice Age (LIA) (Alonso and Gonzalez, 1998). Other contributions focus on ice flow patterns reconstructions during the MIE (Rodríguez-Guitián et al., 1995; Marquínez and Adrados, 2000; Flor, 2004) and the identification of evidence of the Little Ice Age in Picos de Europa Area (González Trueba, 2005, 2007).

As in the Pyrenees, the Pioneer Stage of knowledge in NW Iberia started in the 19th century, but the Advanced Stage was reached later than in the Pyrenees. Because of this, during the Mapping Stage it had been common to assume the initial Pyrenean chronologies as model for NW Iberia glacial evolution. The first absolute geochronological data for the Cantabrian Mountain (Redes Natural Park and Picos de Europa National Park) were not available up to this century (Jiménez-Sánchez and Farias, 2002). Recent advances on glacial evolution in NW Iberia involve the use of Geographic Information Systems (GIS) for quantitative glacial reconstructions and the integration of limnogeological studies with geomorphological models and geochronological data (Cowton et al., 2009; Moreno et al., 2010; Rodríguez-Rodríguez et al., 2011). As shown below, these studies point to a local glacial maximum older than the global LGM.

3.2. Selected sites of NW Iberia

Three sites of NW Iberia have been selected as examples of geomorphological and geochronological glacial records, namely (from W to E): Trevinca Massif, Redes Natural Park and Western Massif of Picos de Europa. Five new dates are included in this review for the last two sites.

3.2.1. The Trevinca Massif

The Trevinca Massif is one of the most western mountain areas of northern Spain, located southwest of the Cantabrian Mountain Range and close to the Spain – Portugal border (42º10'N and 6º50' W) (Fig. 1b). The bedrock is made up of Cambrian to Ordovician igneous and metamorphic rocks, intensely fractured (Ollo de Sapo Domain; Díez-Montes, 2006). The highlands are characterized by a smooth topography with altitude ranging between 1600-2100 m a.s.l. (Peña Trevinca, 2128 m a.s.l.). Several glacial U-shaped valleys cut the margins of this high plateau, and glacial cirques are scarce. The main glacial valleys of the Trevinca Massif are the Bibei and Barxacoba in the west side, and Segundera-Cárdena and Tera in the east side (Sanabria sector). End and lateral moraines are locally well-preserved in both sides of the massif, as well as ice contact and fluvioglacial sediments at altitudes as low as 940 m a.s.l.

The glacial record of the Trevinca Massif has been studied since the end of the 19th and during the early 20th centuries (Fernández Duro, 1879; Taboada, 1913; Halbfass, 1913; Stickel, 1929). Taboada (1913) attributed the glacial landforms and deposits in the Sanabria sector to the Riss and Würm glacial cycles. Subsequently, Llopis-Lladó (1957) ascribed the local glacial evolution to three glacial cycles (Mindel, Riss and Würm), and subdivided the younger one in a glacial maximum stage (Würm I) and two epiglacial recession stages (Würm II and III) based on the weathering grade of till deposits and their correlation with the adjacent fluvial terrace levels. The first palynological studies carried out on the lacustrine deposits located close to the Sanabria Lake (Laguna de las Sanguijuelas and Sanabrian Marsh ponds) and on the Trevinca highlands (Laguna de la Roya and Laguna Cárdena ponds) only provided Holocene records (Menéndez-Amor and Florschütz, 1961; Allen et al., 1996; Muñoz-Sobrino et al., 2004).

Only recently glacial evolution reconstructions, calculation of ELA´s, and local MIE-LGM chronologies are available (Rodríguez-Guitián et al., 1995; Cowton et al., 2009; Rodríguez-Rodríguez et al., 2011; Pérez-Alberti et al., 2011). During the local MIE, the highlands of the Trevinca Massif were covered by an ice cap more than 440 km^2 and with the ELA located at an altitude of 1500-1600 m (Cowton et al., 2009). The ice cap was drained by outlet glaciers more than 20 km long through the Bibei, Barxacoba, Segundera-Cárdena, and Tera glacial valleys, placing the glacial fronts at altitudes between 990 and 940 m a.s.l. in the west and east sides of the massif, respectively (Rodríguez-Guitián and Valcárcel, 1994; Rodríguez-Rodríguez et al., 2011). Ice thickness for the local MIE in the Sanabria sector was in the range of 200-300 m on the flat uplands up to 454 m at the Tera and Segundera-Cárdena valleys confluence (Rodríguez-Rodríguez et al., 2011). In Bibei Valley, ice thickness was up to 500 m, based on altitude differences between the valley bottom and the moraines preserved at Cepedelo site (Rodríguez-Guitián and Valcárcel, 1994). Recent geomorphological, sedimentological and chronological studies carried out on Pías and Sanabria sites have provided the first dates in the Trevinca Massif.

Pías site is a glacial deposit located close to the confluence between the Bibei and Barxacoba glacial valleys (1010 m a.s.l.), in a quarry placed south of the Pías Reservoir (Pérez-Alberti et al., 2011). The sediment outcrop consists of three units deposited in a post-MIE phase. The lower one is made up of fine grained sediments, with some intercalations of coarser detrital material deposited in a fluvioglacial environment. The middle unit is composed of two diamicton levels —indicative of two advances of the Bibei glacier— separated by fluvioglacial sands and gravels deposited in a proglacial environment downstream the Bibei glacial front. Finally, the upper unit is made up of waterlaid deposits with some diamictite levels, overlain by organic rich deposits. This upper unit has been interpreted as a braided-type fluvioglaciar to fluvial environment. Three OSL samples from the sandy levels of the lower unit have reported ages of 27 ± 2, 31 ± 3, and 33 ± 3 ka (Pérez-Alberti et al., 2011), indicative of a local MIE older than 30 ka for the west side of the Trevinca Massif.

<Fig. 3. >

Sanabria site presents a moraine complex located at the end of the Tera Valley constituted by several lateral moraines more than 6 km long, nine frontal moraines, and some undifferentiated till deposits (Rodríguez-Rodríguez et al., 2011). Geomorphological evidence supports a deglaciation with 10 episodes of glacier front retreat and stabilization after the local MIE. Sanabria Lake is located in the inner side of this moraine complex and occupies a glacial over-deepening depression (3.5 km^2) at 1000 m a.s.l. The bottom of the lacustrine sedimentary sequence shows massive to banded sands and silts interpreted as proglacial lake deposits. Two ^{14}C AMS dates from the base and the top of this basal clastic unit constrain its age between 25,584±374 and 14,494±347 cal. yr BP (dates from bulk sediment and terrestrial plant remains, respectively). The Sanabria sequence also reveals an episode of glacial re-advance between 13.1 and 12.3 cal. ka BP with deposition of more clastic facies before the definitive onset of organic-rich sedimentation prevailing during the Holocene. Another core (SAN08-SM1) retrieved from an ice-dammed deposit formed behind the MIE lateral moraine at San Martin de Castañeda has provided a ^{14}C AMS minimum age of 21,833±358 cal. yr BP for the local MIE. Since both cores did not reach the base of the proglacial sequence, the local MIE in the east side of the Trevinca Massif should be older than the base of the recovered Sanabria sequence (25.6 cal. ka BP) (Rodríguez-Rodríguez et al., 2011).

<Fig. 4. >

Taken together the OSL and ^{14}C AMS data from both the Pias and the Sanabria sites, the MIE of the Trevinca Massif would have taken place prior to the global LGM.

3.2.2. Redes Natural Park

Redes Natural Park is located on the headwaters of the Nalón River, in the Northern slope of the Cantabrian Mountain Range (Fig. 1b) (43º13’N, 4º58’O). The bedrock consists of Cambrian to Ordovician sedimentary rocks including limestone, quartzite sandstone and alternations of silt, lutite and sandstone materials (Álvarez-Marrón et al., 1989). The main mountain range (Torres Peak, 2104 m a.s.l.) is located towards the South and contains numerous glacial landforms and deposits as cirques, valleys, end and lateral moraines, tills, ice-dammed and fluvioglacial deposits above 930-1100 m altitude.

Previous research (Jiménez-Sánchez, 1996; Jiménez-Sánchez and Farias, 2002; Jiménez-Sánchez et al., 2002) allowed to establish a local GM phase (Phase 1) followed by two subsequent phases of glacial front retreat and stabilization (Phases 2 and 3). During the MIE glacier fronts descended to 1300-950 m (increasing from E to W), and glacier tongues extended up to 5 km-long (Phase 1). End moraines evidence the stabilization of glacial fronts first at 1300-1500 m (Phase 2) and afterwards at 1500-1700 m (Phase 3).

Two areas of Redes Natural Park have provided detailed chronologies for the local GM: Monasterio and Tarna valleys (Fig1b). Four lateral moraines trending N-S appear in Vega de Brañagallones area, in Monasterio Valley, at altitudes between 1200 and 1250 m a.s.l. (Fig. 5.) During the local MIE (Phase 1), an alpine glacier flowed from S to N and developed a set of lateral moraines that blocked a tributary stream to the East, forming the ice-dammed deposit of the Vega de Brañagallones (0.60 km^2 surface). A 36.7 m-long core (S1) drilled in 1998 showed a lacustrine clastic sequence, and a bulk sediment sample close to the base (35.6 - 35.5 m depth) gave an age of 28,990 ± 230 uncal. yr BP (Jiménez-Sánchez and Farias, 2002). If we calibrate this age using the CalPal Software and the CalPal 2007 Hulu curve (Weninger et al., 2007; Weninger and Jöris, 2008) gives a calibrate age of 33,485±362 cal. ka BP that represents a minimum age for the local GM. . The remaining lateral moraines of this set represent smaller glacier fluctuation during its retreat and thinning after the local GM. Therefore, the age of these moraines could provide an upper limit for the local GM. An outcrop of the second outermost lateral moraine (Fig. 5) was sampled for OSL dating (Brañag-1). The OSL analysis on quarzt grains of the sandy fraction of the till matrix gave a result of 23,967±1841 yr (Table 1). Therefore, the local GM was coeval or younger than 29 ka in Monasterio Valley and the glacial retreat after local GM was already taking place at least at about ca 24 ka.

<Fig. 5. & Table 1 >

The Tarna Valley (Nalón River headwater) is located to the east of Redes Natural Park and extents from the Tarna Pass (1490 m a.s.l.) to 850 m a.s.l. Glacial evidence includes till deposits, lateral moraines and polished surfaces in quartzite bedrock at the head of the valley. The glaciers reached as low as 930 m a.s.l. in this valley during local MIE (Phase 1). Fluvioglacial deposition and complex landslides occurred during subsequent glacial retreat. Landslides affected both the Paleozoic bedrock and the till and fluvioglacial deposits, and the age of their placement could provide a minimum age for glacial retreat between the local GM phase and the subsequent Phase 2. A new OSL sample (Tarna-1) from an outcrop of one of these complex landslides gave an age of 22,983 ± 2,321 yr (Fig. 6 and Table 1). At the headwaters of the same valley (1415 m a.s.l.), a peat bog developed in a glacial hollow originated by ice during Phase 2 provided a basal age of 20,640 ± 300 uncal. yr BP (Jiménez-Sánchez and Farias, 2002) —calibrated in this work at 24,579±421 cal. ka BP with the CalPal Software and the CalPal 2007 Hulu curve (Weninger et al., 2007; Weninger and Jöris, 2008)—. This age represents as a minimum age for Phase 2 in the area (Fig. 6). Both new dates from Tarna Valley support a local GM prior to global LGM.

<**Fig. 6.** >

To sum up, the available glacial chronologies of Redes Natural Park suggest that: i) local GM is coeval or younger than ca 33.5 cal. ka BP; ii) at ca 24.6 cal. ka BP, glaciers would have already retired at altitudes higher than 1430 m.

3.2.3. Picos de Europa: Enol-Comeya area

The Picos de Europa, included in the Cantabrian Mountain Range, is one of the most extensive calcareous massifs in the world. It is formed by an imbricate thrust system of Variscan age, piling up a > 1000 m thick stack of Carboniferous carbonate series. The partial alpine reactivation of these thrusts and the development of new E-W faults (Marquínez, 1992; Alonso et al., 1996) are responsible of the current landscape, with peaks higher than 2600 m above sea level only 28 km far from the Cantabrian Sea coast. Glacial and karstic landforms are the most outstanding landscape features of these mountains. Four S to N flowing rivers have carved deep and steep valleys and canyons, and divide this alpine relief, into Western, Central and Eastern massifs.

