

The Periglaciation of Europe



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1 Geographical Distribution of Periglacial Phenomena in Europe

The vast surface of the European continent (~10 million km²) extending from subtropical (36°N) to polar latitudes (71°N) displays a complex relief configuration including numerous peninsulas, islands, archipelagos, capes, bays and high mountain ranges exceeding 2,500–3,000 m. In addition, mainland Europe constitutes a large peninsula surrounded by the relatively warm Mediterranean Sea and the cooler Atlantic Ocean in its southern and western-northern flanks, respectively. This unique geographical setting determines the occurrence of a wide spectrum of climate regimes, which influence the nature of the geomorphological processes affecting the Earth surface processes in this continent.

Taking into account French's (2007) broad definition of the periglacial environment as the non-glacierised areas where the mean annual air temperatures (MAAT) oscillate between $-2\text{ }^{\circ}\text{C}$ and $+3\text{ }^{\circ}\text{C}$, currently in Europe periglacial regions are only located in mountain ranges, northern Scandinavia and the high-altitude islands (i.e. Iceland, Jan Mayen, Svalbard) (Fig. 1). Following north–south temperature gradients, there is a strong latitudinal influence of the periglacial domain: whereas the lower elevation of the periglacial belt in the Sierra Nevada is located at ~2,650 m, in the Alps it is located at ~1,950 m and in northern Scandinavia periglacial processes

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are currently active down to sea level. In addition, in some of the high-latitude islands, such as Svalbard, periglacial processes are also widespread in glacier-free areas.

Therefore, the present-day active periglacial domain in southern and central Europe is restricted to upper parts of the highest mountain ranges although periglacial evidence is also found at lower altitudes, and even in the flat lowlands, and in mid-low altitude massifs (Oliva et al. 2018b). Actually, deep permafrost at a depth of several hundred metres from the surface has been also detected in Poland (see chapter “[The North European Plain](#)”). In these regions, we find a combination of active features with others inherited from previous cold-climate stages, namely during Quaternary glacial periods. In northern Europe, where vast regions were under large ice sheets during the Last Glacial Cycle (i.e. Scandinavia, Iceland, Svalbard, large part of Great Britain, and northern part of North European Plain) active periglacial processes can coexist in the coldest non-glaciated areas together with periglacial phenomena formed during the deglaciation of current glacier-free environments (Fig. 2).

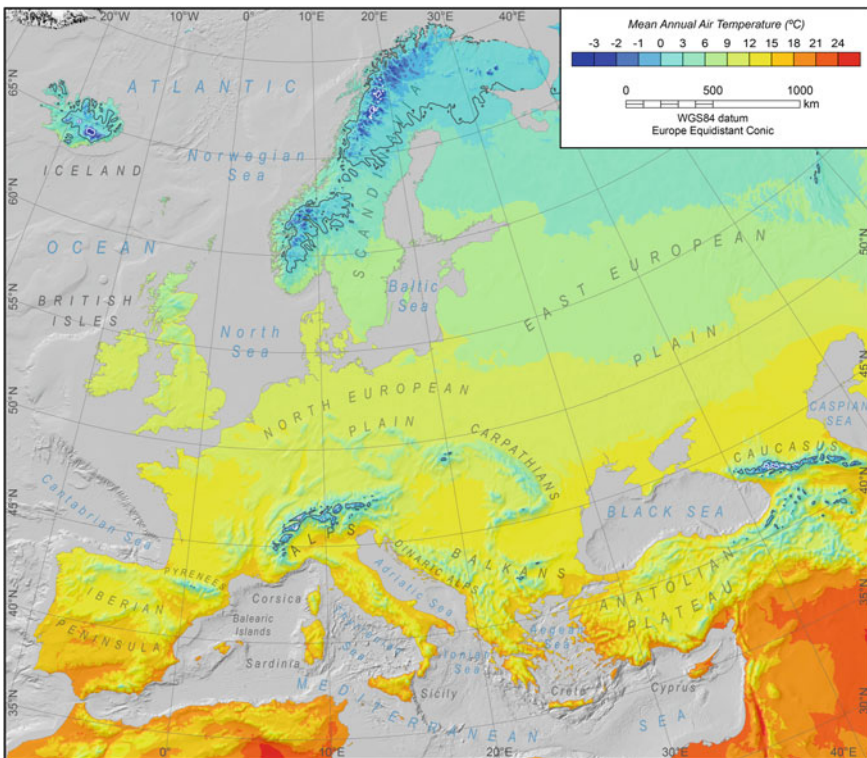


Fig. 1 Map of Europe showing present-day MAAT. The white and black contour line encircles the areas included between -3 and 2 °C following the general definition of French (2007) for the periglacial environment. The line represents the transects shown in Fig. 2

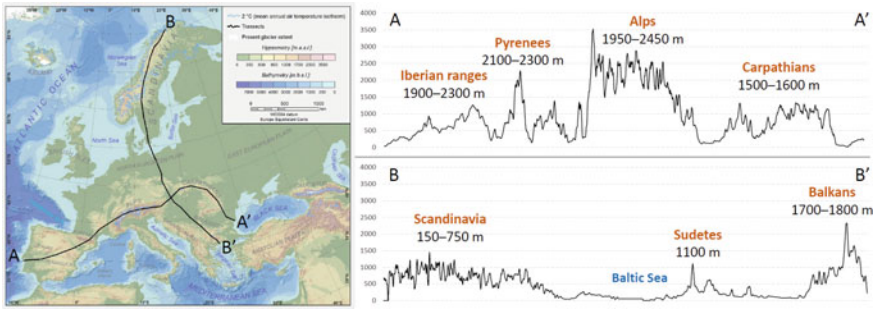


Fig. 2 North-South (A-A') and West-East (B-B') transects showing the lower limit of periglacial processes and permafrost belt in European mountains

In this section, we will integrate knowledge presented separately in the different chapters in order to provide a more general overview on European periglacial landscapes. We will first focus both on present-day landforms, their current activity, ground thermal dynamics and factors controlling their movement. Subsequently, we will examine the distribution and types of relict landforms formed in the past, which can be used to infer Late Quaternary climate conditions. Finally, we will also describe how, in some cases, these periglacial features have been altered by human activities.

2 Ground Thermal Regime of Present-Day Periglacial Processes

Present-day periglacial processes in Europe are generally driven by seasonal frost dynamics, as permafrost conditions are only recorded in the highest mountain ranges as well as in the lowland regions of northern Scandinavia and high-latitude archipelagos, such as Svalbard. Consequently, periglacial landscapes include, thus, cold non-glacial landforms generated by both seasonal frost and permafrost ground thermal regimes.

