ICE PATCH ORIGIN, EVOLUTION AND DYNAMICS IN A TEMPERATE HIGH MOUNTAIN ENVIRONMENT: THE JOU NEGRO, PICOS DE EUROPA (NW SPAIN)

BY

ENRIQUE SERRANO,[1] JUAN J. GONZÁLEZ-TRUEBA,[2] JOSÉ J. SANJOSÉ,[3] and LUIS M. DEL RÍO[4]

[1] Departamento Geografía. Universidad de Valladolid. Valladolid, Spain
[2] Departamento Geografía, Prehistoria y Arqueología. Universidad de el País Vasco. Vitoria. Spain
[3] Departamento Expresión Gráfica. Universidad de Extremadura. Cáceres. Spain
[4] Dapartamento Física aplicada. Universidad de Extremadura. Cáceres. Spain

Serrano, E., González-Trueba, J.J., Sanjosé, J.J. and Del Río, L.M., 2011. Ice patch origin, evolution and dynamics in a temperate high mountain environment: the Jou Negro, Picos de Europa (NW Spain). Geografiska Annaler: Series A, Physical Geography, ••, ••–••. DOI: 10.1111/j.1468-0459.2011.00006.x

ABSTRACT. The Jou Negro ice patch has the greatest surface area of the few that remain in the Cantabrian mountain range. This paper studies its development since the Little Ice Age, its structure and current dynamics. The ice patch lost mass throughout the twentieth century following the general air temperature rise in the region. It was originally a glacier, but from the beginning of the twentieth century it became a glacial ice patch, showing considerable loss of mass after the 1980s. The internal structure with two basal layers of saturated and partially frozen sediments upon which the glacial ice lies, favours contemporary movement. A considerable mass loss in 2009 has been detected associated with small displacements by sliding. The considerable clast cover and topoclimate favours the maintenance of the ice patch, which is in a state of imbalance with current environmental conditions.

Key words: Ice patch, paraglacial environments, Picos de Europa

Introduction

The loss of ice mass is common to most mountains, especially in the temperate high mountain of the Iberian Peninsula. The recession of glaciers results in a transition from ice bodies into ice patches, with melting of non-moving residual ice masses, and glacial environments are frequently replaced by periglacial environments. Marginal ice bodies are today a research field of increasing interest. Several authors have pointed out the importance of knowing the processes of change between the former ice masses and the new non-glaciated morphogenetic systems (Grudd 1989; Harris and Murton 2005; Hughes 2009). The study of marginal glacial and periglacial areas provides knowledge on the last changes from a glacial to a periglacial or deglaciated environment. The processes involved are sensitive indicators of environmental change and climate fluctuations.

Ice patches can be found in the Pyrenees and the Cantabrian Mountains, in the northern Iberian Peninsula. In the Pyrenees there has been constant concern over the differentiation between moving ice masses, unmoving and annual or multi-annual snow accumulations (e.g., Galibert 1965; Martínez de Pisón and Arenillas 1988; Martínez de Pisón *et al*. 1995, 1997; Serrano 1998; González-Trueba *et al*. 2008). On the southern Pyrenees eight glaciers of the eighteen identified in the 1988 by Martínez de Pisón and Arenillas (1988) have become ice patches (González-Trueba *et al*. 2008). In the southern European mountains, such as the Appennines, the Balkans, the southern Alps, Slovenian Alps or Sierra Nevada, always at low altitude, glaciers and ice masses still exist whose study and monitoring reveal the difficulty in determining their activity and dynamics (Grunewald and Scheithauer 2010). Since the 1980s, small European glaciers have lost a moderate surface area but a large volume of ice as temperatures have increased, without clear signs of stabilization (González-Trueba 2007a; Hughes 2009; Grunewald and Scheithauer 2010). Glaciers in transition to ice

DOI: 10.1111/j.1468-0459.2011.00006.x

patches have been studied in the southern Alps, known as "glacionevé" (Assier 1993; Gellatly *et al*. 1994), in the Apenines, where the Calderone glacier, covered by clasts over the last 10 years, has been reclassified as "debris covered glacieret" (Orefice *et al*. 2000; Diolaiuti *et al*. 2006, Diolaiuti and Smiraglia 2010), and in Sierra Nevada, where their disappearance has been witnessed (Gómez Ortiz *et al*. 1996, 2002, 2004). In the eastern Mediterranean mountains recent changes and annual responses in both volume and extension have also been detected (Hughes 2007, 2008, 2009; Hughes and Woodward 2009; Grunewald and Scheithauer 2010). The maintenance of these small ice masses is attributed to topoclimatic factors, mainly shady aspects, snow avalanching and wind drift. Their low altitude and always north-facing orientation means that the little glacier masses survive until a threshold is surpassed (Braithwaite *et al*. 2003; Hughes 2008, 2009), at which point glaciers are transformed into ice patches.

Glaciers, snow patches and ice patches must be distinguished, in order to acquire suitable knowledge on snow and small ice masses, enabling assessment of their different dynamics and evolution, and to use them as geoindicators of change. Definitions such as small glaciers (and the frequent use of the synonym glacierets), snow patches or nevés and ice patches are frequently used but very vague and unclear and in need of precise definition.

Small glaciers are a very common feature in all mountains and high latitudes of the World. They are true glaciers in which movement and internal motion are present and include types such as niche, wall side, cliff or apron glaciers (Gerrard 1990; Benn and Evans 1998). The term *glacieret* was introduced by Lliboutry (1965) to classify small glacial ice bodies and are considered to be very small glaciers (Hoffman *et al*. 2007; NSIDC 2009). Some papers consider them to be accumulations of firn or nevé with little or no movement, originated by snow-drift and avalanches (Benn and Evans 1998; WGMS 2005), or by ice deposition in cold-bottom or karst dolines in high mountains (Barsch 1996; Guyton 2001; Rau *et al*. 2005, WGMS 2005; EUNIS 2009). However, glacierets are small glaciers, the product of larger ancient glaciers, still showing motion or ice deformation, although both very low. They have a glacial origin, glacial ice and are never generated by new snow accumulation.

A snow patch is a relatively small area of snow cover remaining after the main snowmelt period (Kotlyakov and Smolyarova 1990; IPA 2005; NSIDC 2009). It is a snow accumulation by wind drift and fall, and conserved by topoclimatic factors. The snow patches occur throughout the years in the same place, melting in late summer, or may sometimes survive for several years. A snow patch is considered to be present when the snows lasts for at least two consecutive years, but this period is not necessary for the development of nivation processes and landforms (Thorn 1988; Christiansen 1998).

Differences between small glaciers, ice patches and snow patches are defined by their dynamics (motion or not motion), structure (ice, firn, snow, folds) and genesis (glacial or nival). Glacierets are formed by glacier ice deformed by motion with corresponding glacier internal structures and current internal motions. Snow patches may be composed by firn or massive ice (regelation and segregation ice), but do not show deformation by internal motion. The processes are very different. Whereas glacierets are the product of glaciers, generated by glacial processes and originating glacial landforms, snow patches are built up by nival accumulation processes (fall, wind) and the resulting landforms are nival. Literature on non-deformed or deformed ice and firn in snow and ice patches is scarce and to know the origin of ice deformations becomes complicated when the ice mass shows structures derived from uplift or pressures from overlying or underlying material.