Glacial features were described in Picos de Europa Mountains early in the 20th century (Hernández Pacheco, 1914; Obermaier, 1914). Glacial deposits and erosion forms, ELA altitudes, ice flow patterns and glacial reconstructions for the last glacial cycle are relatively well-known (Miotke, 1968; Smart, 1986; Farias et al., 1996; Gale and Hoare, 1997; Alonso, 1998; Marquínez and Adrados, 2000; Jiménez-Sánchez and Farias, 2002; Flor, 2004; González Trueba, 2007; Moreno et al., 2010). Most authors agree that during the local MIE, glacial ice would have covered the highest areas of the Picos de Europa and flowed in a radial pattern. Nevertheless, some glacial landforms could have originated during a previous glaciation (Obermaier, 1914; Gale and Hoare, 1997; Flor, 2004).

The Covadonga Lakes —Enol and Ercina, (43º 16'N, 5ºW, 1030 m asl) — located in the Western Massif of the Picos de Europa Mountains contain a unique record of glacial activity. The location of the lakes is controlled by the lithology —contact between carboniferous limestone and siliciclastic rocks (Marquínez, 1989) — and glacial history — over carved basins with the occurrence of end and lateral moraines —. Glacial arêtes, end and lateral moraines and glacial valleys as Enol and Bricial valleys are well preserved in the area and some rockfall and avalanche deposits are indicative of gravity processes after ice retreat.

The Enol Lake (0.08 km^2) occupies a glacial hollow eroded by a glacier tongue during the local GM. This glacier was initially stabilized at 1030m, as marked by till deposits, and subsequently retreated towards the South (Fig. 7a). Cores drilled in the Enol Lake in 2004 reached proglacial sediments deposited in the lake during the glacial retreat. The recovered proglacial sequence spans from 37,151-38,729 cal. yr BP til 26 ka (Moreno et al., 2010). Between ca. 26 to 18 ka BP, glacial dynamics was still greatly influencing lake sedimentation (glaciolacustrine environment) because of the close proximity of the ice. The Enol chronology demonstrates that the onset of proglacial sedimentation in Enol Lake after glacial retreat took place at least ca 38 ka BP, earlier than the global LGM (Moreno et al., 2010).

<Fig. 7. >

The Comeya hollow (also reported as Comella) is a 1.2 km^2 basin of smooth relief located 700 m downstream from the Enol Lake (Fig. 7b). Initially, a tectonic origin was suggested as an E-W fault- bound graben filled by sediments (Farias et al., 1996). Recently Flor (2004) proposed a glacial origin due to erosion during a previous glaciation. Two cores (SC1 and SC2; 42.5 and 56.7 m-depth, respectively) were drilled in Comeya in 1991. The sedimentary sequence shows torrential and proglacial lacustrine sediments derived from the ablation waters of Enol-Ercina glacier front located at 1030 m altitude (Fig. 8). The SC2 core sequence consists of five main stratigraphic units (Farias et al., 1996; Jiménez-Sánchez and Farias, 2002). A level of lacustrine organic–rich sediments at 35.5 m depth provided the first radiometric ages for the region: 40,480±820 uncal. yr BP (Jiménez-Sánchez and Farias, 2002) —44,118±885 cal. ka BP with the CalPal Software and the CalPal 2007 Hulu curve (Weninger et al., 2007; Weninger and Jöris, 2008) (Fig. 8) —. The sedimentation of lacustrine facies in Comeya hollow after 40 ka BP would have been synchronous with the Enol Glacier ablation during the local GM (Jiménez Sánchez and Farias, 2002).

<Fig. 8. >

SC2 core curated at the Geology Department (Oviedo University) was resampled for new ^{14}C and OSL analyses (Fig. 8 and Table 2). A piece of wood at 6m-depth gave an age of 9,109±74 cal. yr BP, and predates the onset of peat deposition. A bulk sediment sample from the top of the lacustrine sequence (10 m deep) gave an age of 14,734±326 cal. yr BP for the end of the lacustrine deposition. A third sample for OSL was taken from a cemented sand and silt level with dispersed gravels at 42.60-42.80 m deep. Precautions were taken to assure that the inner part of the core had not been exposed to light before analyses. The age of 44,966±3,337 yr (Table 1) is coherent with previous ^{14}C dates of the Comeya sequence. Both ^{14}C and OSL dates indicate that the Comeya hollow was already fed by melt waters from the Enol glacier ca 45 ka ago. Although no evidence of till deposits have been found at altitudes lower than 1030 m in this sector, it seems likely that over-deepening processes contributed to the origin of the Comeya hollow basin. Therefore, the age of Comeya infill must be synchronous with the local GM or even younger.

<**Table 2**>

To sum up, the glacial record of Picos de Europa area demonstrates that: i) the MIE reached 1030 m a.s.l., as deduced from the end moraines, and the local GM in Picos de Europa would have occurred at least ca 45 ka BP; ii) if glaciers would have over-deepened the Comeya hollow in a previous cycle, the local MIE could be even older; and iii) at about ca 38 cal. ka BP glaciers would have been retreated up to 1030 m, and a proglacial lake appeared in Enol.

3.3. Other glacial evidence from NW Iberia

The Ancares Range (Cuiña Peak, 1998 m a.s.l.) located to the North of the Trevinca Massif and West of the Cantabrian Mountains also had extensive glaciations. Geomorphological evidence support a local MIE phase with three glacier tongues descending to 850 m a.s.l., along the Porcarizas, Burbia, and Ancares valleys (Pérez-Alberti et al., 1992; Kossel, 1996). This glacier system would have been 13 km-long and 260 to 280 m-thick, with ELA´s ranging between 1400 and 1500 m altitude. South of Ancares, Courel Mountains had a similar glacier dynamics (Rodríguez-Guitián et al., 1995) with local MIE followed by two subsequent phases of glacier front retreat and stabilization. Radiocarbon dates obtained from peat bog in a glacial cirque (17.4 ka BP) imply almost a total glacial recession by the time of the global LGM (Muñoz-Sobrino et al., 2001). West of the Trevinca Massif, surface exposure ages obtained on glacial polished surfaces and moraine boulders in the Queixa-Invernadoiro and Gerêz-Xurés mountain ranges, have provided considerable older ages (Vidal-Romaní et al., 1999). The local MIE moraine in the Queixa-Invernadoiro Range gave ages of 126.1±13.2 ka ^{21}Ne and two drumlins located upwards of 21.6±16.9 and 15.4±6.9 ka ^{21}Ne. In the Gerêz-Xurés Range, two glacial surfaces gave ages of 238.3±17.2 and 130.7±16.8 ka ^{21}Ne for the local MIE and subsequent deglaciation, respectively.

These exposure ages would imply local MIE phases correlating with marine isotope stages MIS 6 and MIS8, respectively.

In the western sector of the Cantabrian Mountains (Somiedo - Babia area), glacial deposits and landforms have been described (Muñoz Jiménez, 1980; Menéndez-Duarte and Marquínez, 1996; Alonso and Suárez-Rodríguez, 2004), but without a chronological model. Jalut et al. (2004, 2010) described the glacial features of the Sil Valley in the southern slope of the Cantabrian Mountains, and provided ^{14}C AMS dates from the base of two glaciolacustrine deposits (Villaseca and La Mata, at 1317 and 1500 m a.s.l., respectively). The results placed the local MIE at 890 m a.s.l. and the timing of local GM at least prior to 41,150 cal yr BP (Jalut et al., 2010). In the eastern end of the Cantabrian Mountains, recent ^{14}C AMS dates from lacustrine records have established minimum ages for local MIE: In the Trueba Valley between 29,149 – 28,572 cal yr BP when the glacial front reached 750 m a.s.l. (Serrano et al., 2011), and in the Vega Naranco Valley, between 28,630-28,330 cal yr BP for a post-MIE glacial phase (Pellitero-Ondicol, 2011).

5. Discussion

5.1. Northern Iberian and global glacial chronologies

Chronological data reported in this paper (26 dates from the Pyrenees and 12 from NW Iberia) and the SPECMAP-1 oxygen-isotopic curve for the last 300 ka (Martinson et al., 1987) as a reference for the succession of glacial and interglacial isotopic stages are shown in Fig.9.

<Fig. 9. >

Northern Iberian glacial ages appear clustered in two groups: i) ca. 21.3 to 97 ka corresponding to the last glacial cycle (isotope stages 2 to -5d), and ii) ca. 122.1 to 263 ka corresponding to previous glacial cycles (from Eemian to isotope stages 6 to 8). The first group dates (last glacial cycle), are from end moraines (synchronous to local GM, or indicating stabilization episodes after local GM retreat), peat bog and lacustrine deposits filling up previously glaciated hollows, and loess (indicators of minimum ages for the local GM). The second group (glacial cycles previous to the last one) includes OSL dates from fluvioglacial deposits or fluvial terraces, and cosmogenic surface dating in moraine and erratic boulders or glacial polished surfaces.

The Pyrenean data indicate that the local GM (i.e. the Würmian MIE) occurred earlier (MIS 4-MIS 5) in that region than the global LGM (MIS 2), though the ice extent was different in the Central than in the Eastern Pyrenees. Glacial advances during MIS 4 and MIS 2 occurred in both regions, but in the Central Pyrenees, MIS 4 and MIS 2 ice tongues are clearly separated in space, and MIS2 re-advance was restricted to headwater positions (Gállego Valley) while in the Eastern Pyrenees, the ice extent was almost similar during the MIS 4 and MIS 2.

The compiled geomorphological and geochronological data for the last glacial cycle in NW Iberia also suggest that the local GM took place prior to the global LGM. Thus, in the Trevinca Massif, ^{14}C AMS chronologies would imply a local GM older than 25.6 ka BP, (Sanabria core SAN04-3A) while OSL dating of fluvioglacial deposits support a local GM older than 33±3 ka. In Redes Natural Park, local GM was coeval or younger than ca 33.5 ka, and by ca 25 cal. ka BP glaciers would have retreated higher up than 1415 m.a.s.l. Last local GM in Picos de Europa occurred prior or synchronous to ca 45 ka. The NW Iberia data set fits the first scenario (local GM earlier than global LGM of MIS 2) established by Hughes and Woodward (2008).

Both in Pyrenees and NW Iberia, there is ample evidence that the last local MIE was not the most extensive one during the Quaternary. This is the case of the Aragón, Gállego, Cinca, and Ariège valleys in the Pyrenees, and the Gêrez-Xurès, and Queixa-Invernadoiro mountain ranges in NW Iberia. More extensive glaciers than those associated to the last glacial cycle occurred during stages MIS6 and MIS8.

5.2. Asynchronous glacial maxima: dating techniques limitations or climatic variability?

Are the observed differences between local MIE and global LGM related to dating techniques or could they be explained by a different regional response to climate fluctuations? The possibility of dating errors has already been discussed in previous works (Hughes and Woodward, 2008; García-Ruiz et al., 2010). For example, radiocarbon ages could be affected by ageing, because of contamination with recycled old carbon (Bordonau et al., 1993) or the influence of a carbonate-rich bedrock (Pallàs et al., 2006). However, the glacial chronologies from non-carbonate areas (Sanabria) and carbonate – rich areas (Enol) are coherent and some radiocarbon samples are terrestrial organic remains, minimizing the possibility of errors based on ^{14}C dating procedures. In addition, the choice of surfaces for exposure dates can lead to sampling errors and to ages younger than usual.

The coherence of one set of dates obtained with a technique in a similar morphostratigraphic context has usually been considered an argument to validate the chronological model. For example, in the Pyrenees, the exposure dates from the Noguera Ribagorzana Valley (Pallàs et al., 2006) and the Têt Valley group around the global LGM. Nevertheless, the exposure dates from a small tributary of the Querol Valley, known as Malniu (Pallàs et al., 2010), and the Ariège Valley (Delmas et al., 2011, 20012) confirm the occurrence of a Würmian MIE prior to the global LGM. OSL dates in the Central Pyrenees are also consistently showing older ages. And in addition, the still few OSL data from NW Iberia (Pias Site, Redes Natural Park and Comeya SC2 dating) are also coherent with local GM older than the LGM.