Seasonal frost in Europe is favoured by the moderately low temperatures recorded in winter throughout the continent, particularly across the mountains, which promoted the freezing of the first decimetres to several metres of the ground. Negative temperatures during the cold season of the year are widespread across the continent and diurnal one-sided freezing can result in the ephemeral formation of some frost action features (i.e. frost heaving, needle ice). However, periglacial environments are those where processes related to ground freezing are continuous during a certain time of the year and the consequences of freeze-thaw oscillations result in identifiable features across the landscape (French 2007). For example, even if the average winter air temperature of Stockholm and Warsaw are approximate $-1\text{ }^{\circ}\text{C}$ —their mean annual air temperatures (MAAT)s are ~ 7 and $9\text{ }^{\circ}\text{C}$, respectively—these climate conditions are not able to significantly modify the landscape.

Normally, there are substantial differences between air and ground temperatures as well as factors driven by the local lithology (e.g., spatio-temporal patterns of snow cover, topographical conditions, albedo), which can contribute to the cooling or warming of the ground regardless atmospheric conditions. For example, in the Sierra Nevada, south Iberia, the seasonal frost environment extends from above 2,500 m to the summit level (3,300–3,400 m) where the MAATs is 0 °C, but the annual soil temperatures average is 2–2.5 °C (Oliva et al. 2016a). Generally, the calendar, duration and depth of seasonally-frozen ground in Europe is highly variable and largely controlled by the snow precipitation regime, topographical and lithological conditions. In latitudes >65 °N and in high mountains exceeding 2,000–2,500 m, the frozen ground can reach depths >1–1.5 m and persist between 6 and 9 months, and even longer in sheltered environments (chapters “[The Iberian Peninsula](#)”–“[The Mediterranean Islands](#)”).

Snow cover constitutes a key element controlling periglacial activity in mid-latitude mountain environments (Zhang 2005). Its duration and thickness have a direct effect on ground temperatures as well as on soil moisture conditions. Generally, a thicker snow cover isolates the ground from air temperature oscillations and determines rather stable ground temperatures at negative values close to ~0 °C, favouring also a delayed thawing of the ground; by contrast, years with a thinner snow cover favour a deeper seasonal frost layer with colder ground temperatures, although these values tend to increase rapidly as we approach the warm season of the year (Oliva et al. 2014a). These contrasted patterns are even more pronounced in southern Europe, where Mediterranean climates determine low precipitation—albeit with high inter-annual and intra-annual variability—together with relatively higher temperatures and high insolation. Here, the role of snow is crucial for ground temperatures and thus for periglacial processes, such as it happens in the semiarid mountain range of the Sierra Nevada, southern Spain (Fig. 3) (Oliva et al. 2014b; Gómez-Ortiz et al. 2019).

Snow melting provides water during late spring and summer, when the ground subsurface can still be frozen. This moisture supply is important for the vegetation cover, which in turn affect the activity of periglacial slope dynamics, such as cryoturbation or solifluction processes (Ulfstedt 1993). The degree of vegetation cover and moisture availability are key factors controlling soil development in the periglacial belt of mid-latitude mountains, particularly in southern Europe. Well-developed soils favour geomorphic stability, whereas areas where soil processes are weak enhance periglacial activity, such as it usually happens in the upper periglacial belt (Oliva et al. 2009). Other factors such as the lithology and topography play also a decisive role on the ground thermal regime. Certain lithologies, such as volcanic coarse-grained soils, have a major influence on the freezing and thawing of the ground depending on their degree of saturation (Ishikawa et al. 2016). This is clearly seen in Iceland, where seasonal frost prevails up to 800–900 m shifting to permafrost conditions at higher elevations (Etzelmüller et al. 2007). The texture of the sediments is also a key element as shown by rock glaciers, where the openwork coarse blocks favour the presence of cold air trapped within inter-clast voids and thus the existence (and preservation) of permafrost beneath (Jones et al. 2019). Besides, mineralogical composition of sediments or regolith also significantly affects thermal conductivity, which is highly

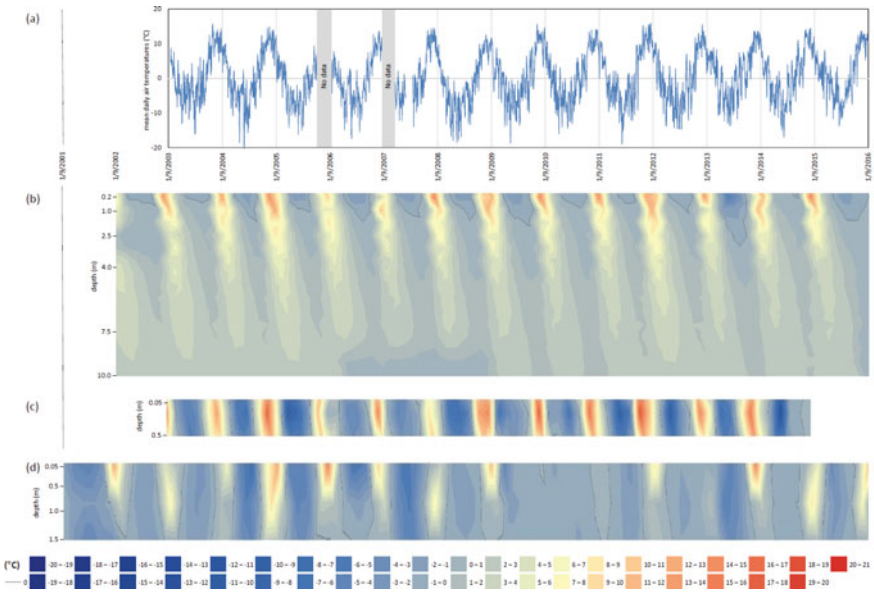


Fig. 3 Ground thermal regime in several periglacial landforms in the Sierra Nevada: sorted circle at the Machos summit plateau at 3300 m (c), rock glacier of the Veleta cirque floor at 3100 m (d), and bedrock temperatures at the Veleta Peak at 3380 m (b) and evolution of air temperatures at the Veleta peak at 3398 m. Figure published in Gómez-Ortiz et al. (2019). The ground generally freezes in autumn and thaws in late spring. The high interannual variability of the seasonal frost thickness (ranging between 0.6 to 2 m) is determined by the presence/absence, thickness and duration of snow in the ground

variable for siliciclastic and carbonaceous material (Hrbáček et al. 2017). Topographical conditions create a wide spectrum of microclimate regimes within the mountain systems, determining substantially different patterns of radiation in northern and southern slopes that have a critical importance for ground thermal regime (Gruber and Haeberli 2007). In Europe, as in the rest of the Northern Hemisphere, this is reflected in colder conditions in north-exposed slopes than in those facing south. As a result, for a similar altitude in the Alps, Pyrenees or Scandinavian Alps, we may have permafrost in the northern slope and a deep and long seasonal frost regime in the southern side (Fig. 2). In addition, the aspect has major implications for periglacial dynamics as it conditions, among others, the intensity of frost shattering, persistence of long-lying snow-patches and the vegetation cover.