Ice patches are ice bodies without movement by flow or internal motion. They are differentiated from snow patches by their ice content and from glaciers by the absence of internal motion. Kuhn (1995) points out that the absence of activity, not size, is the determinant factor in the distinction between glaciers and ice patches. If the ice mass deforms, then it should be considered a glacier. Our classification involves two genetic mechanisms (Fig. 1):

1. Ice patches generated by snow accumulation and firn development. The dominant structure of the ice mass is horizontal or subhorizontal, without postdepositional and postdiagenetic internal deformations. This type of ice patch is indicative of a phase of snow accumulation leading to the formation of a glacier when the ice mass allows for flow and internal motion. Galibert (1960, 1965) points at frozen snow on alpine walls (named "caparace de neige glaceé") as a possible origin of ice patches,

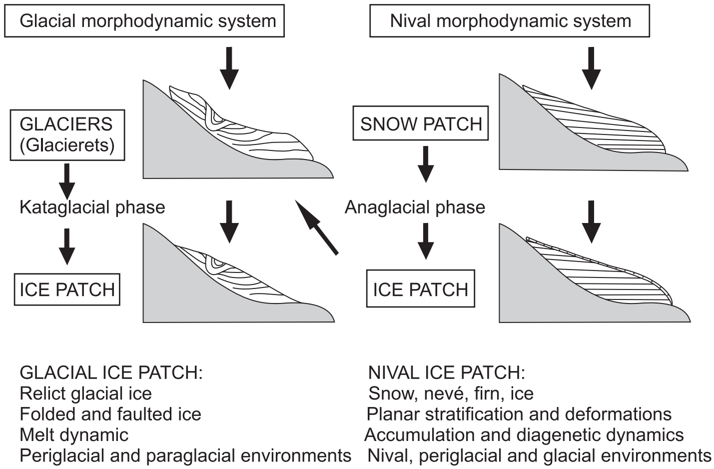

Fig. 1. Types and characteristics of ice patches. See text for further discussion.

although he studies the decline of glaciers and not their genesis. They are generated during anaglacial phases, and very little is known about these processes, probably because their genesis is currently very small worldwide. This type of dynamics is not present in the Iberian Peninsula.

2. Ice patches formed by relict glacier ice or dead ice. These are the remains of old glaciers that have lost mass to the point that their internal motion and flow have ceased. Kuhn (1995) points out that the absence of activity, not size, is the determinant factor in the distinction between glaciers and ice patches. The ice patches maintain glacial internal structures, with folds and fault structures derived from their internal motion during the glacial phase. They are generated during kataglacial phases (Martinez de Pisón, and Arenillas, 1988) and are the last witnesses of relict glacial ice of old glaciers. Their decline or disappearance is a sign of significant hydrological and geomorphological changes.

The two types of ice patches show different attributes, genesis, evolution and processes indicating different environments (nival and anaglacial in the former case, paraglacial and kataglacial in the latter) and dynamics (fully active in the former, relicts in the latter). We propose, therefore, a different designation, depending upon their origin: nival ice patch in the former case and, glacial ice patch in the latter (Fig. 1). The genetic and structural differences differentiate the two types of ice patch. Nival ice patches show an ice core surrounded by firn and nevé, whereas glacial ice patches do not necessarily have firn or nevé. In both cases readjustment movements or basal sliding can be detected. However, these are mass movements, not internal ice motion, derived from imbalances due to melting, mass change, saturation of the basal deposits or the accumulation of snow or ice.

This paper studies the recent evolution and current dynamics of the Jou Negro ice patch, in the Picos de Europa (NW Spain). Focus is given on the morphodynamic surface transformations accompanying melt and annual changes in the ice mass and its surroundings. The ice patches present in the Picos de Europa are the last physical signs of the glaciers that existed until just a few decades ago (González-Trueba 2006, 2007a; González-Trueba *et al*. 2008) and are precise and sensitive indicators of change in the Cantabrian high mountain environment.

Study site

The Jou Negro cirque is located in the Picos de Europa Central Massif (Cantabrian Mountains), in the north of the Iberian Peninsula (43°12′03″N 4°51′12″W, Fig. 2). The Picos de Europa massif is an oceanic mountain range, 20 km from the Atlantic coast, the first orographic barrier for the rain-bearing winds from the Atlantic Ocean. The climate is warm (MAAT, 10°C at 1608 m a.s.l.),

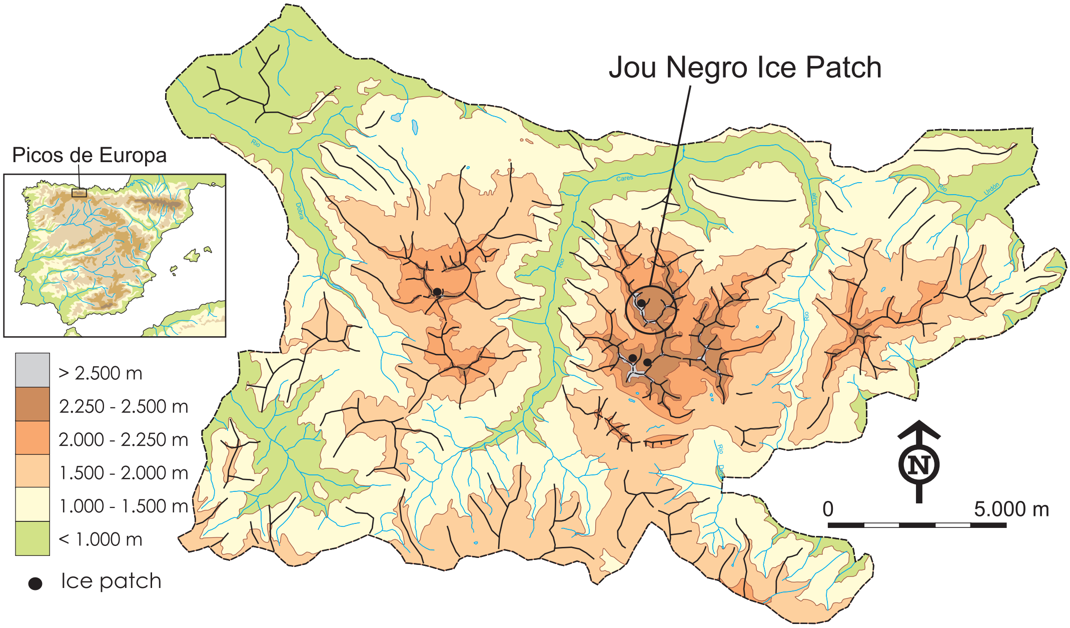

Fig. 2. Location of the Jou Negro ice patch in the Picos de Europa Central Massif. Black points indicate the four ice patches located in the Picos de Europa.

characterized by high precipitation of up to 3000 mm yr^{-1} at the summits. Precipitation falls mainly as snow during the winter, but rainfall and cloudiness are high throughout the year. The bedrock is Carboniferous limestone affected by a succession of northerly dipping thrust faults and the relief is the result of tectonic, glacial, karstic and periglacial Quaternary processes. The Picos de Europa was the only glaciated massif of the Cantabrian Mountains during the *Little Ice Age (LIA)* between the fourteenth and nineteenth centuries (González-Trueba 2006, 2007a, 2007b). At present four ice patches exist in the Picos de Europa (Fig. 2), the only ice mass in the Cantabrian Mountains, all of them small ice masses occupying a total of 61 850 m^2 (González-Trueba 2006, 2007b). The Jou Negro ice patch is the most extensive ice body in the Cantabrian high mountain, with a 0.6 ha surface area (Table 1).