A useful approach to solve this controversy is the integration of data coming from the analyses of lake sediments and from the geomorphological studies and several dating techniques. This methodology is the best approach to infer the history of glacial retreat as includes multiproxy studies and provides the geomorphological framework for accurate chronologies. Tramacastilla Paleolake (García-Ruíz et al., 2003) and El Portalet peat bog (González-Sampériz et al., 2006) in Central Pyrenees, and Enol Lake (Moreno et al., 2010), and Sanabria Lake (Rodríguez-Rodríguez et al., 2011) in NW Iberia are good examples of integration of multiproxy data. In all these sites, glacial landforms and sediments were mapped in detail and successive scenarios in the glacier evolution were reconstructed using lake sediments chronologies. In Enol and Sanabria, the base of the lacustrine sedimentary sequences gave the age of the proglacial lakes formation, providing a minimum age for local GM (Moreno et al., 2010, Rodríguez-Rodríguez et al., 2011). Sedimentary hiatus in lake sequences provide evidence of phases of glacier activity (González-Sampériz et al., 2006).

Most of the ^{14}C AMS and OSL glacial chronologies reported here fit with first scenario of glacial evolution described by Hughes and Woodward (2008). These authors followed earlier hypotheses of Gillespie and Molnar (1995) explaining the asynchroneity of glacier maxima between continental and mountain glaciers around the world as a result of the particular sensitivity of mountain glaciers to regional climate as well as global conditions. Florineth and Schlüchter (2000) had also explained the differences in glacier development between Scandinavia, the Alps and Southern Europe and proposed that during the MIS 4 and 3, the location of the Polar Front around 46º N, favoured meridian circulation and the growth of glaciers in western Scandinavia, Pyrenees, Vosgues and northern Alps. After 30,000 yr BP, climate was increasingly drier, and the glaciers in south Europe grew again though their advance remained restricted to upstream positions. Preuser et al. (2007) and Ivy-Ochs et al. (2008) showed similar glacial evolution in the western Alps. Local differences in the ice extent in the Pyrenees during the last maximum could be explained by regional climate dynamics. Calvet et al. (2011) proposed that the higher activity of the Balearic low atmospheric pressure centre during the MIS 2 could explain the differences in glacier extent between Eastern (larger advances) and central Pyrenees.

Our review of available data favours a regional response to global climate variability as the main forcing to explain the observed glacial asynchronies. However, new dates and detailed geomorphologic surveys are necessary to refine the age models and further define the regional differences.

6. Conclusions

The review of glacial chronologies in northern Spain Mountains evidences the different research history of both regions: in the Pyrenees the glacial chronology is better known and absolute dating techniques have been applied for more than 30 years ago, while in NW Iberia, quantitative data on glacial evolution are still few, and the first chronological data have been published in the last decade.

Most of the Pyrenean ages obtained from moraines, glacio-lacustrine deposits and fluvial terraces point to local GM earlier than global LGM, that is, during MIS 4, between 50 and 70 ka. A second re-advance coinciding with the global LGM has been detected in talus screes, loess deposits and moraines, although it has different extent in the Central and the Eastern Pyrenees. Minor advances after the global LGM are represented by many frontal moraines. The use of cosmogenic isotopes and OSL dating techniques indicates the occurrence of glaciations prior to the last glacial cycle corresponding to MIS 6 (about 170 ka) and even to MIS 8.

In NW Iberia, ^{14}C AMS dating of marginal deposits (synchronous as or younger than local GM), peat bogs and proglacial sediments in glacial lakes established after glacial retreat have constrained a minimum age for local GM. OSL techniques applied to glacial and fluvioglacial deposits provide ages for the local GM. All of the ages point to local GM earlier than global LGM, with minimum ages between ca 26 to 45 ka. Our review integrating the analysis of glaciolacustrine deposits with the geomorphological study of the glaciated areas and absolute age control defines glacial evolution models with local GM prior to LGM in the Pyrenees and NW Iberia.

The sensitivity of mountain glaciers to local and global climatic changes could explain satisfactorily the regional differences observed, as it has been postulated in other mountainous areas of the Mediterranean Region (Hughes et al., 2010). However, the possibility of a bias introduced by the different dating methodologies cannot be totally discarded. New data are necessary to refine the chronological models and to identify the geographic variability in glacial evolution. The implementation of multidisciplinary strategies involving geomorphological, geochronological, limnogeological and other palaeoenvironmental techniques are necessary to improve our knowledge of glacial evolution in the Iberian Peninsula.

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Table 1

OSL results for the new dataset recently taken in the Redes Natural Park and the Picos de Europa National Park (laboratory at *Universidad Autónoma de Madrid*, Spain). The location of the samples is shown in figures 5, 6 and 8.

Sample label (Lab. Reference)	Sample data	Equivalent dose (Gy)	Annual Dose (mGy/año)	Age (yr)
MAD-5598SDA	Silt-sand sediments of a till outcrop (Brañag-1)	101.86	4.25	23,967±1,841
MAD-5594SDA	Silt sediments from a complex landslide outcrop (Tarna -1)	70.56	3.07	22,983±2,321
MAD-5560SDA	Sand sediments of the SC2 core nucleus (at 42m depth)	67.45	1.50	44,966±3,337

Table 2

New ^{14}C AMS ages obtained for the samples recently taken in the Comeya SC2 core. The calibrated ages have been calculated using the CalPal Software (Weninger et al., 2007) and the calibration dataset of Weninger and Jöris (2008). The location of the samples is shown in Fig. 8.

Sample label	Type of sample	Conventional ^{14}C age (yr BP)	Calibrated results (1σ 68.3% prob.) (yr cal BP)	Calendaric age (yr cal BP)	$^{13}C/^{12}C$ ratio ($^{0}/_{00}$)
CNA429	Wood trunk	8,140 ± 50	9,034-9,183	9109±74	-30.78±0.23
CNA428	Bulk sediment	12,460 ± 70	14,407 -15,060	14734±326	-25.12±0.83

Fig. 1. Setting of the study areas, (a) Pyrenees and (b) NW Iberia, with the location of the main sites described in this work.

Fig. 2. Glacial record in Aragón River Valley (Pyrenees); a) field picture showing the main moraines and the available OSL data; b) interpretative sketch showing till evidence; c) geomorphological map of the Villanúa site (based on García-Ruiz et al., 2011).

Fig. 3. Geomorphological map, stratigraphic logs of the main sedimentary sequences and OSL age data available for the Pias site (Trevinca Massif, NW Iberia) (based on Pérez-Alberti et al., 2011).

Fig. 4. Geomorphological map for the Sanabria Lake site (Trevinca Massif, NW Iberia) showing the main glacial features and a field picture of the ice-dammed deposit of San Martin de Castañeda. The location of the cores referred in the text is shown (based on Rodríguez-Rodríguez et al., 2011).

Fig. 5. Glacial record in Vega de Brañagallones site (Monasterio Valley, Redes Natural Park): a) field picture showing the ice-dammed deposit now covered by alluvial deposits, and the main moraines; b) geomorphological map (after Jiménez-Sánchez and Farias, 2002). The OSL result, which is reported in this work for the first time (Table 1), and the location of the core S1 are also shown.

Fig. 6. Glacial record in Tarna Valley site (NW Iberia); a) Outcrop picture of the complex landslide recently sampled for OSL dating and result obtained (Table 1); b) glacial valley U-shaped located next to Tarna Pass, with the location of the peat bog previously dated with ^{14}C AMS (Jiménez-Sánchez and Farias, 2002); c) Geomorphological map of Tarna Valley, showing the location of the chronological data.

Fig. 7. Glacial evidence from the Western Massif of Picos de Europa National Park: pictures of the Enol Lake (a) and the Comeya hollow sedimentary filling (b) including

the location of the cores drilled in both basins (Farias et al., 1996; Jiménez-Sánchez and Farias, 2002; Moreno et al., 2010). The SC2 core record and the chronological data available are reported in Fig. 8.

Fig. 8. Geomorphological map of Comeya–Enol site with the location of the cores drilled in the Comeya hollow and the Enol Lake (Farias et al., 1996; Jiménez-Sánchez and Farias, 2002; Moreno et al., 2010). SC2 core sedimentary record with the new ages reported in this work for the first time (one OSL and two ^{14}C AMS ages; Tables 1 and 2) and the first data reported in Jiménez-Sánchez and Farias (2002). All the ^{14}C AMS ages have been expressed as uncalibrated results (the calibrated ages can be consulted in the text and in Table 2).

Fig. 9. Correlation of the chronological data reviewed for the Pyrenean and NW Iberia mountains in this paper and the Oxygen-isotopic record for the last 300 ka obtained from the SPECMAP1 Project (Martinson et al., 1987). The time span covered by the latest glacial Marine Isotope Stages (MIS 2- 8) is highlighted in light-blue, and the kind of glacial evidence used to get the ages is shown with different dot patterns.

Figure

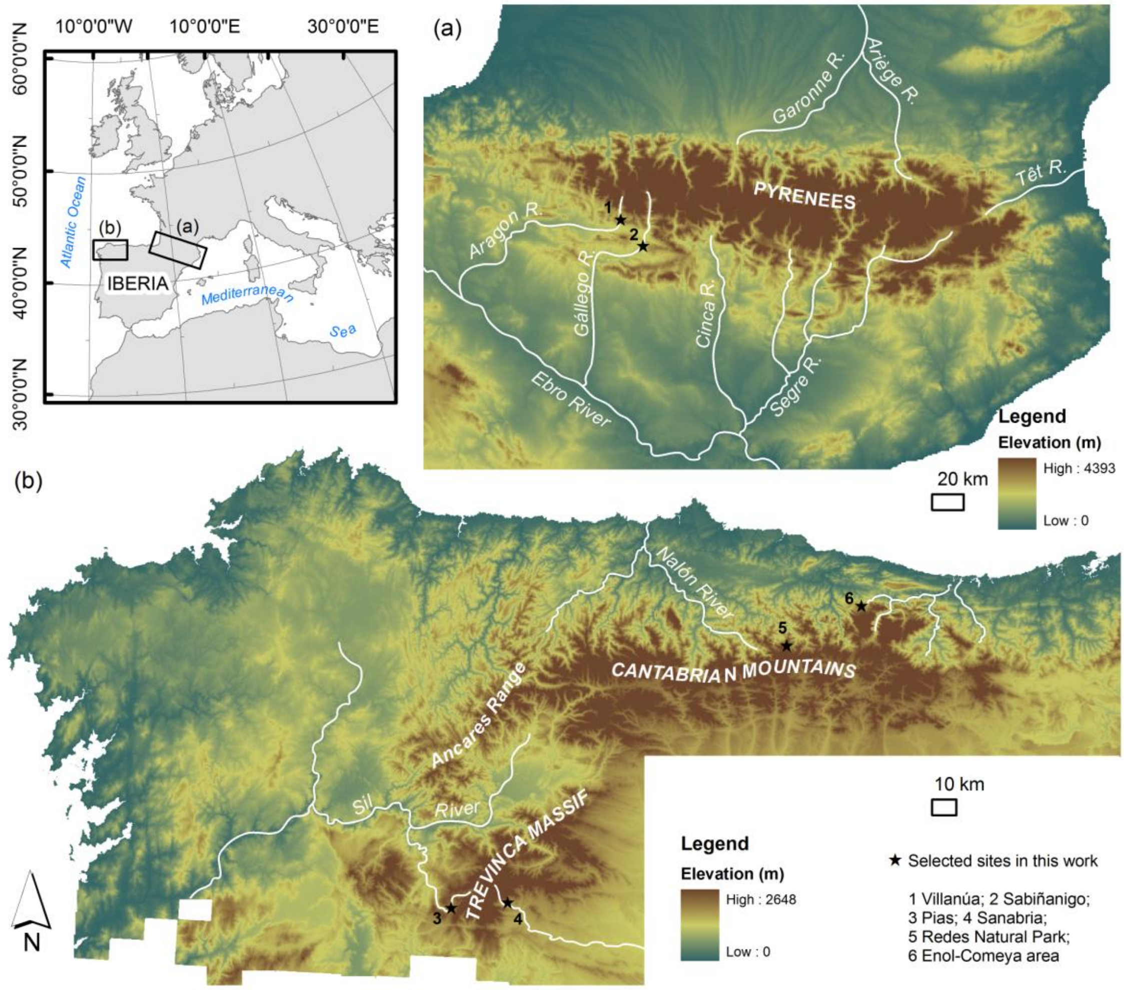

10°0'0"W
10°0'0"E
30°0'0"E
60°0'0"N
50°0'0"N
40°0'0"N
30°0'0"N
Atlantic Ocean
(b)
(a)
IBERIA
Mediterranean
Sea
(a)
Garonne R.
Ariège R.
Têt R.
PYRENEES
Aragon R.
Gállego R.
Cinca R.
Segre R.
Ebro River
1
2
Legend
Elevation (m)
High : 4393
Low : 0
20 km
(b)
Nalón River
CANTABRIAN MOUNTAINS
Ancares Range
Sil
River
TREVINCA MASSIF
5
6
3
4
N
10 km
Legend
Elevation (m)
High : 2648
Low : 0
★ Selected sites in this work
1 Villanúa; 2 Sabiñanigo;
3 Pias; 4 Sanabria;
5 Redes Natural Park;
6 Enol-Comeya area

Fig. 1.