Nowadays, the altitude of the lower limit of alpine permafrost in Europe generally increases towards the east and from north to south (Fig. 2). In the mountain ranges surrounding the Mediterranean Sea, permafrost is rare <2500 m, turns discontinuous between 2,500 and 2,800/3,000 m and becomes continuous in ice-free areas above this elevation. In the Pyrenees, the lower boundary of permafrost is set at 2,650 m, and its occurrence is possible above this level and probable >2,900 m, always lower in northern aspects (Serrano et al. 2019a). This is slightly higher than in the Southern

Alps (>2,400 m; Bodin et al. 2009), Rila Mountain (>2,350 m; Dobiński 2005) but lower than in the Mount Olympus (>2,700 m; Dobiński 2005), or central Anatolia and NE Turkey (>2,800–3,400 m; Gorbunov 2012). In more southern (and higher) mountains exceeding 3,000 m, such as Sierra Nevada, no permafrost belt is detected—neither in the Atlas range, in northern Africa—although isolated permafrost patches are located in the highest northern cirques, in areas that hosted glaciers during the LIA (Oliva et al. 2016a; Gómez-Ortiz et al. 2019). Indeed, microtopoclimatic conditions, such as a reduced snow cover in winter, can promote the existence of permafrost patches at relatively low altitudes of the Central Apennines on La Majella and M. Velino massifs (~2,400 m), as well as the lithological characteristics of the ground, such as volcanic rocks, that allow the existence of permafrost in the highest active European volcano (Mt. Etna) at elevations (>2,900 m; Oliva et al. 2018a, b).

In Central Europe north of 45 °N, as MAAT decrease, the alpine permafrost belt is found at lower elevations always dependent on the relief configuration and aspect. This is the case of the Alps, where active rock glaciers, borehole monitoring and extensive geophysical surveying point to a lower boundary ranging from 2,200 to 2,600 m (chapter “The Alps”). In line with this, permafrost is also found at lower altitudes in the South Carpathians (>1,900 m; chapter “The Carpathians”), the Eastern Carpathians (>1,800 m; chapter “The Carpathians”), or the Tatra Mountains in Western Carpathians (>1,930 m; Dobiński 2005). In northern Europe, permafrost is only found in Scandinavia and Iceland, or High Arctic Archipelagos such as Svalbard. In the Scandinavia case, permafrost appears between 1750 and 1050 m (Gisnås et al. 2017), most of which is either discontinuous or sporadic, being continuous only about the 2%. However, in northern and central Iceland, it can be found at lower elevations, down to 800 m (Etzelmüller et al. 2007; Czekirka et al. 2019).

Permafrost temperatures in Europe range between -1 to -5 °C (Fig. 4). At 2,900 m in the Pyrenees, permafrost temperatures are of ca. -1 °C (Serrano et al. 2019a), whereas in the Alps they range between -2 and -3 °C at 3,300–3,400 m, and between 0 and -1.5 °C at 2,600–2,950 m (Bodin et al. 2015; PERMOS 2021). In Scandinavia, permafrost temperatures are close to 0 °C on bedrock at elevations below 1,000 m, as well as in the northern part of the peninsula at altitudes of ~300–500 m; in the mountains, however, at 1,500–1,800 m, permafrost temperatures are of ~ -2.5 °C (Christiansen et al. 2010). In Svalbard, cold permafrost is widespread, with values of ~ -1.2 to -5.1 °C (Christiansen et al. 2019). The active layer is highly variable and depends on the nature of the surface. Bedrock sites tend to show much deeper active layer thicknesses than permafrost environments composed of unconsolidated sediments (e.g. active rock glaciers). In recent years, boreholes have provided a wide range of active layer thicknesses, oscillating from 2.8 to 11 m in the Swiss Alps (PERMOS 2021) to 0.6–4.6 m in Svalbard (Christiansen et al. 2019). In all areas, permafrost temperatures record a warming trend, which oscillated between 1.5 °C on Svalbard and 0.4 °C in the Alps between the late 1990s and 2018, accompanied by a deepening of the active layer, which thickened between ~10% and 200% (Etzelmüller et al. 2020).

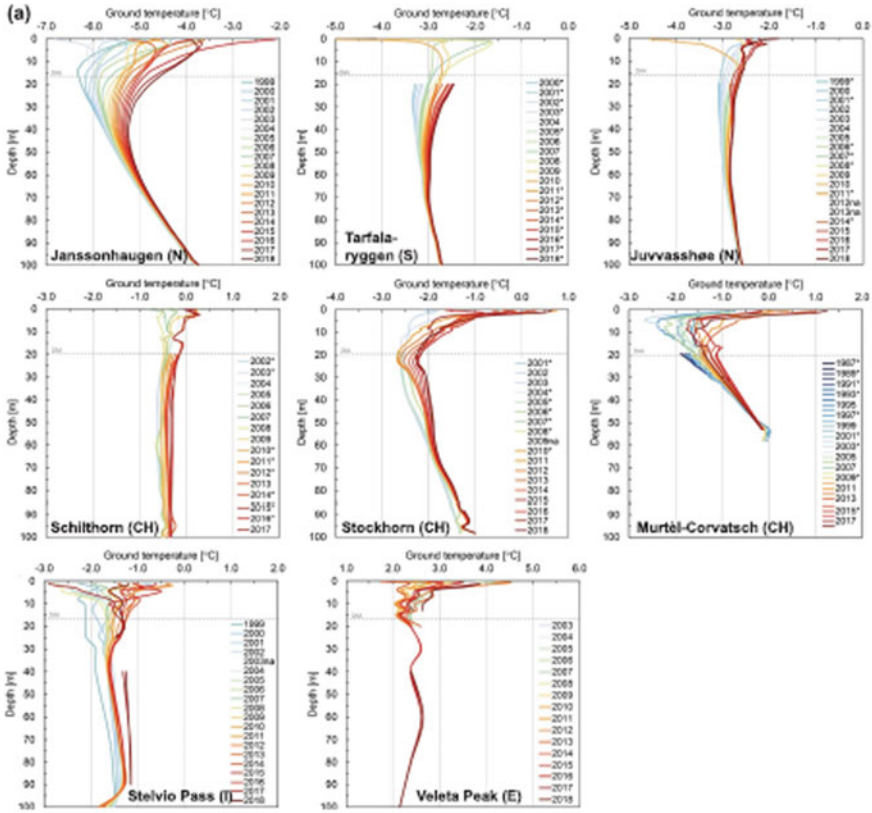


Fig. 4 Mean annual ground temperature profiles from 1987 to 2018 at different sites along a transect from Sierra Nevada in Spain (37 °N) to Svalbard (78 °N) established during the PACE (Permafrost and Climate in Europe) project (figure published in Etzelmüller et al. 2020). The figure shows how ground temperatures have warmed across all the permafrost sites over the last decades