The Jou Negro is a glaciokarstic depression located at 2220 m a.s.l. on the north face of Torre Cerredo (2648 m a.s.l.), the highest peak of the Cantabrian Mountains. The cirque is shaped by nivo-karstic landforms (karst depressions, sinkholes and karren) and subsurface drainage, which explains the good preservation of LIA moraines located at the bottom (Fig. 3). The cirque is surrounded by a 400 m high wall with snow and debris avalanche tracks and the wall is covered on the lower part by debris cones and talus. Erratic blocks and periglacial landforms and processes occupy the nowadays deglaciated areas (Fig. 3).

Methodology

Using complementary geomatic, geophysical and geomorphological techniques, as well as the analysis of climatic and historical data, we have obtained information on the variation and evolution of the ice mass in relation to environmental changes. Historical and recent documentary records (pictures and written documents) of the LIA glacier and the ice patch during the twentieth century were useful sources of information. Twenty three pictures and documents dating as far back as 1893 were analysed to rebuild the Jou Negro ice mass's extension and evolution.

A geomorphological map of the Jou Negro cirque with morphoclimatic information on active landforms and processes on the ice patch and the surrounding environment was produced. The map shows the current dynamics and the spatial relations between the different processes.

Table 1. Surface and thickness changes in the Jou Negro Cirque (1890–2008).

		1890–1900	1980	1995	2008
Type of the ice body		Small Glacier	Ice patch	Ice patch	Ice patch
Surface area	(ha)	5.2	3.5	2.5*	0.6**
Surface loss since the LIA	(ha)	–	1.7	2.7	4.6**
	(%)	0	32.7	52	88.5**
Surface loss compared to previous time period	(ha)	–	1.7	1	1.1**
	(%)	–	32.7	28.5	44**
Annual rate of loss	(m^2 yr^{-1})		212	1000	1460**
	(% yr^{-1})		1.25	10	7.6**
Thickness loss (m)	Low zone	–	?	10	–
	Middle zone	–	15	25 (10)	–
	High zone	–	20–25	45–50	50–55 (2–3)

* Estimation from González-Suárez and Alonso, 1994.
** Estimation of outcropping surface loss, including the recovering by clasts. The ice is preserved under debris.

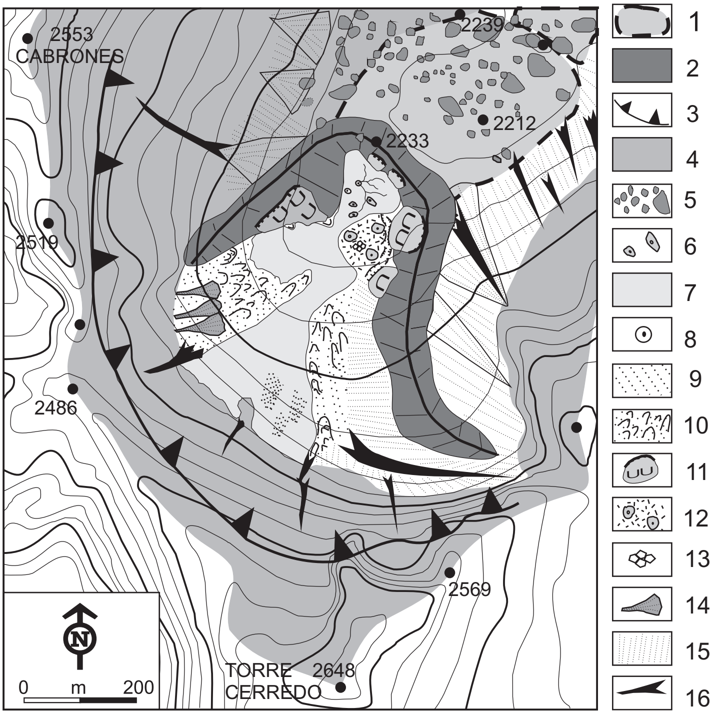

Fig. 3. Geomorphological sketch of the Jou Negro cirque. 1) Glaciokarstic depression. 2) LIA moraine. 3) Glacial cirque. 4) Calcareous wall of the Jou Negro cirque. 5) Erratic blocks. 6) Frost mounds. 7) Ice. 8) Supra-ice sinkholes. 9) Supra-ice debris. 10) Gelifluction lobes. 11) Slope slides. 12) Frost mounds. 13) Patterned ground. 14) Debris cones. 15) Talus cones. 16) Avalanche and debris tracks.

The nearest meteorological stations to the Cantabrian Mountains have fragmentary data and are located far from the study site. There are no climatic data, nor snowfall data, for the Picos de Europa high mountain prior to 2008. Outside the mountain, since 1973, there is a continuous record of meteorological data from meteorological stations in Leon (95 km southwest), Santander (88 km northeast) and Oviedo (86 km northwest). This allows the evaluation of the regional climate setting and to analyse summer temperature series. Small glaciers respond to changes in regional climate, in

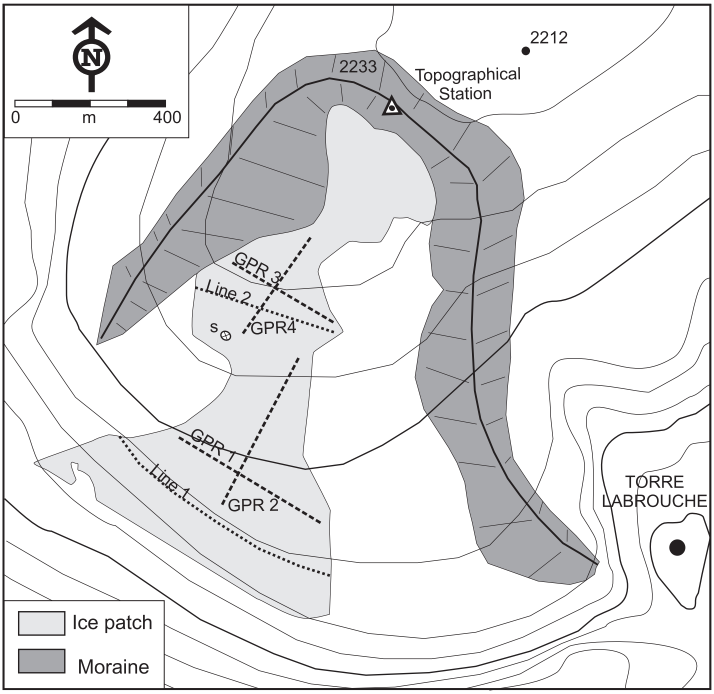

Fig. 4. Location of geodetic control lines (line 1 and 2) and GPR sounding (GPR 1, 2, 3 and 4) in the Jou Negro ice patch. S, mechanical sounding in a moulin.

which summer temperatures and spring snowfall are important factors in predicting mass balance (Ohmura *et al*. 1992; Hoffmann *et al*. 2007; Hughes 2008, 2010). Winter precipitation is not so significant because wind drift and avalanching from walls and channels overcome direct snowfall accumulation. The Leon, Oviedo and Santander climatic record was examined for *mean summer temperatures (MST)* of each year (MST – June/July/August/September) to calculate the deviation to the summer mean temperature of the 1973 to 2009 period. The results were used to evaluate the climate control on melting and geomorphological processes.

Geomatic techniques (topographical survey and terrestrial photogrammetry) were used to identify the main features and annual changes in the ice mass and surroundings. The outcropping ice patch perimeter was measured and two transverse cross sections were measured in 2007 and 2008 (Fig. 4). These provide information on changes in area, volume and melting in response to climate.