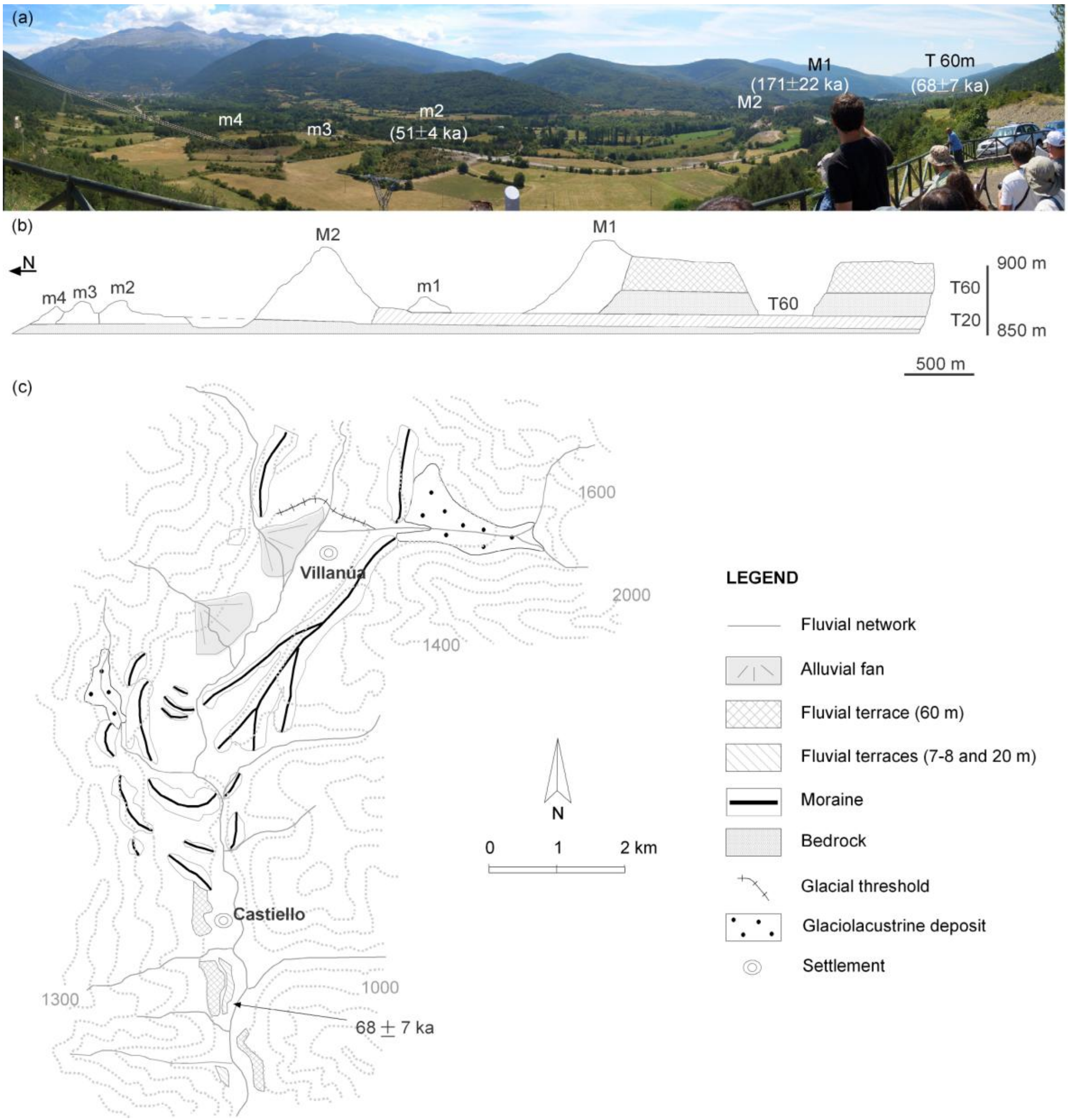
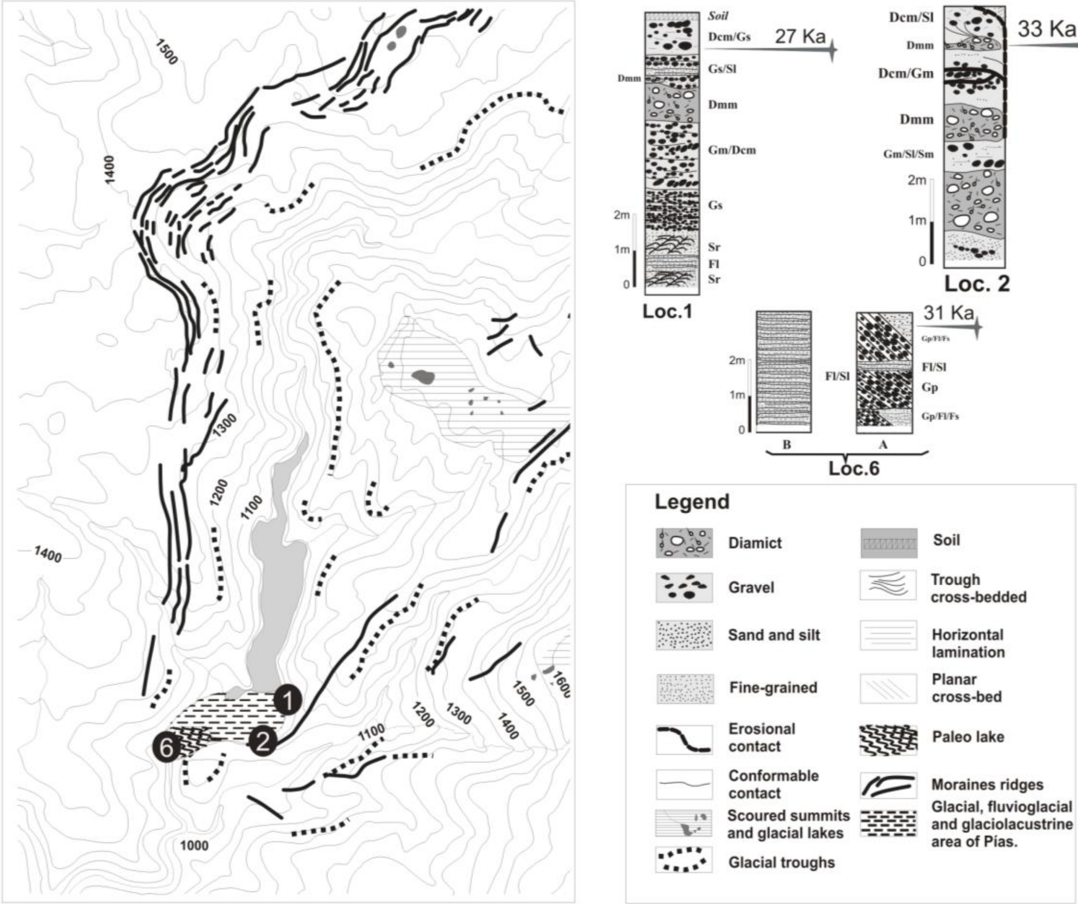
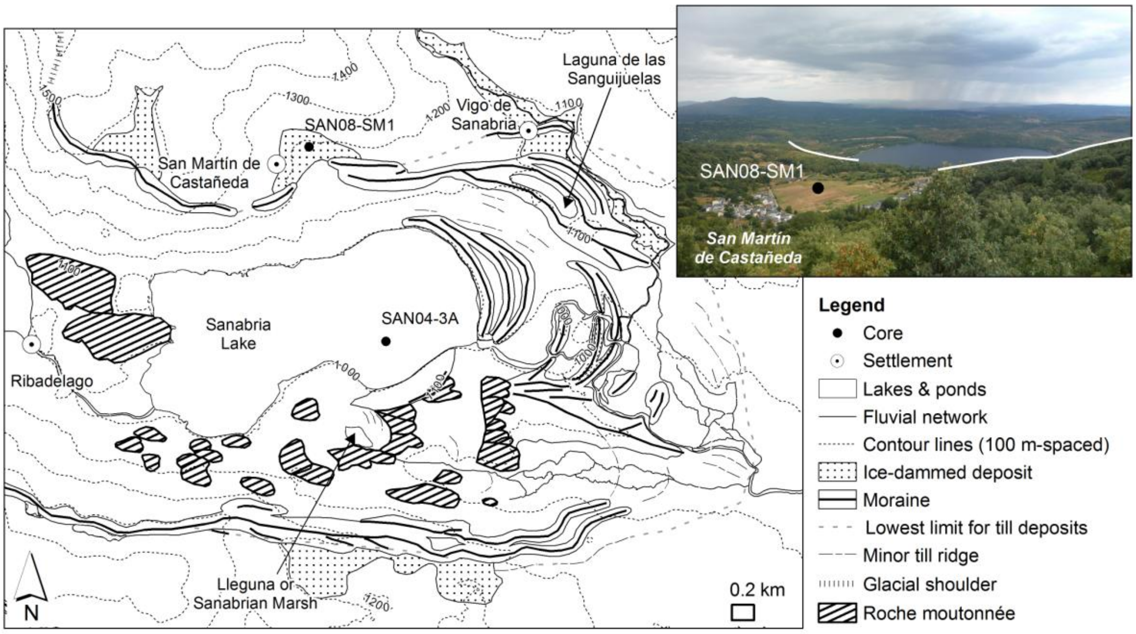

(a)
m4
m3
m2
(51±4 ka)
M2
M1
(171±22 ka)
T 60m
(68±7 ka)
(b)
N
m4
m3
m2
M2
m1
M1
T60
T60
T20
900 m
850 m
500 m
(c)
Villanúa
Castiello
1600
2000
1400
1300
1000
68 ± 7 ka
N
0
1
2 km
LEGEND
Fluvial network
Alluvial fan
Fluvial terrace (60 m)
Fluvial terraces (7-8 and 20 m)
Moraine
Bedrock
Glacial threshold
Glaciolacustrine deposit
Settlement

Fig. 2.



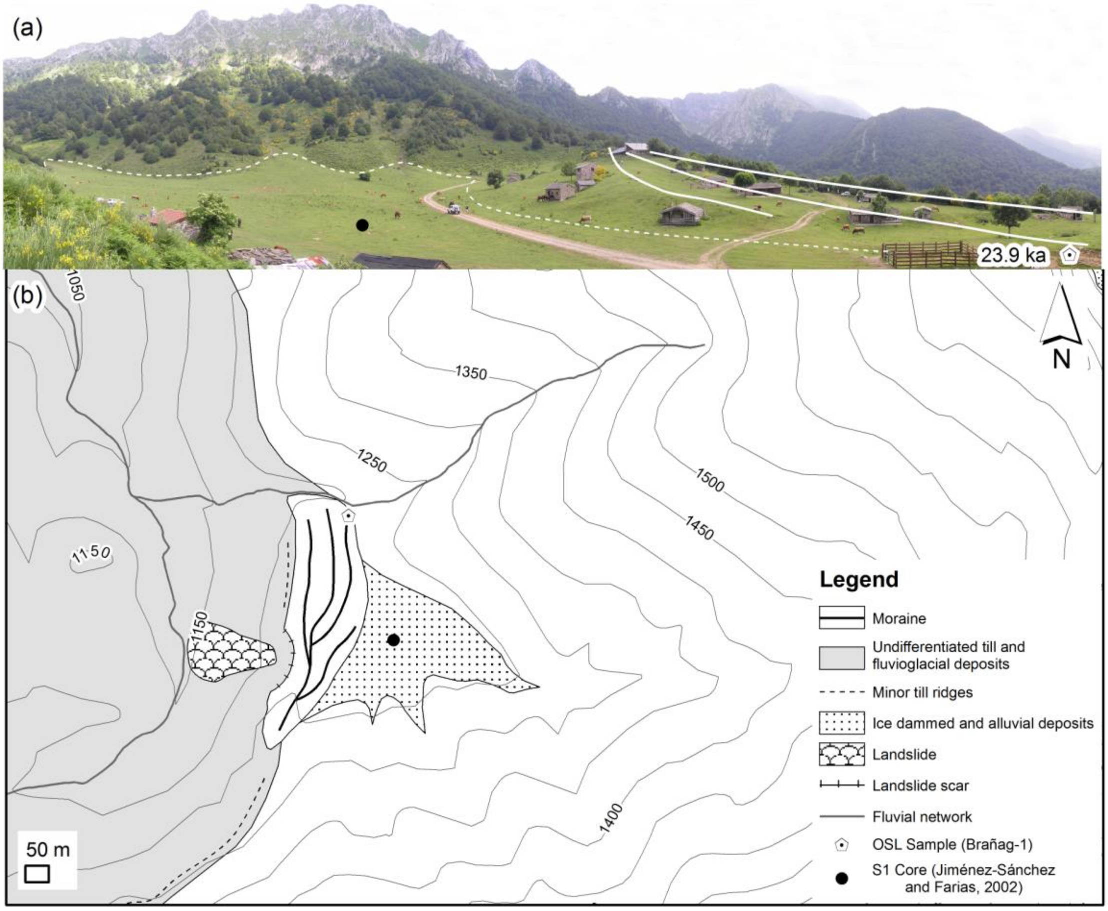
Loc.1
27 Ka
Loc. 2
33 Ka
31 Ka
Loc.6
Legend
Diamict
Gravel
Sand and silt
Fine-grained
Erosional contact
Conformable contact
Scoured summits and glacial lakes
Glacial troughs
Soil
Trough cross-bedded
Horizontal lamination
Planar cross-bed
Paleo lake
Moraines ridges
Glacial, fluvioglacial and glaciolacustrine area of Pías.