3 Present-Day Periglacial Processes

Present-day activity of periglacial landforms in Europe is conditioned by the distribution of seasonal frost and permafrost conditions. In the coldest regions where permafrost is widespread, periglacial processes are generally active and most associated phenomena show a dynamic pattern under current climate regimes. This is the case of ice-free regions in high-latitude regions (e.g. Svalbard, northern Scandinavia and northern Iceland) as well as in the highest mountain ranges (e.g. Alps, Pyrenees, Anatolia mountains) that include a broad spectrum of active periglacial landforms and processes, including widespread thermal-contraction-crack polygons (e.g. ice and sand wedges), permafrost creeping landforms (e.g. rock glaciers, protalus lobes) and other hillslope phenomena (e.g. solifluction lobes, debris flows), cryoturbation features (e.g. sorted circles, stone stripes), or thermokarst phenomena

(e.g. retrogressive-thaw-slumps, lakes), amongst others. However, some permafrost-related features (i.e. pingos) may undergo a process of collapse by hydrostatic and topographic adjustment as part of their natural evolution (French 2007).

However, in seasonal frost environments, active periglacial landforms coexist with others that are inactive or weakly active as present-day climate conditions are not favourable for maintaining periglacial activity. This may be the case of periglacial features that formed under colder climates prevailing during Quaternary glacial periods, which favoured the expansion of permafrost to lower latitudes/altitudes. For example, most rock glaciers in Mediterranean mountains are currently inactive and developed following Late Glacial cold phases (i.e. Bølling–Allerød and Early Holocene; Palacios et al. 2017a, b, 2022; Andrés et al. 2018; Oliva et al. 2018a, b). These colder conditions also favoured the decline of the periglacial belt associated with seasonal frost conditions to the mid-low slopes of the high massifs as well as the occurrence of seasonal frost in moderately high massifs that are currently not affected by cryogenic processes. Stratified scree deposits or talus slopes currently covered by vegetation show evidence of this fact in many mountains in southern and central Europe (Oliva et al. 2018b).

Hence, whereas in southern Europe the periglacial domain is only found in the highest mountains where no settlements and little infrastructure exists (Oliva et al. 2016b), in the high latitudes of Europe a large number of socio-economic activities take place in active periglacial terrain. Here, a deep comprehension of periglacial processes is needed for a proper assessment of landscape management, infrastructures potentially affected by ground ice dynamics as well as reduce the potential risks of societies to natural hazards.

Together with ground thermal regime, the prevailing lithology in periglacial landscapes constitutes also key factor controlling also their degree of activity (French 2007). Whereas some processes are widespread and show a greater development in certain lithologies, such as solifluction dynamics in metamorphic and limestone massifs, others are critically constrained by the bedrock type, such as rock glaciers, which are almost absent in limestone massifs, such as the Picos de Europa in Iberia, or in the Alps. In addition, topography also determines contrasting degrees of activity within the regional area, particularly in mountain areas. Whereas northern slopes tend to show greater periglacial dynamics than southern sites in non-glaciated massifs of southern and central Europe (Fig. 2). In glaciated environments in northern Europe and south-facing slopes of high-mountain regions can be affected by intense periglacial dynamics. However, this pattern is highly influenced by microclimatic conditions, the presence of late-lying snow patches, past environmental conditions, etc.

3.1 Southern Europe

In southern Europe, the active periglacial domain resembles the glaciated domain of the last glacial cycle: The elevation difference between the lower Last Glacial

Maximum (LGM) periglacial zone and present-day periglacial processes is about 800–1000 m (Oliva et al. 2016b). At latitudes below ~45 °N in the European continent, periglacial processes and landforms are mostly concentrated in the main mountain ranges, with minor evidence in mid- and low-mountainous regions as well as in interior plateaus and basins. Many periglacial landforms and deposits are located in environments with relatively high modern MAATs, and are thus inactive or very weakly active. This is the case of solifluction landforms in the Iberian mountains, which are generally relict and inherited from past colder and wetter periods and only show activity above 2,500 m in the Pyrenees to 3,000 m in the Sierra Nevada, where MAAT are <2 °C (chapter “[The Iberian Peninsula](#)”). A wide range of active periglacial phenomena are found across the periglacial belt driven by seasonal frost, such as features associated with cryoturbation (e.g. patterned ground, garlands, earth hummocks), rock weathering (e.g. blockfields, tors), talus slopes and related processes (talus cones, debris flows, or relict rock glaciers), periglacial mass movements (solifluction landforms, landslides) and others that are not so common and associated with certain lithologies (e.g. ice caves) (Fig. 5). In addition, there are also several active morphologies within the marginal areas underlined by permafrost; rock glaciers and protalus lobes are amongst the most abundant, although other features such as block streams and frost mounds have been also reported from high mountain ranges of Southern Europe (chapters “[The Iberian Peninsula](#)”–“[The Mediterranean Islands](#)”).

Monitoring of active periglacial features is, however, very limited and has been mostly done in some of the highest Iberian mountains. This is the case of solifluction, which has been shown to be inactive below 2,800 m in the Sierra Nevada, and only the landforms located at ~3,000 m near late-lying snow patches reported annual displacement rates <1 cm yr⁻¹ (Oliva et al. 2009, 2014b). Other studies in the Pyrenees and the Picos de Europa showed higher movement rates of solifluction features up to 1.8–3.6 cm yr⁻¹ (chapter “[The Iberian Peninsula](#)”). Several rock glaciers have been monitored in the Pyrenees since the early 1990s showing highly variable annual horizontal displacement rates between ~10 and 40 cm yr⁻¹ (Chueca and Julián 2011; Serrano et al. 2019a), with lower rates of 4–11 cm yr⁻¹ in the case of an active protalus lobe in the Maladeta massif (González-García et al. 2017). Some rock glaciers showed evidence of undergoing a rapid degradation of underlying buried ice and permafrost that resulted in higher vertical than horizontal displacements (4 vs 1.3 m during 2001–2016; Gómez-Ortiz et al. 2019). The recent evolution of periglacial talus cones has also been surveyed at elevations of 1900–2200 m in the Cantabrian Mountains, where very active dynamics resulted in surface changes up to 0.5–1 m yr⁻¹ (Serrano et al. 2019b).