GPR prospection was undertaken (equipment RAMAC/GPR device from Måla GeoScience) using 200 MHz (unshielded) and 500 MHz (shielded) antennas, each with its own central control unit. A set of four georeferenced profiles was made in different zones in October 2007 and 2008 (Fig. 3). Fifteen radargrams were recorded in the 2007 campaign, and 13 in 2008. Seven of the radargrams were recorded using the *Common Mid-Point (CMP)* technique with the 200 MHz antenna. The same profiles were performed in both campaigns, such that the topographic positioning of the traces of the profiles (2008 compared to 2007) shows errors of less than 5 cm (Del Río *et al*. 2009).

Results and analysis

Ice patch evolution

Ice patch evolution and climate historical scientific sources (Prado 1860; Saint Saud 1893; Penck 1897) and recent geomorphological studies point to the existence of glaciers in the Picos de Europa during the LIA (González-Trueba *et al*. 2008; González-Trueba 2006, 2007a, 2007b, 2007c). In the Jou Negro, testimonies of experienced climbers, explorers and geographers indicate that there was a glacier until the end of the nineteenth century. The presence of a voluminous and well preserved moraine and internal deformation structures such as ice-folds and faults confirm this fact. It would have been a small glacier (glacieret), without tongue and with a steep ice surface, located at very low elevation (2230 m a.s.l.). It was 280 m long with a surface area of approximately 5.2 ha and ice thickness of around 19–20 m in the frontal part and 30–32 m in the middle part. The *Equilibrium Line Elevation (ELE)* during the maximum advance of the LIA has been estimated by the AAR method at 2287 ± 0.05 m a.s.l. (González-Trueba 2006). It is a very low ELE, controlled by topoclimatic factors, much lower than the regional mean ELE, located above the highest summits. The key factors for the development of the Jou Negro glacier during the LIA were aspect and topographical features. The cirque's steep 400 m high walls protected it from solar radiation and snow avalanches fed it (González-Trueba 2006, 2007a).

During the LIA a glacier developed by net accumulation of snow, firnification, transformation into ice and motion, and the frontal moraine was built-up. The survival of the glacier in the Jou Negro into the first decades of the twentieth century is not supported by testimonies from the period, and some authors deny its existence (Obermaier 1914). In the 1990s different studies point to the presence of an ice-patch inherited from the LIA in the Jou Negro cirque. González-Suárez and Alonso (1994, 1996) support the existence of a glacier, the Jou Negro glacier, but they are rebutted by Frochoso and Castañón (1995) who, based on the size of the ice mass and the absence of glacial activity, reject the idea, and the ice patch presence was accepted (Alonso and González-Suárez 1998). Recent research has focused on the last historical advance, the LIA in the Picos de Europa and the Jou Negro (Serrano and González-Trueba 2002; González-Trueba 2006, 2007a, 2007b; González-Trueba *et al*. 2008).

The reconstruction of the Jou Negro glacier surface extension by mean of historical data and cartographic works confirms the loss of 88.5 % of the outcropping ice surface since maximum expansion (Table 1, Fig. 5). The clast cover hides the ice patch extent at the edges, so the ice patch surface measured in the field can be increased. The real reduction of the surface area can be estimated at between 53 % and 70 %. Up until the 1980s the loss of thickness was dominant over that of surface, with a sharp thinning of 15 metres in the middle part. At present there are no longer any crevasses

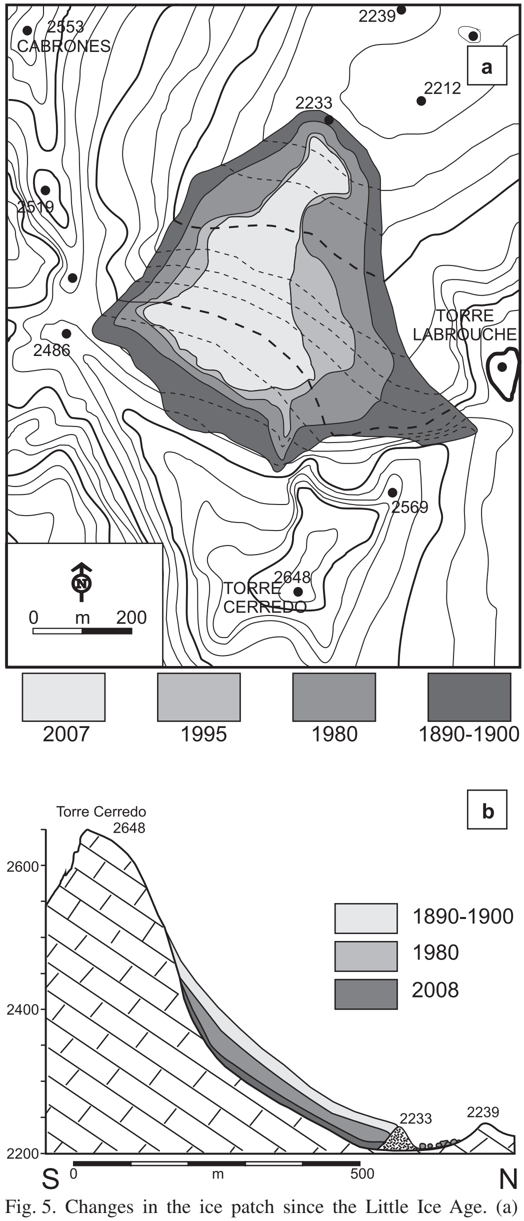

Fig. 5. Changes in the ice patch since the Little Ice Age. (a) surface changes, (b) depth changes.

nor bergschrund in the ice body, only a melt-based false bergschrund and the bevelled ice front. It is, therefore relict glacier ice, or dead ice, forming an ice patch. The analysis of the glacier's structures and possible movement from existing images denotes that the ice has not undergone internal deformation in the last 30 years (1980–2010). Between 1980 and 1995 the ice patch lost around

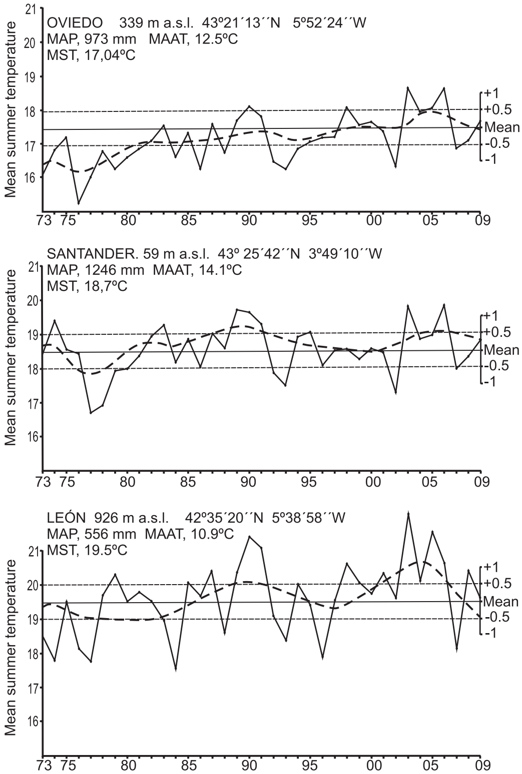

Fig. 6. Mean summer (June to September) temperatures and deviation to the average of 1973–2009 in Santander, Oviedo and León. The dashed line is the 5-year running average.

half of its surface with respect to the LIA glacier. This can be the period of the greatest loss of ice since the twentieth century began, and has left it reduced to a concave tongue-shaped ice mass partially covered by debris. From 1995 to 2007, 76 % of the surface of outcropping ice was lost, however, at a lesser rate than previously thought (Table 1, Fig. 5). Debris cover increased significantly.