Fig. 3.



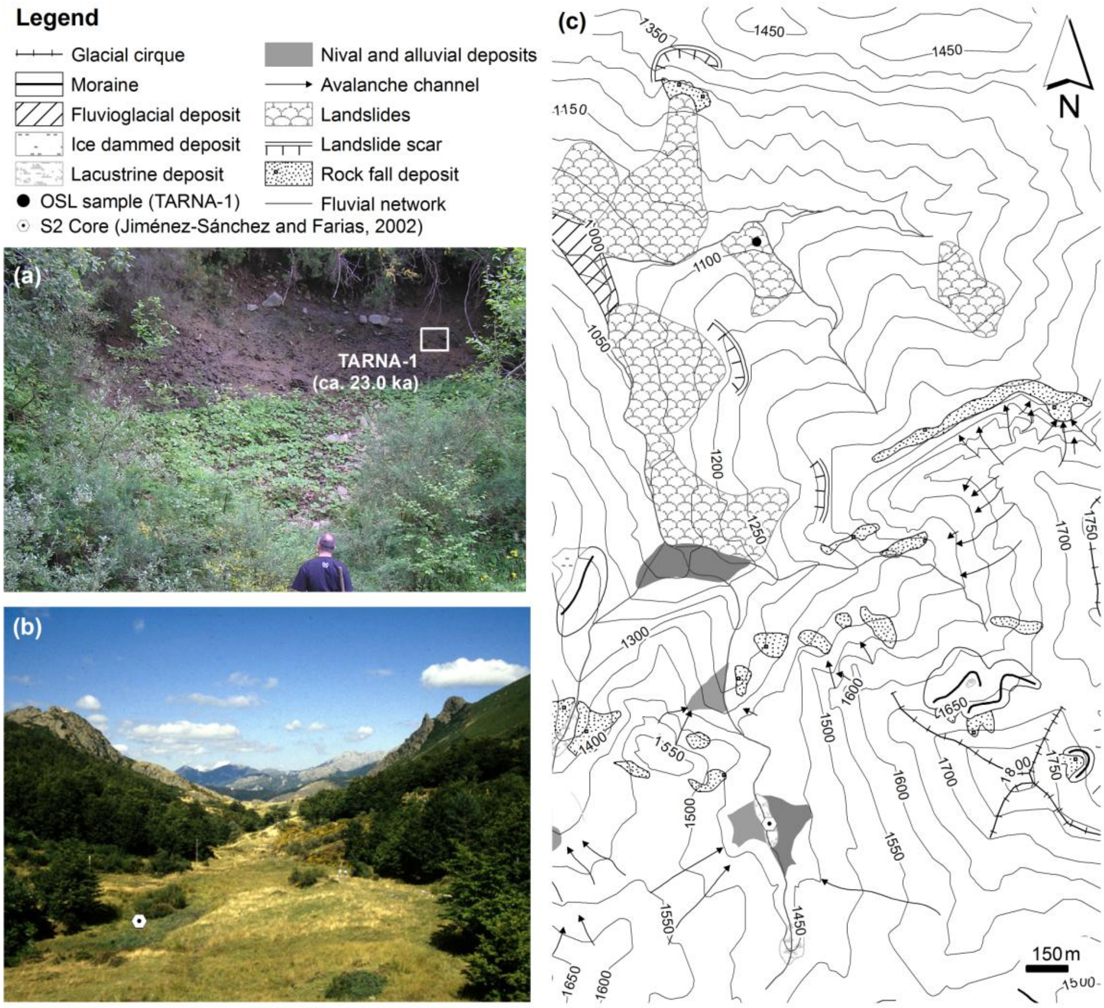
Laguna de las Sanguijuelas
Vigo de Sanabria
SAN08-SM1
San Martín de Castañeda
Sanabria Lake
SAN04-3A
Ribadelago
Lleguna or Sanabrian Marsh
0.2 km
N
Legend
Core
Settlement
Lakes & ponds
Fluvial network
Contour lines (100 m-spaced)
Ice-dammed deposit
Moraine
Lowest limit for till deposits
Minor till ridge
Glacial shoulder
Roche moutonnée

Fig. 4.



(a)
23.9 ka
(b)
1050
1350
1250
1500
1450
1150
1150
1400
N
50 m
Legend
Moraine
Undifferentiated till and fluvioglacial deposits
Minor till ridges
Ice dammed and alluvial deposits
Landslide
Landslide scar
Fluvial network
OSL Sample (Brañag-1)
S1 Core (Jiménez-Sánchez and Farias, 2002)

Fig. 5.



Legend
Glacial cirque
Moraine
Fluvioglacial deposit
Ice dammed deposit
Lacustrine deposit
OSL sample (TARNA-1)
S2 Core (Jiménez-Sánchez and Farias, 2002)
Nival and alluvial deposits
Avalanche channel
Landslides
Landslide scar
Rock fall deposit
Fluvial network
(a)
TARNA-1
(ca. 23.0 ka)
(b)
(c)
N
150 m

Fig. 6.

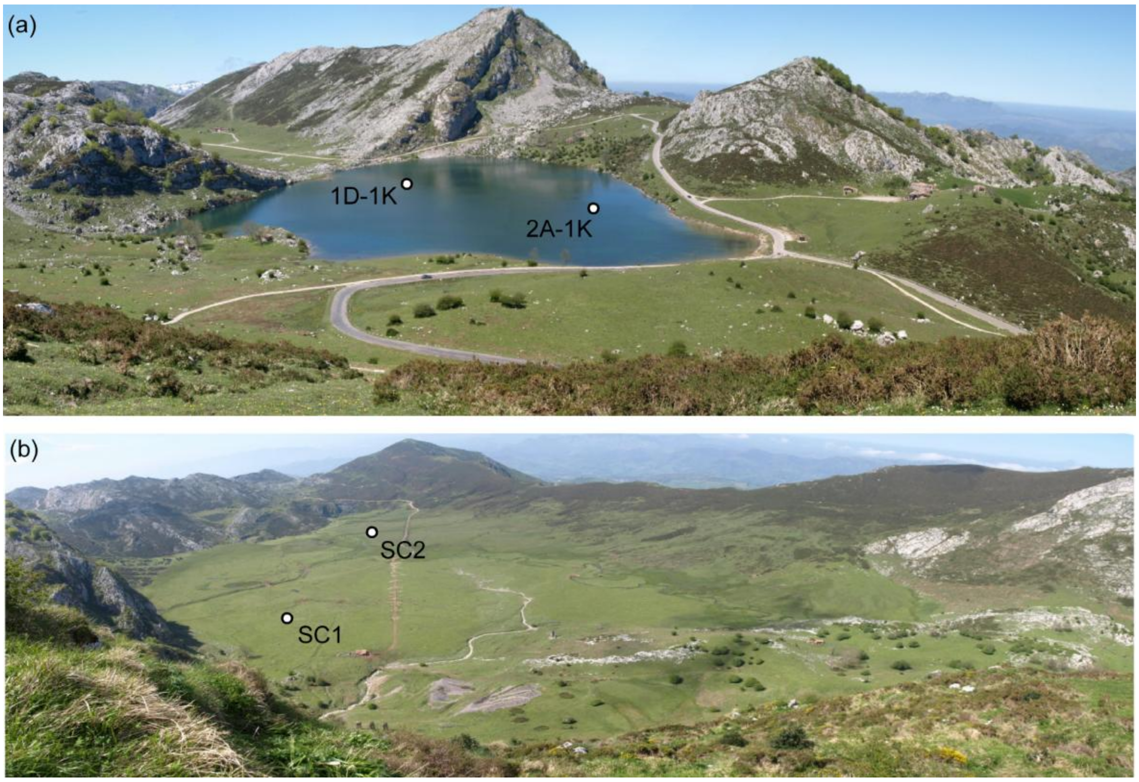
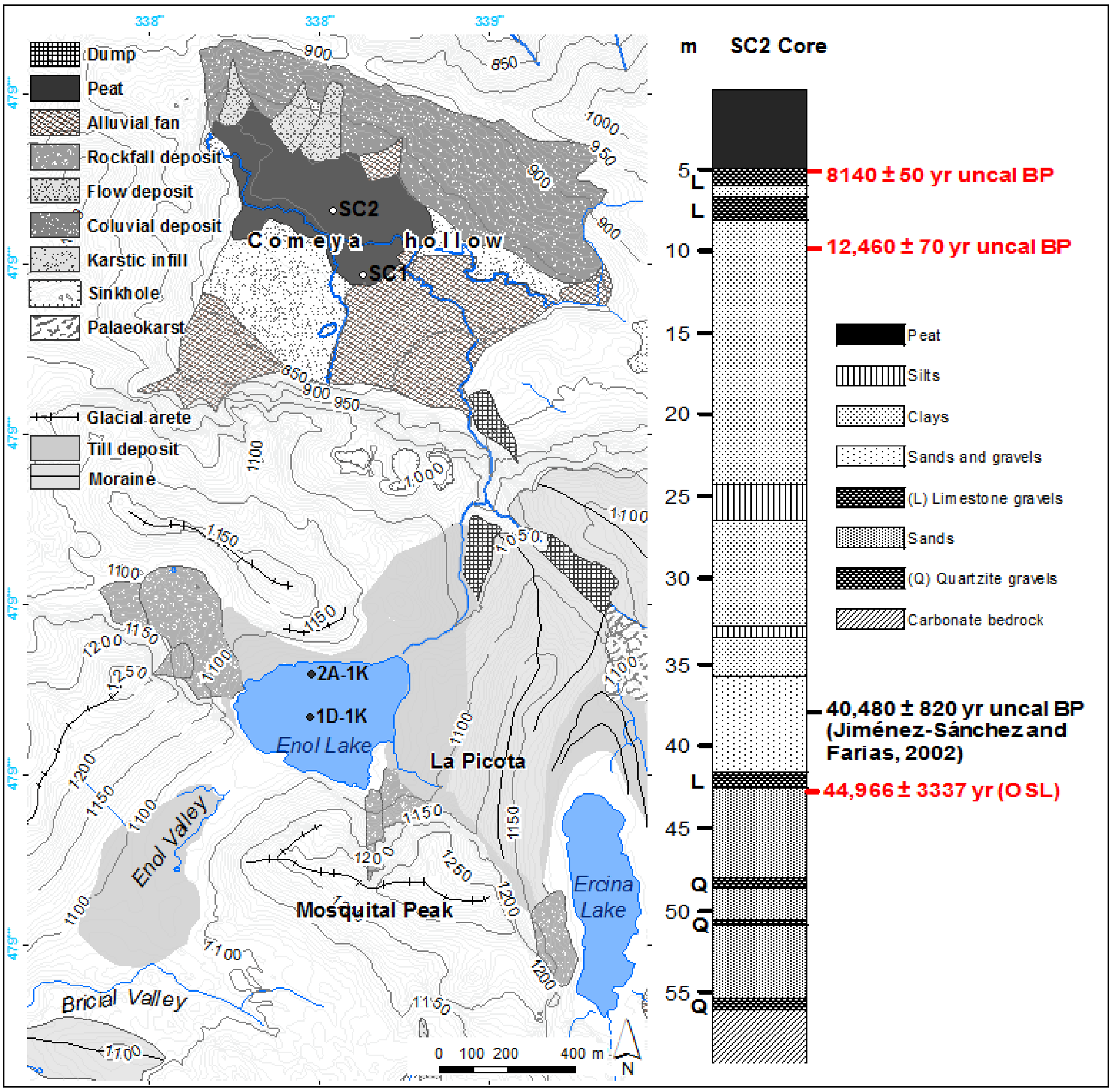


Fig. 7.