3.2 Central Europe

In Central Europe, the present active periglacial domain resembles the highest mountain ranges, i.e. the Alps, Carpathians, or the highest Variscan ones such as

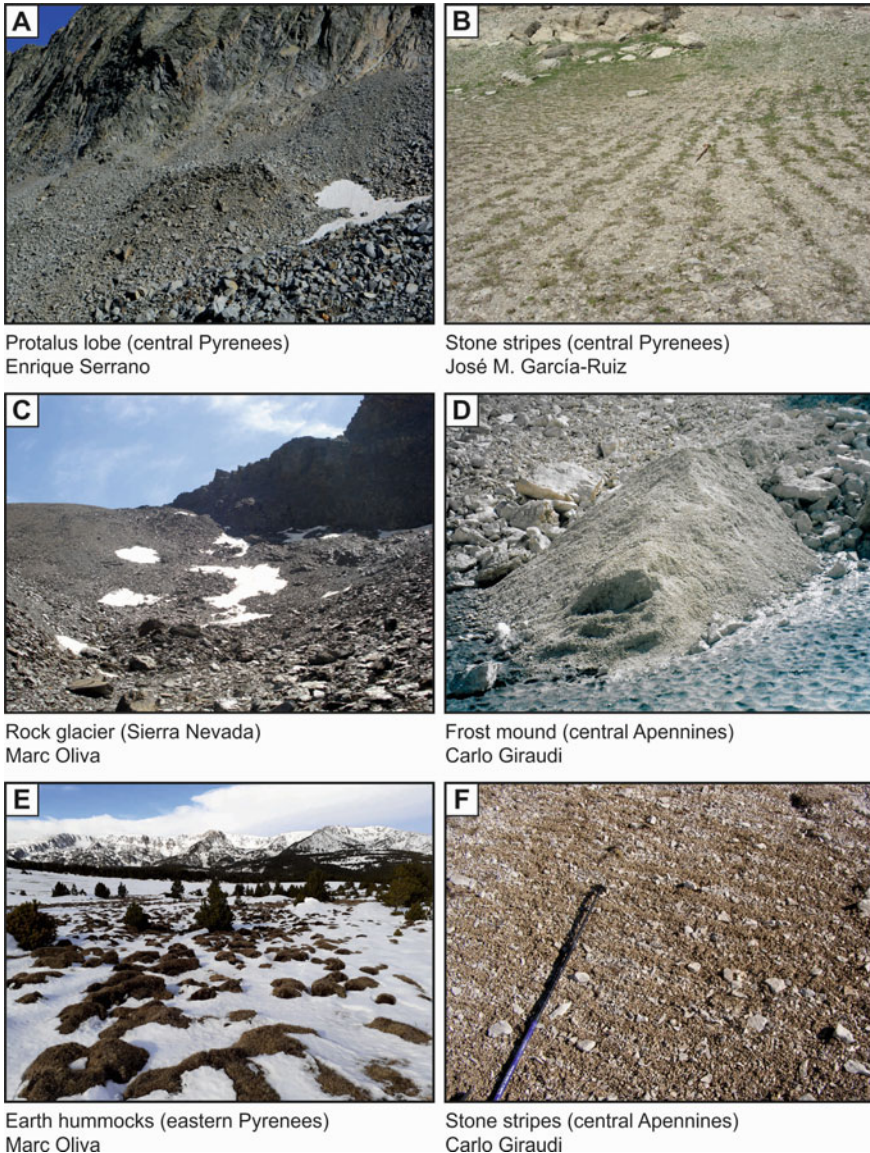


Fig. 5 Examples of active periglacial features in southern Europe

Karkonosze, Bohemian Forest, or Hrubý Jeseník Mts. A significant longitudinal gradient connected with the continentality of the climate could be seen for mountain ranges along $\sim 50^\circ\text{N}$. The lowermost limit of active periglacial processes in the Sudetes lies around 1400 m (chapter “[The Central European Variscan Ranges](#)”), however this altitude rises to 1500–1600 m in the Western and Eastern Carpathians

further to the east (chapter “[The Carpathians](#)”). Significantly higher is the location of the current periglacial realm in the Alps: for the Northern-Eastern Alps it starts at altitudes above 1950 m, but in Southern-Western Alps this altitude lies at least at 2350 m (chapter “[The Alps](#)”). The areas of active periglacial processes in Central Europe were not always glaciated during the last glacial period, as these mountain ranges were mostly covered by valley glaciers during the Pleistocene cold periods and periglacial domain existed also on summits of younger Alpine belts of the Carpathians, or flat unglaciated surfaces of older Variscan ranges. In such location periglacial landforms survived till present and the best evolved periglacial landscape could be found there. This was different for much higher European Alps, where 55% of the area was glaciated during the LGM with the largest glacier coverage in Northern-Western Alps (chapter “[The Alps](#)”) and non-glaciated summit areas were, therefore, spatially restricted. While large areas of the Alps and Carpathians have negative MAAT with the 0 °C isotherm being located at ~2200–2600 m and ~1760–1885 m in the Alps and Carpathians, respectively, the summit areas of Central European Variscan ranges show at present only positive MAAT, although the MAAT was negative at the highest parts of the Karkonosze Mts. even during the LIA (Sobik et al. 2019). The large difference of the altitude of the 0 °C isotherm throughout the Alps reflects rather the north–south elevation gradient with higher altitudes generally for Southern Alps. This is also supported by diverse precipitation amounts, being more than double in Northern Alps compared to Southern Alps.

The Alps and Carpathians represent a classical area of European periglacial landforms research as a very wide spectrum of different active periglacial features induced by frost weathering, cryoturbation, regelation sorting, periglacial slope movement, cryoplanation, nivation, or aeolian and thermokarst processes can be found there (chapters “[The Alps](#)”, “[The Carpathians](#)”). Rather different features induced by periglacial processes could be found in carbonate and igneous rocks of both main Central European mountain ranges, where active periglacial processes occur. On the contrary, the only characteristic periglacial landforms in significantly lower Central European Variscan ranges, which are currently considered as a marginal periglacial domain, are regelation-related sorting features (chapter “[The Central European Variscan Ranges](#)”).