On the surface of the ice patch there are several sinkholes produced by melting water flowing towards the endokarstic system. The vertical sinkholes or "moulins" permitted ice thickness to be measured, and values between 14.3 and 7.5 m were recorded 14 years ago in large moulins that have now disappeared (Alonso and González-Suarez, 1998), and 13 m in small moulins measured in 2007. GPR sounding showed a depth of 8–12 m in the same area, showing that there has been no significant thinning from 1995 to 2010.

The five year MST moving average (Fig. 6) reveals a temperature rise between 0.64°C and 1.3°C, with a mean increase of 1°C for the region. This increase is not linear, and warmer and cooler summers alternate (Table 2). Warm summers dominate since 1989, 25 % of which are concentrated in three periods: 1989–1991, 1997–2001 and 2003–2006. Until 1989, warm summers were only 7.2 % of cases, against 11.2 % of cool summers. The absence of data of spring snowfall makes a detailed analysis of the years most favourable to the retreat of the ice patch impossible. Nevertheless, the succession of warm summers from 1987 to 1991 coincides with a period in which the loss of the ice mass escalates sharply. From 2003 to 2006, four consecutive years with warm summers occurred, including 2003, which was the warmest summer in Europe in the last 500 years (Lauterbacher *et al*. 2004) and the warmest in the region together with 2006. This last period established the extent of the ice patch to what it is in 2010.

Field observations in 1995–1996, 2004 and annually from 2006 to 2009, and comparison with the temperature record, show that spring snowfall together with cloudiness are key factors for the summer melt of the snow cover and of the ice patch. In 1995, 2007 and 2008 snow melted completely over the ice patch at the end of the summer, with temperature conditions close to the mean. During the study period, the ice patch remained snow covered until the middle of August and was snow free in September. On the other hand, in 1996 and 2009 (characterized by summer temperatures close to MST) and in 2004 and 2006 (characterized by warm summers), the ice patch did not become free of snow. The surroundings of the ice patch are completely snow free every year, even with large spring snowfall, as occurred in 2009.

If large quantities of snow fed by avalanches and wind were necessary during the LIA for the glacier to develop in relatively warm conditions, during the retreat, slight increases in winter precipitation and cool summers should have been enough to generate small stabilizations, such as occurred in small European glaciers (see Kaser 2001; Grunewald and Scheithauer 2010). Moreover, in maritime environments winter and spring precipitation is more influential than in continental Alpine or Pyrenean areas (Steiner *et al*. 2008). The oceanic conditions of the Picos de Europa, with high precipitation and cloudiness, favour the persistence of the ice patch. Results show, however, that the summer temperature shows an influence, mainly in the medium and long term in explaining

Table 2. Frequency of cool and warm summers at the Picos de Europa from 1973 to 2009.

Years	Summer type	Oviedo (%)	Santander (%)	León (%)	Mean (%)
1973–1989	Cool	10.8	8.1	16.2	11.7
	Warm	5.4	8.1	8.1	7.2
1989–2009	Cool	8.1	10.8	8.1	9
	Warm	32	27	16.2	25.6
Changes	Cool	−2.7	+2.7	−8.1	–
	Warm	+26.6	+18.9	+8.1	–

the retreat of the ice masses, as the topoclimatic conditions are determinant. The main factors determining the cycles of the melting of snow over the ice patch and of the ice patch itself should be, therefore, snow avalanching and wind blown snow, spring snow fall and summer temperatures.

The internal structure of the ice patch

In the 2007 campaign we analysed two points at the intersections of the profiles of the upper (1) and lower (2) zones of the ice patch by CMP (intersections between GPR 1 and 2, and GPR 3 and 4 in Fig. 4). In Zone 1, the radargram showed a horizontal or semi-horizontal reflection at 11 m depth with a wave propagation velocity of 0.208 m ns^{-1}. In Zone 2, the analysis showed two reflections, at 4.8 and 12.3 m, with wave propagation velocities of 0.187 and 0.170 m ns^{-1}, respectively. In the 2008 campaign, we analysed the two points located at the intersections of profiles 3 and 4 (GPR3 and GPR4 in Fig. 4). In Zone 2, the radargram shows several possible reflections between 1.7 and 5.5 m depth, with propagation velocities in the range 0.209 to 0.168 m ns^{-1}. However, in this case it was impossible to identify other deeper reflections, unlike the case of the 2007 campaign. In Zone 3, the two radargrams recorded at the same point showed two reflections at 1.5 and 9.2 m, with wave propagation velocities of 0.199 and 0.170 m ns^{-1}, respectively (Del Río *et al*. 2009). The values of the propagation velocities, between 0.168 and 0.209 m ns^{-1}, and the dielectric permittivity, whose values were between 3.2 and 2.1, are coherent with those reported in the literature for ice (0.150–0.173 m ns^{-1}) and snow (0.212–0.245 m ns^{-1}) propagation velocities, and for ice (3–4) and snow (1.5–2) dielectric permittivities (Jakobsen and Overgaard 2002; Brandt *et al*. 2007).

The GPR profile has a length of 50 m with a difference in elevation from beginning to end of

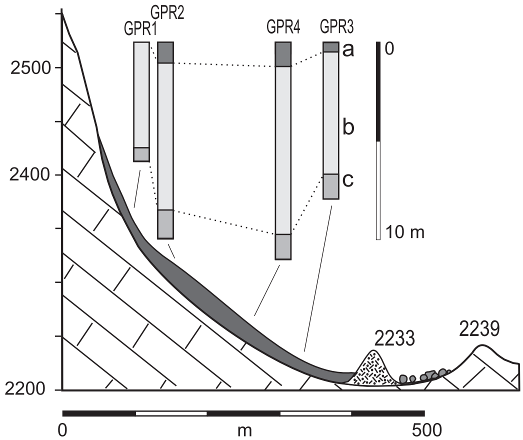

Fig. 7. Internal structure of the Jou Negro ice patch from GPR soundings. (a) mixed layer of ice, snow and clasts, (b) banded glacial ice, (c) clastic layer with glacier ice, detrital material, till and flowing melt water.

+17.4 m. It is possible to appreciate the four-layer internal structure of the ice-patch (Fig. 7) from bottom to top: (1) a thick layer of debris arranged haphazardly that we interpret as subglacial till (c in Fig. 7), (2) a transitional clastic layer corresponding to a zone of glacier ice, detrital material, and melt water, (3) banded glacial ice that formed the main body of the ice patch during the LIA, and reaching between 7 and 10 m depth (b in Fig. 7), and (4) a mixed layer of ice, snow, and clasts of great continuity formed by the accumulation of snow and sleet on the surface, mixed with fragments of debris fallen from the cirque walls (a in Fig. 7).