Dump
Peat
Alluvial fan
Rockfall deposit
Flow deposit
Coluvial deposit
Karstic infill
Sinkhole
Palaeokarst
Glacial arete
Till deposit
Moraine
SC2
SC1
Comeya hollow
2A-1K
1D-1K
Enol Lake
La Picota
Enol Valley
Mosquital Peak
Ercina Lake
Bricial Valley
0 100 200 400 m
N
m
SC2 Core
8140 ± 50 yr uncal BP
12,460 ± 70 yr uncal BP
40,480 ± 820 yr uncal BP (Jiménez-Sánchez and Farias, 2002)
44,966 ± 3337 yr (OSL)
Peat
Silts
Clays
Sands and gravels
(L) Limestone gravels
Sands
(Q) Quartzite gravels
Carbonate bedrock

Fig. 8.

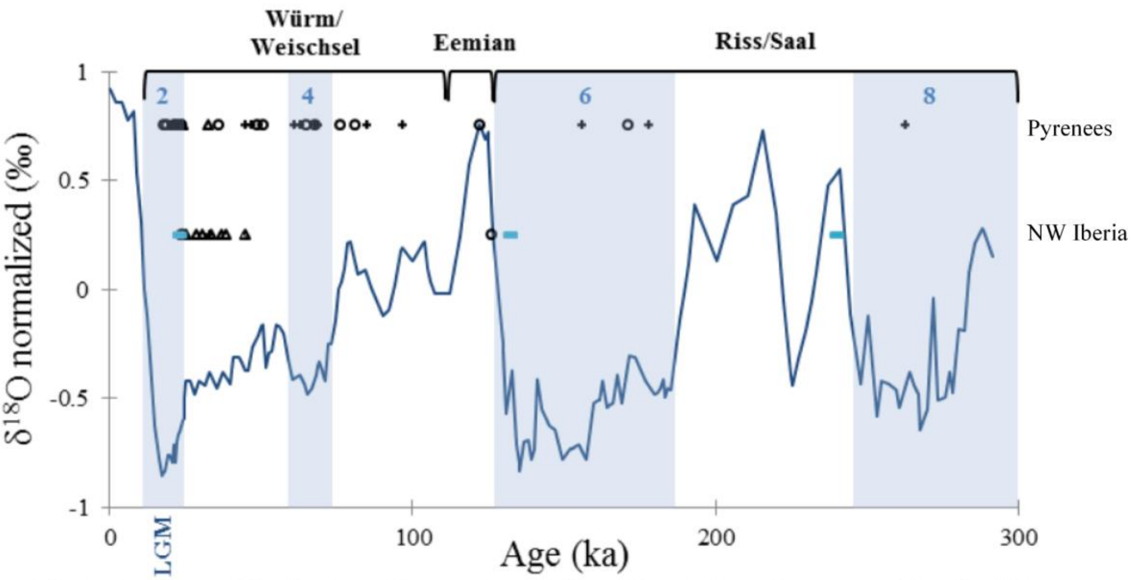

Würm/
Weischsel
Eemian
Riss/Saal
2
4
6
8
Pyrenees
NW Iberia
δ18O normalized (‰)
1
0.5
0
-0.5
-1
0
100
200
300
LGM
Age (ka)
+ Fluvioglacial
○ Moraine & erratics
▵ Lacustrine sediments & peat bogs
- Glacial polish

Fig. 9.

Figure

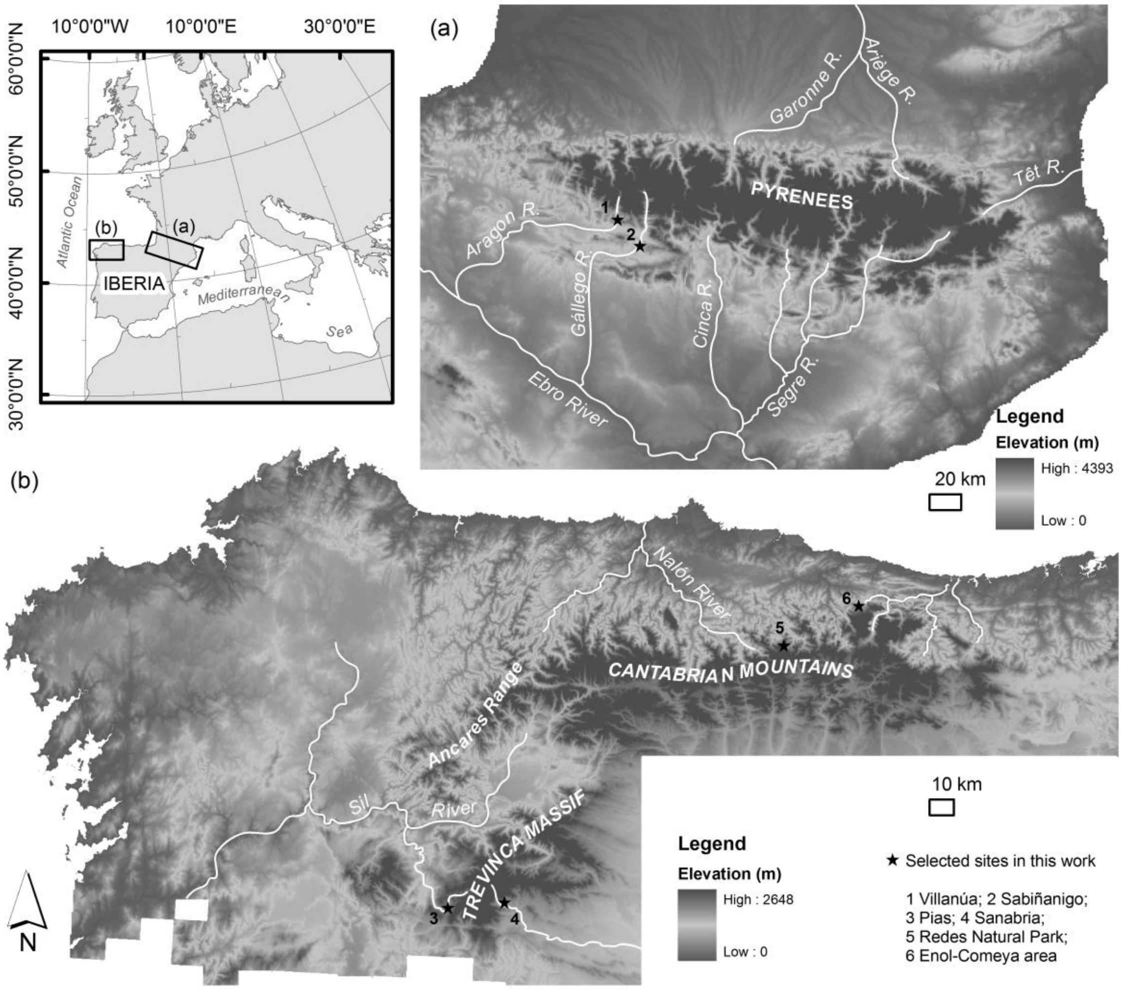
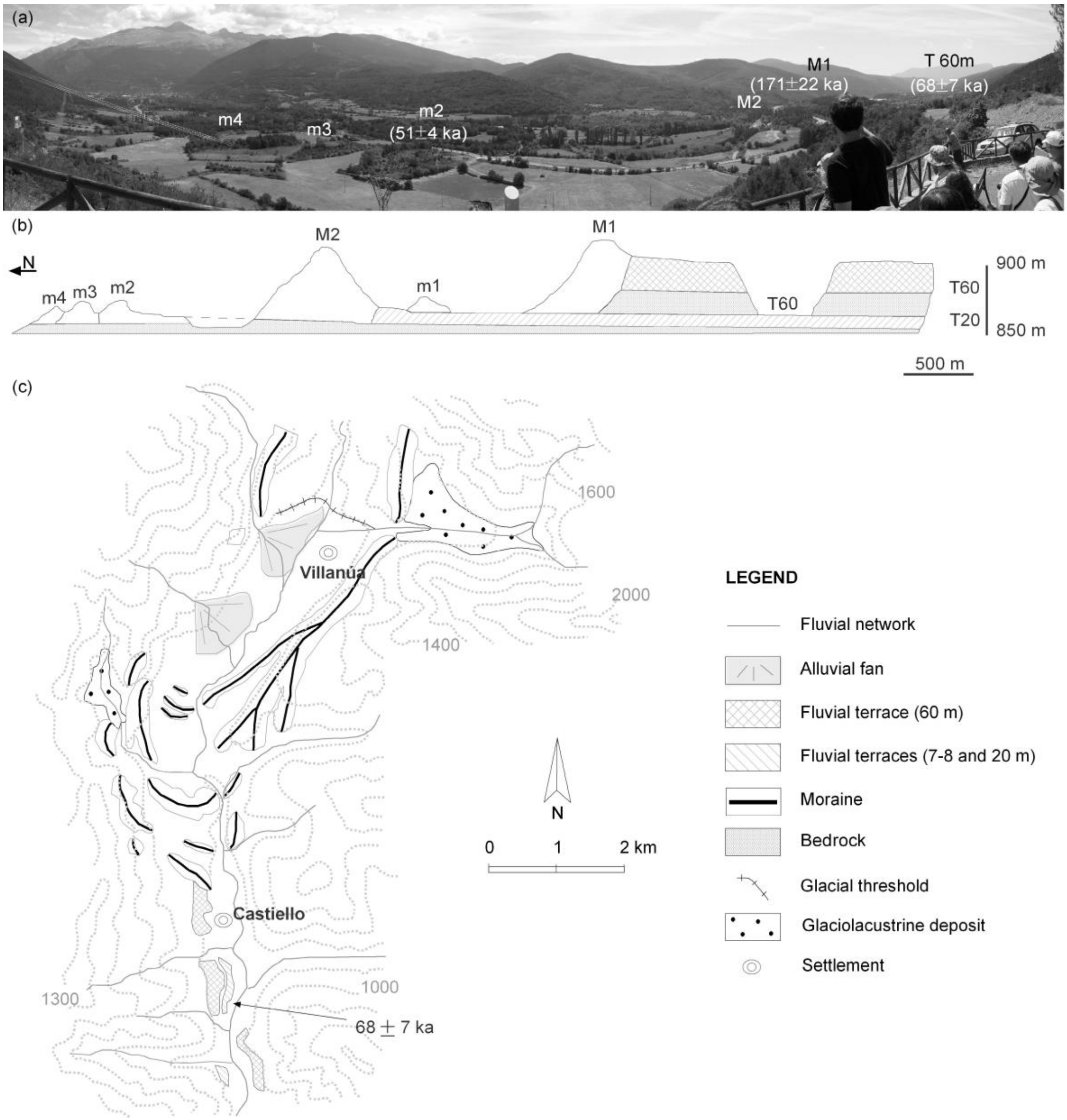
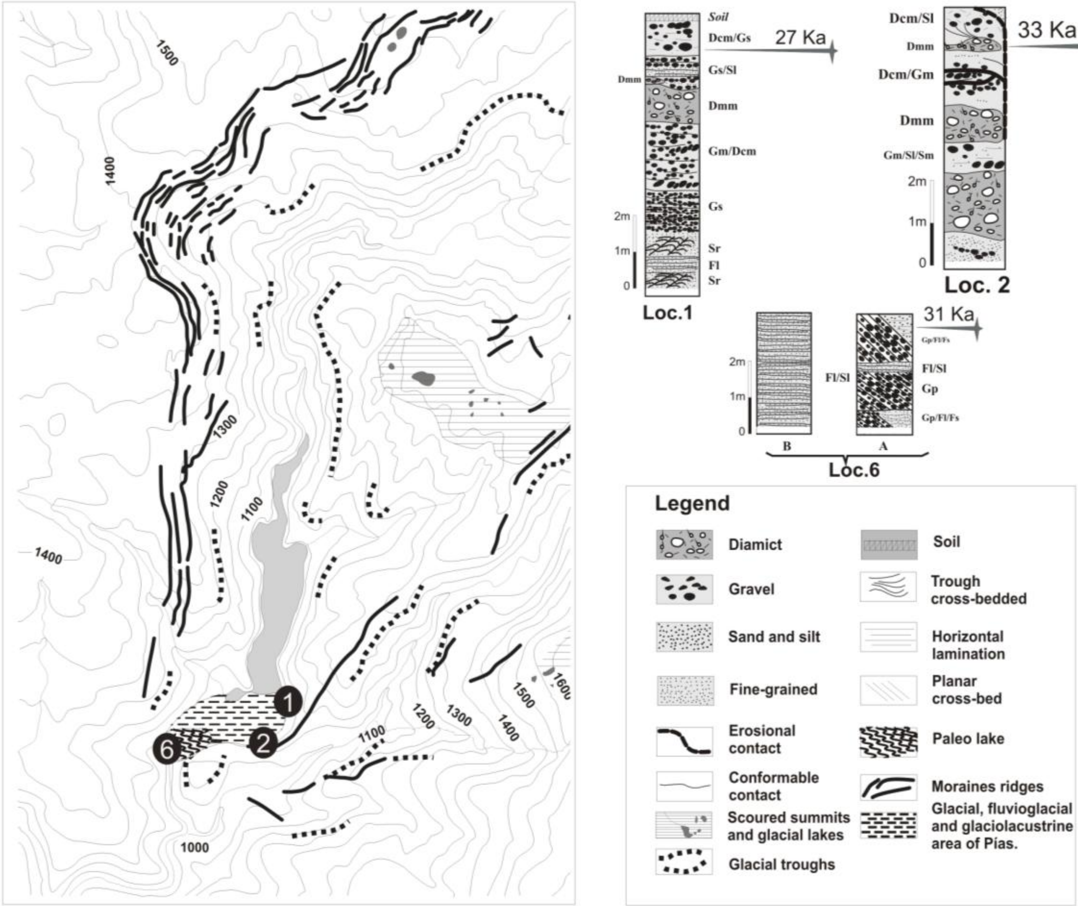