Permafrost monitoring has one of the longest European tradition in the Alps, which is since 1990s transnational and collaborative. However, long term monitoring of rock glaciers started in Switzerland and Austria already in 1920s (Chaix 1923; Finsterwalder, 1928) and many other monitoring activities emerged in 1950s–1980s (e.g.; Pillewizer 1957; Vietoris 1972; Francou and Reynaud 1992). Most of the rock glaciers in the Alps move annually in decimetres to metres (chapter “[The Alps](#)”), but some rock glaciers show an accelerating trend with extraordinary velocities of up to 100 m a⁻¹, probably connected with rapid degradation of permafrost (Delaloye et al. 2010; Eriksen et al. 2018). The slower current periglacial processes are monitored at different places of the Alps and the Carpathians. Present-day solifluction movement rates are 1–20 cm a⁻¹ in the Alps (Matsuoka 2001; Veit 2002; Kellerer-Pirklbauer 2018) and from <1 cm a⁻¹ to >15 cm a⁻¹ in the Carpathians (Urdea 2000; Rączkowska

2007; Onaca et al. 2017b) showing highly variable displacement rates depending on microtopographic and microclimatological conditions (Fig. 6).

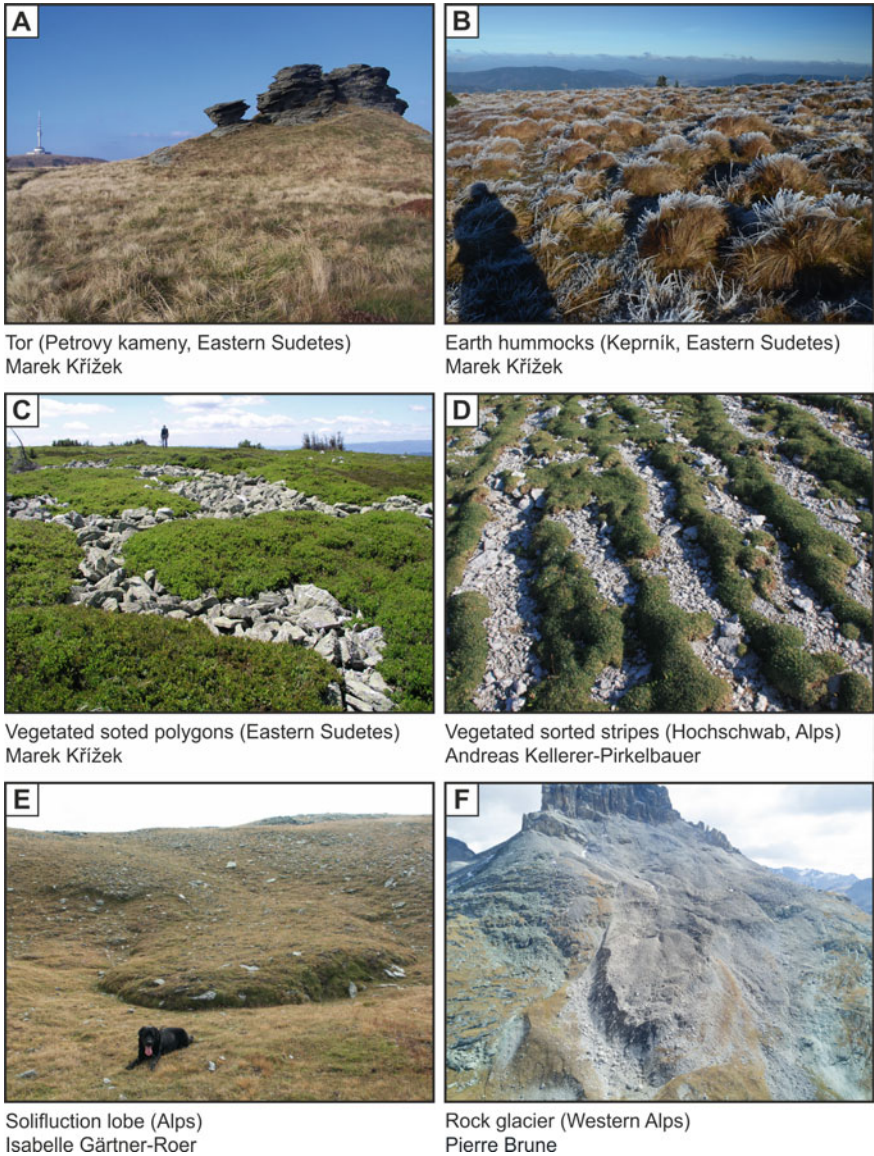


Fig. 6 Examples of active periglacial features in central Europe

3.3 Northern Europe

North European regions were extensively glaciated and shaped by extensive cold-based ice sheets during successive glacial cycles. Thus, active and relict landforms often intercalate with those originated by glaciers, and in some cases, they even survived the passage of the ice. Currently, MAAT are below the freezing point at the highest peaks of the Scandes (<-6 °C) and the Icelandic highlands (-4 °C). However, in Great Britain, the MAAT does not reach the negative values even at the highest Scottish summits ($1-3$ °C). The zonation of the periglacial landforms is problematic in these two areas as active and inactive landforms coexist. Nowadays, periglacial activity is limited to landforms and processes associated to mass movements (solifluction, complex landslides, rock slope failures and debris flows linked to permafrost degradation, and avalanches, with permafrost degradation as a major trigger), cryoturbations (earth hummocks / thúfur, small-scale patterned ground), talus slopes (talus cones, debris flows) and ice-cored formations (rock glaciers, palsas and ice-cored moraines).

Monitoring of the active periglacial features is restricted to a reduced number of landforms. Solifluction lobes have been monitored in the Great Britain, Scandinavia and Iceland. In the first case, studies were carried out in the highest areas of northern Scotland (900 m) for over than 30 years with average rates between 7.8 and 10.6 mm a^{-1} (Ballantyne 2013). In Scandinavia, solifluction lobes have been studied in a number of locations, and the measurement of their displacement velocities revealed a clear relationship with the atmospheric warming (Åkerman and Johansson 2008; Ridefelt et al. 2009; Strand et al. 2020). Douglas and Harrison (1996) measured displacement rates of solifluction lobes in south-eastern Iceland reaching up to 11 mm a^{-1} . Active landslides have also been monitored in Scandinavia, revealing a clear acceleration of their sliding, linked to the degradation of permafrost (Hilger et al. 2021). Rates up to 0.57 and 0.7 m a^{-1} were measured in active landslides of northern Iceland (Wangensteen et al. 2006; Sæmundsson and Petursson 2018). Active rock glaciers are showing evidence of degradation in northern Iceland, with surface subsidence and an almost static state (velocities between 0.29 and 0.74 m a^{-1} ; Wangensteen et al. 2006; Tanarro et al. 2019). In Scandinavia, however, the recent trend of rock glaciers was characterised by an acceleration of their displacement, linked to the permafrost degradation (Eriksen et al. 2018) (Fig. 7).