From the velocities provided by the CMP radargrams, assuming a constant velocity of 0.208 m ns^{-1} in the zone of the GPR 2 profile, the bedrock is found at depths varying from 7 m in the part closest to the south wall to 10 m estimated for the central part of the ice patch. The analysis of the GPR 1 profile gave a practically constant depth to

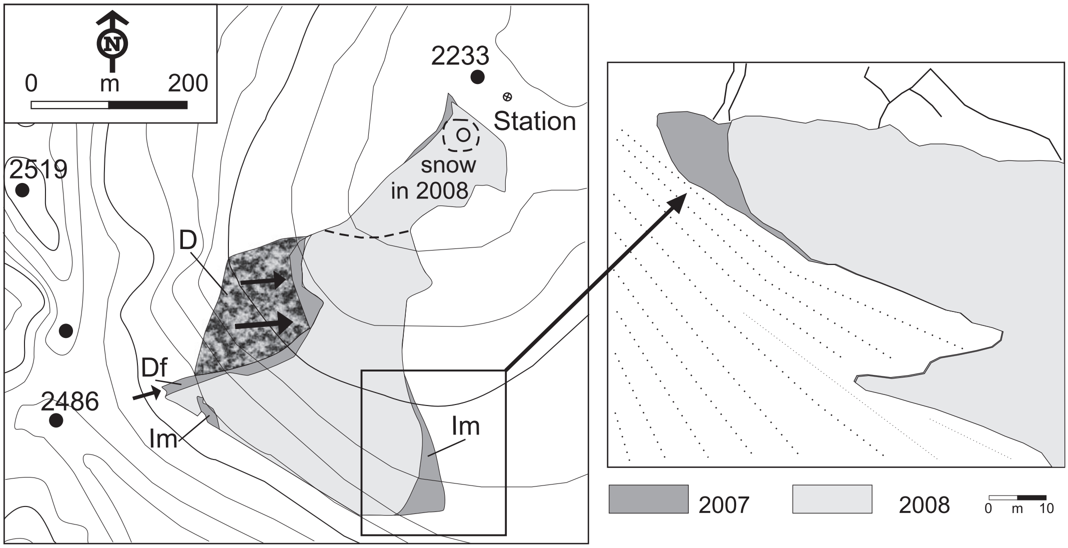

Fig. 8. Changes in the surface of the Jou Negro ice patch between 2007 and 2008. Ice loss and debris cover in the left picture. Ice loss on the SE side in the right picture.

the substrate of 10 m over the entire length of the profile. In the low zone, the GPR 4 profile shows depths to the substrate of 9 m in 2007 and in the 8–12 m range in 2008. In this latter case the greatest depth occurred in the most central part of the ice-patch, and the shallowest close to the moraine. For the GPR 3 profile, in 2007 we found the substrate at depths of 7–10 m. Finally, in the middle part (zone 3), the GPR 6 profile placed the bedrock at 7–10 m, while the GPR 5 profile placed it at 7–12 m.

Geomorphic processes and present-day evolution of the ice patch

The Jou Negro ice patch has a length of 173 m and an orientation of 58° (NE). The average width is 51.93 m, with a minimum value of 20 m and a maximum of 120 m. Comparing the geomatic measurements taken in 2007 and 2008, the outcropping ice surface has fallen by 28.78 %, from 9327 m^2 to 6642 m^2. This rate is four times higher than the annual rate estimated for the last 13 years (7.68 % yr^{-1}), excluding the recovered processes by clast from the wall, without surface loss. In the period 1995 to 2008 76 % of the outcropping ice surface was lost, at an average rate of 1460 m^2 yr^{-1}. During 2007–2008 the rate ascended to 2685 m^2 yr^{-1}, which almost doubles the mean rate of the last 13 years. The surface loss in 2007–2008 took place in the higher limits of the ice patch, where slope is steepest, the ice is in contact with the substrate and there is no debris cover (Fig. 8).

Between 2007 and 2008 at least 2 % of the ice patch surface was covered by clasts. For the period up to 2007, it is not possible to delimitate the extent of the ice patch exactly and, therefore, the rate of melt cannot be estimated. The following two types of processes are involved in the burial of the ice:

1. High frequency and low intensity processes, such as debris flows, rockfalls, runoff and nivation. Debris flow deposits occur at the foot of the SW and SE corridors and from Torre Cerredo, partially burying the ice patch. During spring and summer 2008 debris flows from the SW corridor built a furrow with two lateral ridges and a frontal fan that partly destroyed the ice. From the moraine and the wall of the western side, the rock falls involve a clast cover progression of around 35 m between 2007 and 2008.
2. High intensity and low frequency processes, slides and rock falls extending from walls and moraines. They cover the lower parts of the ice patch and produce accumulations of blocks over it.

The topographical profiles made on the upper and middle parts of the ice patch have permitted

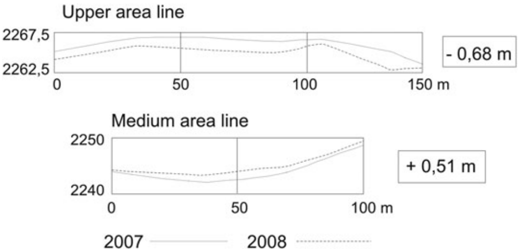

Fig. 9. Topographical changes in the upper (line 1) and medium (line 2) areas of the Jou Negro ice patch surface.

centimetre scale vertical variations to be measured (Fig. 9). In the upper profile an average lowering of 68 cm was observed, whereas in the lower profile an average rise of 51 cm was recorded. The following hypotheses could explain the lowering of the upper profile: the loss of ice mass and consequent thinning by melting; internal motion by flow and slide of the ice mass (which would qualify it as a small glacier), or the slide of the mass. The thickening of the lower profile might be explained by the following hypotheses: an increase in the ice mass or firn (ruled out as the firnification of snow is not possible in one year and is not supported by field observation); the internal motion by flow (ruled out due to the absence of features or internal motion structures – crevasses, folds); or due to a pushing movement of the ice block from the upper part, which partially lifts the lower one. We propose that the observations in both the upper and lower profile may be explained by rotational sliding of the ice body.

This sliding motion involves the balancing of the ice body such that to the south, at the head, it scrapes on the walls of the cirque, producing the partial breakage of the frozen body at the uppermost edges as well as faster melting as a result of being in contact with the rock. This process explains the important loss of mass of the upper part of the ice patch in just one year through the joint action of pressure, partial breakages and ice block falls, combined with the summer melting due to the direct contact with the rock. The internal structure, with a lower layer of clasts saturated during the periods of spring and summer snow melt, behaves as a deforming layer under the ice mass and promotes sliding of the overlying ice body, in this case with a rotational component.

Conclusions

Glacial dynamics at the Jou Negro, initiated during the LIA, ceased at the start of the twentieth century. The remaining ice mass is now a glacial ice patch formed by residual or dead glacier ice, stratified, folded and fractured, but without deformation or internal motion flow and partially covered by clasts. The use of historic documents, geomatic and geophysical techniques showed to be efficient in the quantification of the changes in the ice patch. The current basic morphometric and morphological character of the ice patch has been established and processes covering the ice body, such as rockfalls, slides and debris flows, have been detected. From the analysis of radargrams a four-layer internal structure has been observed.

Present day changes in the Jou Negro ice patch are due to melting and sliding of the ice mass. The ice patch has lost around 53–70 % of its surface since the LIA (88.5 % of outcropping ice). The melt rates increased during the twentieth century, especially from the late 1990s, and in 2007–2008 surface losses of over double the mean of the last 15 years have been detected. The reason for this high rate is not clear yet. The loss of ice is associated with periods with lower spring snow fall and warm summers which favour both melt and rotational sliding of the ice body.

The ice patch persists under a relatively temperate environment (MAAT ca 0°C), by the compensatory effect of topoclimatic factors, such as a northeast-facing aspect, high rates of snow accumulation by avalanches and wind, and the very steep cirque walls. The ice patch is in disequilibrium with the present-day environmental conditions, with a regressive dynamic and a complex response to the regional climatic variations, in which the interannual variability of summer temperatures, extraordinary spring precipitations and snow cover determine the relict ice melt or permanence.

Topoclimatic factors (e.g., snow avalanches, shaded slopes) are not sufficient to protect the ice patch if the spring is not snowy or the summer exceptionally warm. The balance between wet and cold springs and warm summers is the key to understanding the rhythm of melt or accumulation on ice patches in marginal high mountain environments but also on small glaciers in marginal environments (glacier cirques below the regional equilibrium line altitude). Ice patch equilibrium and dynamics are controlled by topoclimatic factors and climatic variability in the short term, and environmental conditions and climate changes medium to long term, similar to small glaciers in marginal environments.