10°0'0"W
10°0'0"E
30°0'0"E
60°0'0"N
50°0'0"N
40°0'0"N
30°0'0"N
Atlantic Ocean
(b)
(a)
IBERIA
Mediterranean
Sea
(a)
Garonne R.
Ariège R.
Têt R.
PYRENEES
Aragon R.
Gállego R.
Cinca R.
Segre R.
Ebro River
1
2
Legend
Elevation (m)
High : 4393
Low : 0
20 km
(b)
Nalón River
CANTABRIAN MOUNTAINS
Ancares Range
Sil
River
TREVINCA MASSIF
5
6
3
4
N
Legend
Elevation (m)
High : 2648
Low : 0
10 km
★ Selected sites in this work
1 Villanúa; 2 Sabiñanigo;
3 Pias; 4 Sanabria;
5 Redes Natural Park;
6 Enol-Comeya area

Fig. 1.



(a)
M1
(171±22 ka)
T 60m
(68±7 ka)
M2
m4
m3
m2
(51±4 ka)
(b)
M2
M1
N
m4
m3
m2
m1
T60
900 m
T60
T20
850 m
500 m
(c)
1600
Villanúa
2000
1400
Castiello
1300
1000
68 ± 7 ka
N
0
1
2 km
LEGEND
Fluvial network
Alluvial fan
Fluvial terrace (60 m)
Fluvial terraces (7-8 and 20 m)
Moraine
Bedrock
Glacial threshold
Glaciolacustrine deposit
Settlement

Fig. 2.


1500
1400
1300
1200
1100
1400
1000
1100
1200
1300
1400
1500
1600
Soil
Dcm/Gs
27 Ka
Gs/Sl
Dmm
Dmm
Gm/Dcm
Gs
Sr
Fl
Sr
2m
1m
0
Loc.1
Dcm/Sl
33 Ka
Dmm
Dcm/Gm
Dmm
Gm/Sl/Sm
2m
1m
0
Loc. 2
31 Ka
Fl/Sl
Fl/Sl
Gp
Gp/Fl/Fs
2m
1m
0
B
A
Loc.6
Legend
Diamict
Gravel
Sand and silt
Fine-grained
Erosional contact
Conformable contact
Scoured summits and glacial lakes
Glacial troughs
Soil
Trough cross-bedded
Horizontal lamination
Planar cross-bed
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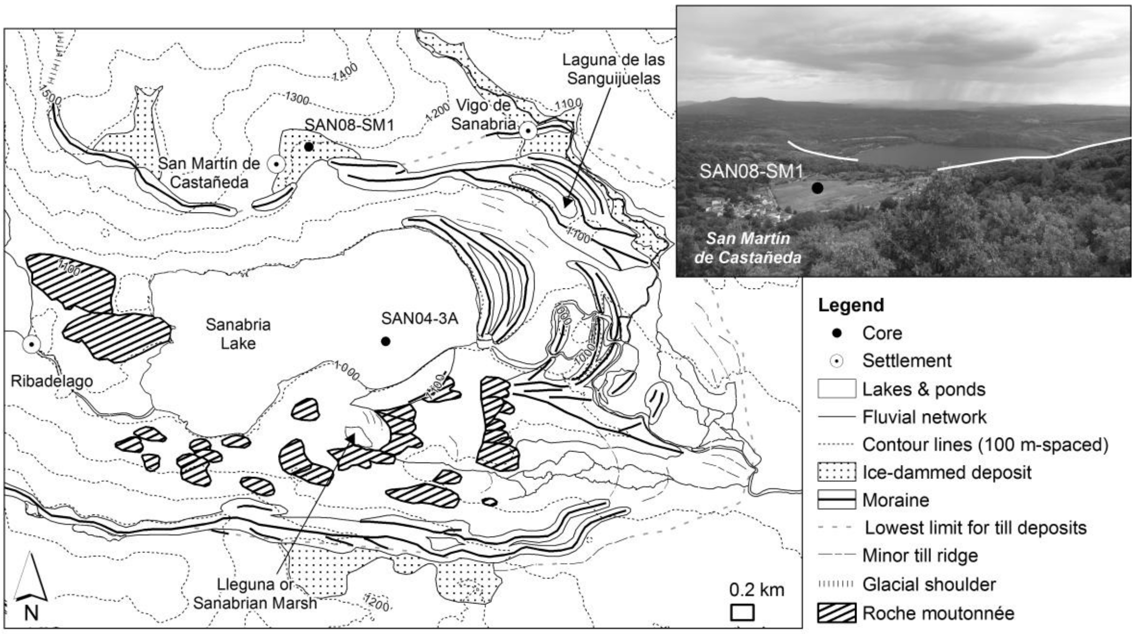
Moraines ridges
Glacial, fluvioglacial and glaciolacustrine area of Pías.

Fig. 3.


Laguna de las Sanguijuelas
SAN08-SM1
Vigo de Sanabria
San Martín de Castañeda
Sanabria Lake
SAN04-3A
Ribadelago
Lleguna or Sanabrian Marsh
N
0.2 km
1400
1300
1200
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Ice-dammed deposit
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Lowest limit for till deposits
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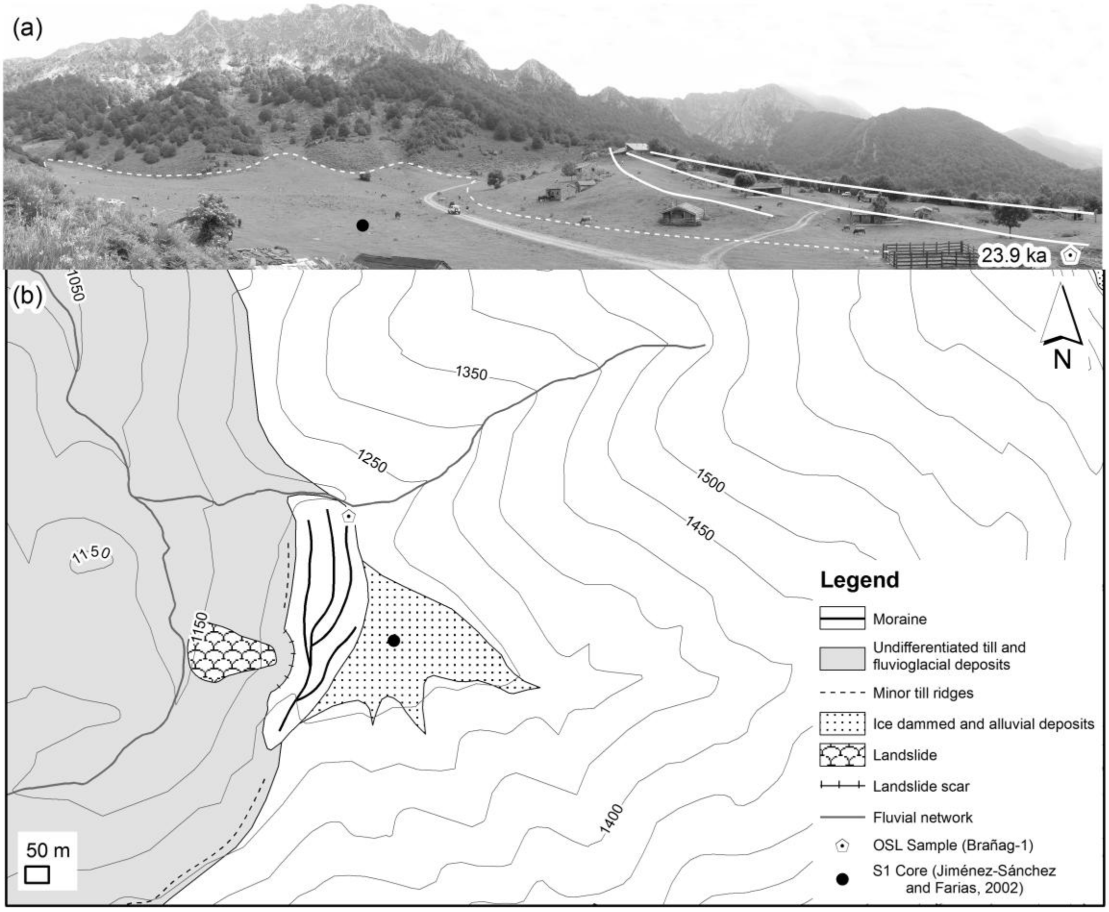
Glacial shoulder
Roche moutonnée

Fig. 4.