4 The Legacy of the Past

Quaternary climate variability determined large-scale variations of the cold-climate geomorphological processes affecting the European continent. The surface extent of glacial and periglacial dynamics expanded to lower latitudes and elevations during colder (glacial) periods and migrated to the higher latitudes and the upper parts of the mountain ranges during warmer (interglacial) periods. Colder climate conditions

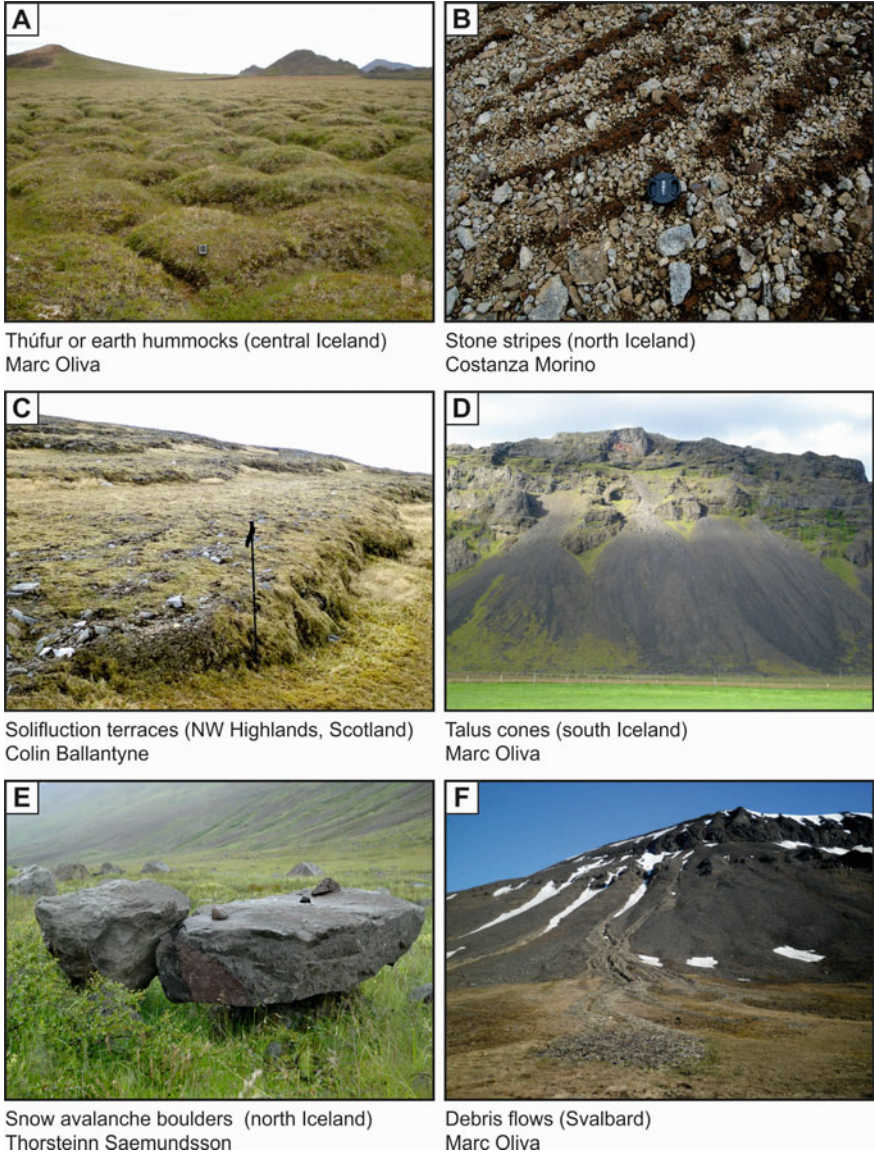


Fig. 7 Examples of active periglacial features in northern Europe

prevailing during glacial periods promoted the inception of the European Ice Sheet Complex that reached latitudes 48 °N (Palacios et al. 2022) as well as a large expansion of glaciers in the currently still glaciated mountain ranges and the formation of numerous glaciers in currently ice-free massifs (Woodward 2009; Allard et al. 2021). By contrast, warmer conditions during interglacial phases—such as the current one,

the Holocene—determined the final melting or substantial retreat of glaciers as well as the shift of permafrost and periglacial processes to higher elevations and northerly latitudes (Oliva et al. 2018b) (Fig. 8).

The timing of the different glaciations and the surface extent occupied by the ice masses across the European continent has been recently summarised in a book that complements the present one focusing on the periglacial dynamics (Palacios et al. 2022). In fact, research on past glacial processes in Europe has received much more attention from scholars than studies on the impact that past periglacial processes had on the current landscape of the continent. Indeed, the timing of development of periglacial landforms is still little known and has been mostly inferred using chronostratigraphic approaches and relative dating techniques.

Reconstructions of past periglacial dynamics have been done using different methods depending on the nature of periglacial deposits. Radiocarbon dating has been the most widespread method to infer the past activity of periglacial features