Acknowledgements

This work was funded by the projects CGL-2010-19729/BTE and CGL2007-65295/BTE (Ministry of Science & Innovation and FEDER) and OAPN 0534/2010 of Ministry of the Environment and Rural and Marine Environment. We thank A. Atkinson, Angel Fernández, Diego Fernández, R. Martín, J. de Matías and I. Rico, for field assistance and P.D. Hughes and an anonymous reviewer for helpful comments.

Enrique Serrano, Departament of Geography, Universidad de Valladolid, Paseo Prado de la Magdalena s/n, 47011 Valladolid. Spain
Email: serranoe@fyl.uva.es

Juan José González-Trueba, Departamento de Geografía, Prehistoria y Arqueología, Universidad del País Vasco, C/Tomás y Valiente s/n, 01006 Vitoria-Gasteiz, Spain
Email: jjgtrueba@mail.com

José Juan Sanjosé, Departamento de Expresión Gráfica, Universidad de Extremadura
Email: jjblasco@unex.es

Luis Mariano Del Río, Departamento de Física Aplicada. Escuela Politécnica de Cáceres, Universidad de Extremadura
Email: lmdelrio@unex.es

References

Alonso, V. and González-Suárez, J.J., 1998. Presencia de hielo glaciar en los Picos de Europa (Cordillera Cantábrica). El helero del Jou Negro. [Glaciar ice ocurrence in the Picos de Europa. The Jou Negro ice patch]. *Cuaternario y Geomorfología*, 12(1–2), 35–44.

Assier, A., 1993. *L'Englacement des Alpes de Sud Franco-Italienes du Petit Age de Glace a la fin du XXéme siecle*. Institute de Geographie Alpine-CNRS, Grenoble.

Barsch, D., 1996. *Rockglaciers. Indicators for the Present and Former Geoecology in High Mountain Environments*. Springer Verlag, Berlin.

Benn, D.I. and Evans, D.J.A., 1998. *Glaciers and Glaciation*. Edward Arnold, London.

Braithwaite, R.J., Zhang, Y. and Raper, S.C.B., 2003. Temperature sensitivity of the mass balance of mountain glaciers and ice caps as a climatological characteristic. *Zeitschrift Gletscherkunde und Glazialgeologie*, 38, 35–61.

Brandt, O., Langley, K., Kohler J. and Hamran, S.E., 2007. Detection of buried ice and sediment layers in permafrost using multi-frequency Ground Penetrating Radar: A case examination on Svalbard. *Remote Sensing of Environment*, 111, 212–227.

Christiansen, H.H., 1998. Nivation forms and processes in unconsolidated sediments, NE Greenland. *Earth Surface Processes and Landforms*, 23, 751–760.

Del Río, L.M., Tejado, J.J., de Sanjosé, J.J., Atkinson, A., Serrano, E., González, J.J. and Fernández, A., 2009. Ice patch structure and depth using GPR techniques: a first approach to the Jou Negro ice patch (Picos de Europa, Spain). In: *Proceeding 5th International Workshop on Advanced Ground Penetrating Radar* (IWAGPR 2009). Granada. 122–126.

Diolaiuti, G. and Smiraglia, C., 2010. Changing glacier in a changing climate: how vanishing geomorphosites have been driving depth changes on mountain landscape and environment. *Geomorphologie*, 2, 131–152.

Diolaiuti, G., Smiraglia, C., Pelfini, M., Beló, M., Pavan, M. and Vassena, G., 2006. The recent evolution of an Alpine glacier used for summer skiing (Vedretta Piana, Stelvio Pass, Italy). *Cold Regions Science and Technology*, 44, 206–216.

European Nature Information System (EUNIS), 2009. *EUNIS habitats Thesaurus Framework for Nature Conservation*. European Economic Community (EEC). http://eunis.eea.europa.eu/glossary.jsp. 20 March 2010.

Frochoso, M. and Castañón, J.C., 1995. Comments on "Glaciers in Picos de Europa, Cordillera Cantábrica, northwest Spain" by González-Suárez and Alonso. *Journal of Glaciology*, 41, 430-432.

Galibert, G., 1960. Les processus actuels d'évolution du relief de Haute Montagne dans le massif de Zermatt (Alpes Valaisannes). *Bulletin de l'Association de Géographes Français*, 290–291, 70–86.

Galibert, G., 1965. *La haute montagne alpine. L'évolution actuale des formes dans les hautes massifs des Alpes et dans certains reliefs des comparation*. CNRS, Toulouse.

Gellatly, G.F., Smiraglia, C., Grove, J.M. and Latham, R., 1994. Recent variations of ghiacciao del Calderone, Abrizzi, Italy. *Journal of Glaciology*, 40, 486–490.

Gerrard, A.J., 1990. *Mountain Environments*. Belhaven, London.

Gómez-Ortiz, A., García, A., Ortega, F., Salvador, F., Sánchez, S., Sanz, C., Schulte, L. and Simón, M., 2002. *Geomorphological Map of Sierra Nevada. Glacial and Periglacial Geomorphology*. Junta de Andalucía, Sevilla.

Gómez-Ortiz, A., Schulte, L. and Salvador, F., 1996. Contribución al conocimiento de la deglaciación reciente y morfología asociada del Corral del Veleta (Sierra Nevada). [Contribution to the knowledge on recent deglaciation and morphology of the Corral del Veleta]. *Cadernos do laboratorio Xeolóxico de Laxe*, 21, 543–558.

Gómez-Ortiz, A., Schulte, L., Salvador, F., Palacios, D., San José, J.J. and Atkinson, A., 2004. Deglaciación reciente de Sierra Nevada. Repercusiones morfogénicas, nuevos datos y perspectivas de estudio futuro. [Recent deglaciation of Sierra Nevada. Morphogenetic results, new data and future perspectives]. *Cuadernos de Investigación Geográfica*, 30, 147–168.

González-Suárez, J.J. and Alonso, V., 1994. Glaciers in Picos de Europa, Cordillera Cantábrica, northwest Spain. *Journal of Glaciology*, 40, 198–199.

González-Suárez, J.J. and Alonso, V., 1996. Reply to the comments of Frochoso and Castañón on "Glaciers in Picos de Europa, Cordillera Cantábrica, NW Spain" by González-Suárez and Alonso. *Journal of Glaciology*, 42, 386–389.

González-Trueba, J.J., 2006. Topoclimatical factors and very small glaciers in Atlantic Mountain of SW Europe: The Little Ice Age glacier advance in Picos de Europa (NW Spain). *Zeitschrift für Gletscherkunde und Glazialgeologie*, 39, 115–125.

González-Trueba, J.J., 2007a. *La Pequeña Edad del Hielo en los Picos de Europa*. [Little Ice Age in the Picos de Europa], Universidad de Cantabria-Fundación Marcelino Botín, Santander.

González-Trueba, J.J., 2007b. *El paisaje natural de los Picos de Europa: Geomorfología y sus implicaciones geoecológicas en la alta montaña cantábrica. [Natural Landscape of the Picos de Europa*. Geomorphology and geoecological results in the Cantabrian high Mountain], Centro de Investigación del Medio Ambiente (CIMA), Gobierno de Cantabria, Santander.

González-Trueba, J.J., 2007c. *Geomorfología del Macizo Central de los Picos de Europa, con Mapa Geomorfológico a escala 1/25.000*. [Geomorphology of the Central Massif of Picos de Europa, with a geomorphological map to 1/25,000 scale], OAPN, Ministerio de Medio Ambiente, Madrid.