(a)
23.9 ka
(b)
1050
1350
1250
1500
1450
1150
1150
1400
N
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Legend
Moraine
Undifferentiated till and fluvioglacial deposits
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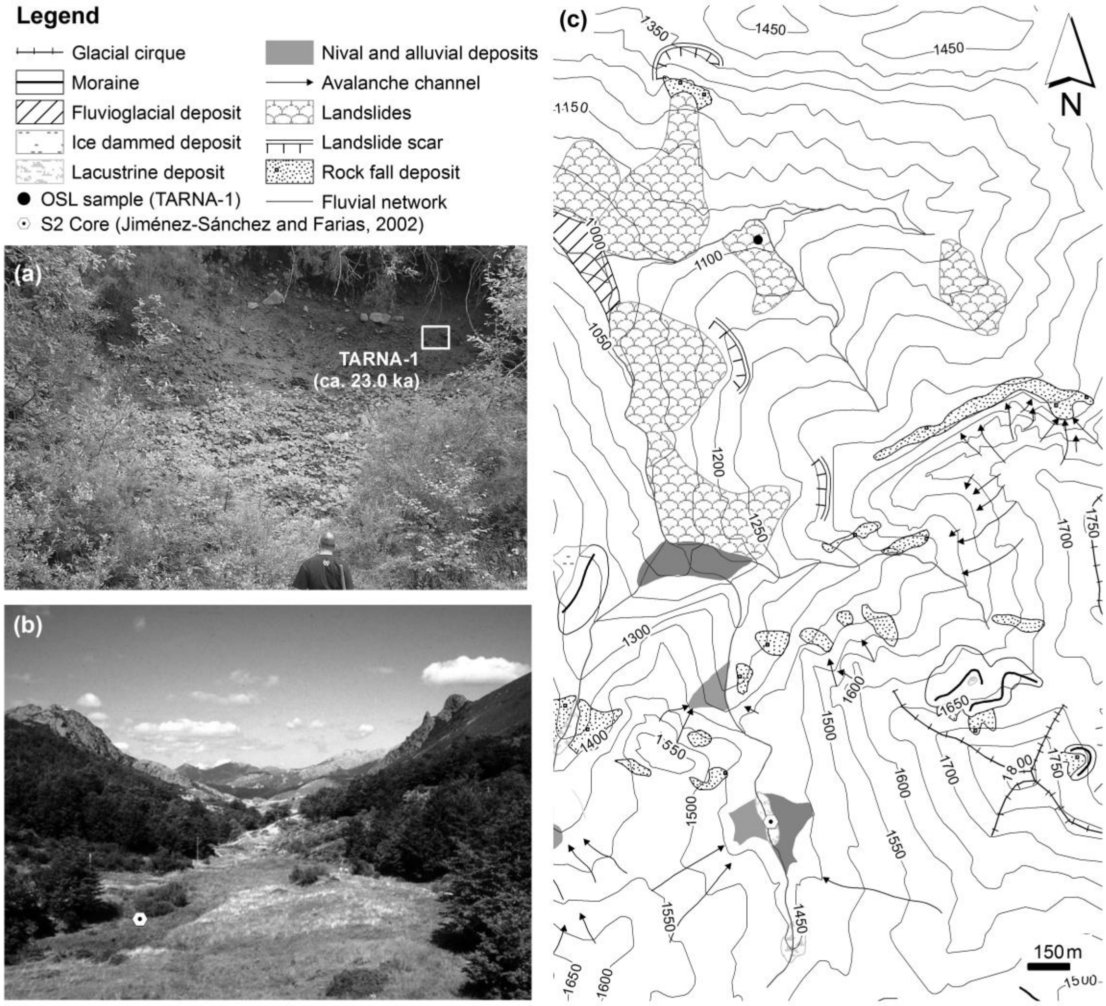

Legend
Glacial cirque
Moraine
Fluvioglacial deposit
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Lacustrine deposit
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(a)
TARNA-1
(ca. 23.0 ka)
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N
150 m

Fig. 6.

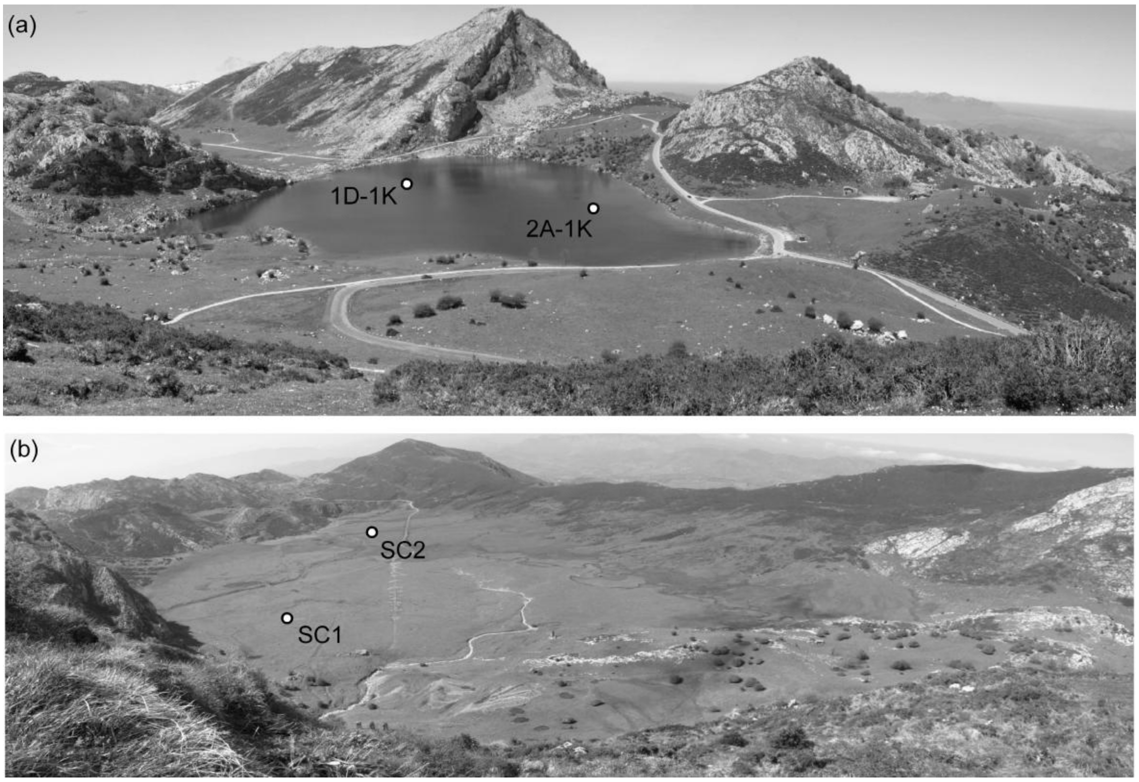
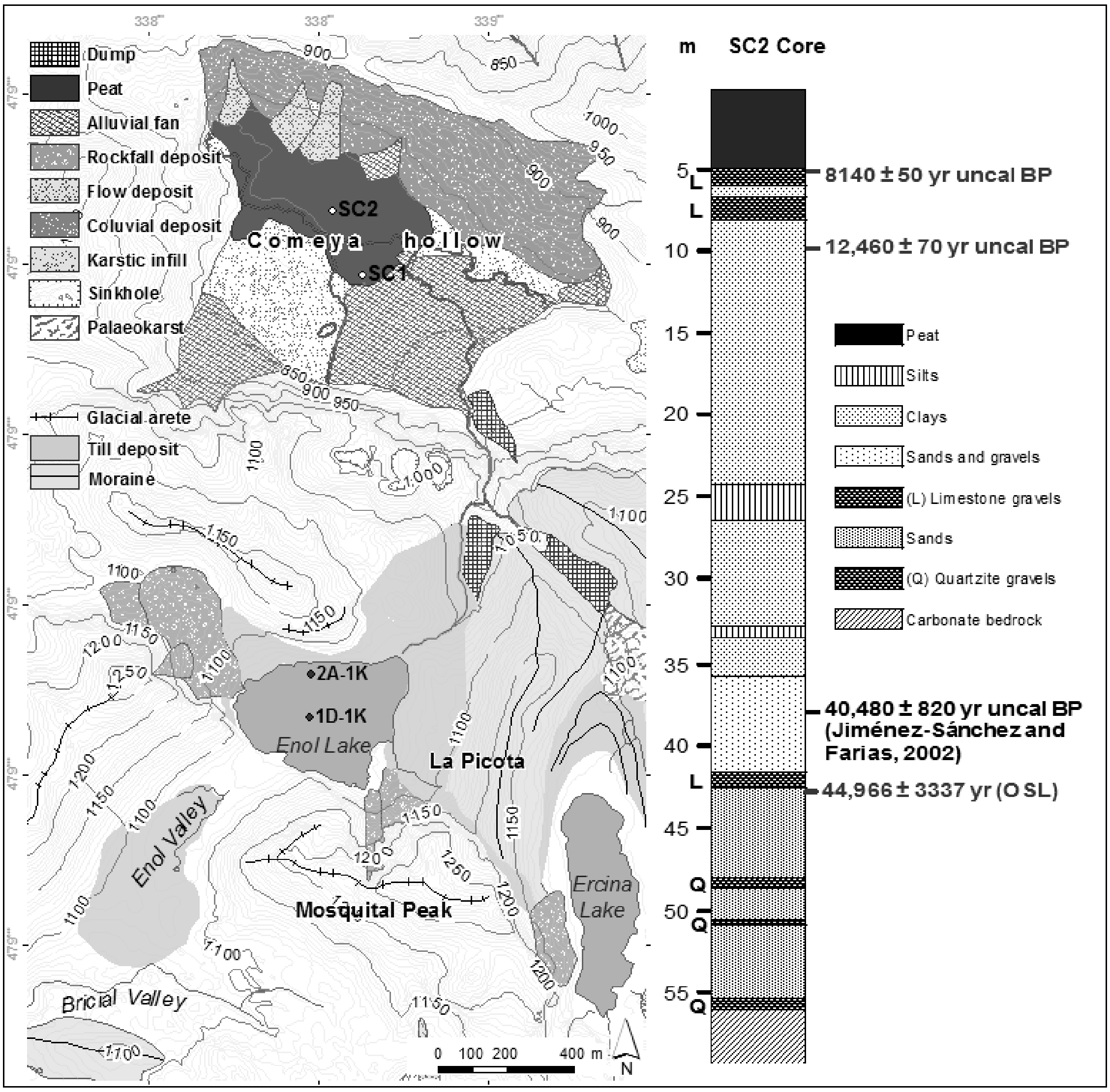
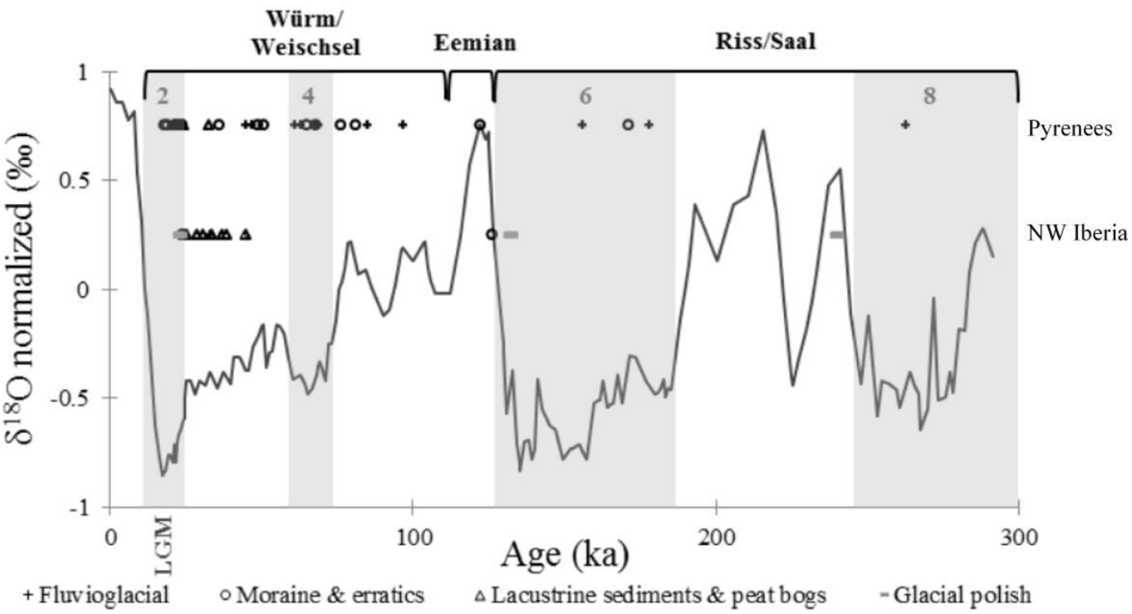


Fig. 7.



Dump
Peat
Alluvial fan
Rockfall deposit
Flow deposit
Coluvial deposit
Karstic infill
Sinkhole
Palaeokarst
Glacial arete
Till deposit
Moraine
SC2
SC1
Comeya hollow
2A-1K
1D-1K
Enol Lake
La Picota
Enol Valley
Mosquital Peak
Ercina Lake
Bricial Valley
0 100 200 400 m
N
m
SC2 Core
8140 ± 50 yr uncal BP
12,460 ± 70 yr uncal BP
40,480 ± 820 yr uncal BP (Jiménez-Sánchez and Farias, 2002)
44,966 ± 3337 yr (OSL)
Peat
Silts
Clays
Sands and gravels
(L) Limestone gravels
Sands
(Q) Quartzite gravels
Carbonate bedrock

Fig. 8.


Würm/
Weischsel
Eemian
Riss/Saal
2
4
6
8
Pyrenees
NW Iberia
δ¹⁸O normalized (‰)
1
0.5
0
-0.5
-1
0
100
200
300
LGM
Age (ka)
+ Fluvioglacial
○ Moraine & erratics
△ Lacustrine sediments & peat bogs
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