González-Trueba, J.J., Martin Moreno, R., Martínez de Pisón, E. and Serrano, E., 2008. Little Ice Age glacier advance and current glaciers in the Iberian Peninsula. *The Holocene*, 18, 551–568.

Grudd, H., 1989. Small glaciers as sensitive indicators of climatic fluctuations. *Geografiska Annaler, Series A: Physical Geography*, 72, 119–123.

Grunewald, K. and Scheithauer, J., 2010. Europe's southermost glaciers: response and adaptation to climate change. *Journal of Glaciology*, 56, 129–142.

Guyton, B., 2001. Glaciers of California. *Modern Glaciers, Ice Age Glaciers, the Origin of Yosemite Valley and a Glacier Tour in the Sierra Nevada*. University of California Press, Berkeley.

Harris, C. and Murton, J.B., (eds), 2005. *Cryospheric Systems: Glaciers and Permafrost*. The Geological Society of London, Special Publications, 242, London.

Hoffman, J., Matthew J., Fountain, A.G. and Achuff, J.M., 2007. 20th-century variations in area of cirque glaciers and glacierets, Rocky Mountain National Park, Rocky Mountains, Colorado, USA. *Annals of Glaciology*, 46, 349–354.

Hughes, P.D., 2007. Recent behaviour of the Debeli Namet glacier, Durmitor, Montenegro. *Earth Surface Processes and Landforms*, 32, 1593–1602.

Hughes, P.D., 2008. Response of a Montenegro glacier to extreme summer heat waves in 2003 and 2007. *Geografiska Annaler, Series A: Physical Geography*, 192, 259–267.

Hughes, P.D., 2009. Twenty-first century glaciers in the Prokletije Mountains, Albania. *Arctic, Antarctic and Alpine Research*, 4, 455–459.

Hughes, P.D., 2010. Little Ice Age glaciers in Balkans: low altitude glaciation enabled by cooler temperatures and local topoclimatic controls. *Earth Surface Processes and Landforms*, 35, 229–441.

Hughes, P.D. and Woodward, J.C., 2009. Glacial and periglacial environments. In: Woodward, J.C. (ed.), *The Physical Geography of the Mediterranean Basin*. Oxford University Press, Oxford. 353–383.

IPA (International Permafrost Association), 2005. *Glossary of Permafrost and Related Ground Ice Terms*. IPA. The Arctic Institute of North America. The University of Calgary. Calgary.

Jakobsen R. and Overgaard, T., 2002. Georadar facies and glaciotectonic structures in ice marginal deposits, northwest Zealand, Denmark. *Quaternary Science Reviews*, 21, 917–927.

Kaser, G., 2001. Glacier–climate interaction at low latitudes. *Journal of Glaciology*, 47, 195–204.

Kotlyakov, V.M. and Smolyarova, N.A., 1990. *Dictionary of Glaciology*. Elsevier, Rotterdam.

Kuhn, M., 1995. The mass balance of very small glaciers. *Zeitschrift für Gletscherkunde und Glazialgeologie*, 31, 171–179.

Lliboutry, L., 1965. *Traité de Glaciologie*. Masson, Paris.

Luterbacher, J., Dietrich, D., Xoplaki, E., Grosjean, M. and Wanner, H., 2004. European seasonal and annual temperature variability, trends, and extremes since 1500. *Science*, 303, 1499–1503.

Martínez de Pisón, E. and Arenillas, M., 1988. Los glaciares actuales del Pirineo español. [Present day glaciaers in the Spanish Pyrenees] In: *La nieve en el Pirineo Español* [*The Snow in the Spanish Pyrenees*]. MOPU, Madrid. 29–98.

Martínez de Pisón, E., Navarro, P., Martínez, I., Cantarino, I., Pedrero, A. and Arenillas, M., 1995. Observaciones morfológicas sobre la evolución de los glaciares del Pirineo español entre 1982 y 1993. [Morphological observations on the evolution of glaciers in the Spanish Pyrenees between 1982 and 1993], In: *La nieve en las Cordilleras españolas 1993/94* [The snow in the Spanish Mountains 1993/1994]. Programa ERHIN, MOPTMA, Madrid. 253–281.

Martínez de Pisón, E., Navarro, J., Martínez, R., Pedrero, A., Arenillas, M. and Bisbal, L., 1997. Observaciones sobre el estado de los glaciares del Pirineo español. Campaña 1994. [Observations on the state of glaciers in the Spanish Pyrenees. Fieldwork 1994], In: *La nieve en las Cordilleras españolas*. [The snow in the Spanish Mountains] Ministerio de Medio Ambiente, Madrid. 253–265.

NSIDC (National Snow and Ice Data Center), 2009. NSIDC snow glossary. http://nsidc.org/cgi-bin/words/glossary.pl. (12 June 2010).

Obermaier, H., 1914. *Estudio de los glaciares de los Picos de Europa.* [*Study on the Glaciers of Picos de Europa*], Trabajos del Museo Nacional de Ciencias Naturales, Serie Geología 9, Madrid.

Ohmura, A., Kasser, P. and Funk, M., 1992. Climate at the equilibrium line of glaciers. *Journal of Glaciology*, 38, 397–411.

Orefice, M., Pecci, M., Smiragglia, C. and Ventura, R., 2000. Retreat of Mediterranean glaciers since the LIA: Case of study of Ghiacciaio del Calderone, Central Apennninos, Irali. *Arctic, Antarctic and Alpine Research*, 32, 197–201.

Penck, A., 1897. Die Picos de Europa und das kantabrische Gebirge. *Geographische Zeitschrift*, 278–281.

Prado, C. de, 1860. Valdeón, Caín, La Canal de Trea. Ascensión a los Picos de Europa en la Cordillera Cantábrica [Valdeón, Caín, La Canal de Trea. Climbing to the Picos de Europa in the Cantabrian Mountains]. *Revista Minera*, 62, 92–101.

Rau, F., Mauz, F., Vogt, S., Singh, S.J. and Raup, B., 2005. Illustrated GLIMS glacier classification guidance for the GLIMS glacier inventory. GLIMS Guides and Tutorials, 2005. Versión 1.0. http://www.glims.org (7 Feb. 2011).

Saint-Saud, A.A., 1893. Les Picos de Europa (Monts Cantabriques). Etude Orographique. *Annales Club Alpine Français*, Paris.

Serrano, E., 1998. *Geomorfología del Alto Gállego (Pirineo aragonés) [Geomorphology of the High Gállego (Aragonese Pyrenees)]*. Fundación Fernando El Católico. Zaragoza.

Serrano E. and González-Trueba, J.J., 2002. Morfología y evolución glaciar en los Picos de Europa [Geomorphology and glacial evolution in the Picos de Europa]. In: Redondo, J.M., Gómez, A., González, R.B., Carrera, P. (eds), *El modelado de origen glaciar en las Montañas Leonesas*. Universidad de Leon, León. 249–267.

Steiner, D., Pauling, A., Nussbaumer, S.U., Nesje, A., Luterbacher, J., Wanner, H. and Zumbühl, H.J., 2008. Sensitivity of European glaciers to precipitation and temperature – two case studies. *Climatic Change*, 90, 413–441.

Thorn, C., 1988. Nivation: a geomporphic chimera. In: Clarck, M.J. (ed.), *Advances in Periglacial Geomorphology*. Wiley and Sons, Chichester. 3–32.

WGMS (World Glacier Monitoring Survey), (2005). *Fluctuations of Glaciers 1995–2000*. IUGG-UNEP-UNESCO, Paris.

Manuscript received Jul., 2010, revised and accepted Nov., 2010