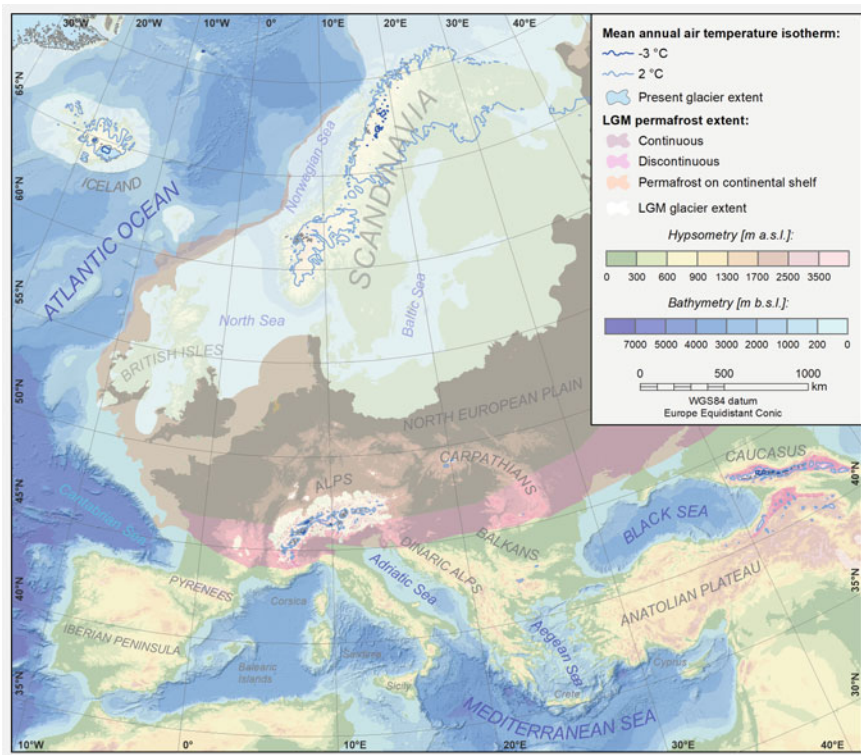


Fig. 8 Map of Europe showing the distribution of continuous, discontinuous permafrost, permafrost on continental shelf and glaciers during the LGM (Lindgren et al. 2016) versus current Mean annual air temperatures (MAAT) isotherms of 2 °C (defining current periglacial realm) and –3 °C (defining current continuous permafrost) and current glacier extent

as it allowed dating organic remnants trapped by periglacial processes within sedimentary sequences; it has been used to date stratified screes (Ruiz-Fernández et al. 2019), solifluction beds (Oliva et al. 2011), ice-wedges (Opel et al. 2018) or pingos (Yoshikawa and Nakamura 1996) or aeolian loess deposits (Oliva et al. 2014c). More recently, research studies on periglacial deposits promoted the cross-comparison of different (absolute) dating techniques, such as radiocarbon and luminescence dating (Palstra et al. 2021), as well as the implementation of cosmic-ray exposure (CRE) dating on some periglacial features whose chronology was only relative. This is the case of CRE on relict rock glaciers that are reporting ages of development following their deglaciation, and revealing thus a paraglacial origin in many features from both south (Palacios et al. 2017b; Andrés et al. 2018) and north Europe (Andrés et al. 2019; Fernández-Fernández et al. 2020). In addition, relative dating techniques, such as Schmidt-hammer exposure-age dating, have been recently used to establish absolute ages for nival landforms, such as the formation of pronival ramparts by frequent snow-avalanche (Matthews and Wilson 2015). However, the age of development of some periglacial landforms, such as blockfields or features associated pattern ground features such as sorted circles, remains still uninvestigated.

The wide variety of periglacial landforms and deposits that are no longer active under present-day climate conditions include a combination of features that formed during the Late Quaternary under much colder periods than today to others developed during the deglaciation as well as other minor features that generated during the last historical cold phase, the Little Ice Age (LIA). Logically, the degree of preservation of periglacial landforms is generally better for recent periods than for longer timescales as erosion may have degraded these features and make more challenging the reconstruction of past landscape dynamics. In this sense, the preservation of landforms tends to be greater in granitic and metamorphic massifs and lesser in limestone lithologies where karst dissolution has reshaped periglacial landforms inherited from past periods.

4.1 Last Glacial Cycle

From the end of the Eemian Interglacial (~114 ka) till the onset of the Holocene (11.7 ka), Europe underwent the last massive cooling and associated glaciation, with some periods of warmer conditions (interstadials) and others with significantly colder temperatures (stadials) (chapter “[Periglacial Landforms Across Europe](#)”). Colder conditions favoured the formation of mountain glaciers and a large European Ice Sheet in northern Europe advancing South to latitudes 50 °N (Stroeve et al. 2016; Batchelor et al. 2019), together with an expansion of winter sea ice to latitudes surrounding the southern fringe of the European Ice Sheet (Denton et al. 2010). These conditions favoured significantly different oceanographic and atmospheric circulation patterns than present-day, which in turn affected spatial and temporal precipitation regimes. The combination of both temperatures and moisture regimes during

the Last Glacial Cycle favoured the expansion of cold-climate geomorphological processes, including glacial and periglacial dynamics.

Southern Europe was not extensively glaciated and glaciers remained confined in mountain regions as small ice fields and alpine glaciers descending through the valleys. The depression of the Equilibrium Line Altitude (ELA) in Mediterranean mountains during the LGM—compared to present-day values—ranged between 800 and 1,200 m, which suggests a minimum temperature difference of ca. 6–10 °C (Oliva et al. 2018b). The ELAs in glaciated mountains in southern Europe increased towards the East: whereas in NW Iberia the ELA was located <1,000 m, it increased to >1,500–2,000 m in the western and central Mediterranean Basin (Hughes et al. 2006), and reached elevations of 2,500–3,000 m in mountains of Turkey and Lebanon (Messerli 1967).

The northern fringe of the Mediterranean regions included areas with permafrost occurrence during the Last Glacial Cycle, such as central France and the northern Balkan region (Brown et al. 2001; Vandenberghe et al. 2014). The cold and arid conditions prevailing in these permafrost regions during the coldest phases of the last glaciation, such as the LGM (chapter “[Quaternary Climate Variability and Periglacial Dynamics](#)”), favoured a scarce vegetation cover and an intense aeolian activity, which explains the thick loess deposits existing in central-northern Europe (Zeeden et al. 2018; Chapter 11) and northern Italy (Cremaschi et al. 2015). In some internal depressions across the Mediterranean region, evidence of the existence of ice wedges and cryoturbation features has been associated with deep seasonal frost conditions or even isolated patches of sporadic permafrost, such as in the Spanish Meseta (chapter “[The Iberian Peninsula](#)”), or in the Pannonian basin (chapter “[The Italian Peninsula](#)”). Loess deposits including cryoturbation evidence may reveal also the existence of permafrost conditions in areas surrounding the valley glaciers in the southern Alps, depressions at the foot of the northern Apennines and even at the Po floodplain (Cremaschi et al. 2015).

At higher elevations in the mountain regions of southern Europe, below the glaciated domain was the periglacial belt conditioned by the occurrence of permafrost at moderately high elevations and seasonal frost at the lower parts of the mountains high-altitude plateaus (e.g. at 600–1,200 m). Due to topographical conditions, permafrost was also widespread in non-glaciated environments above the snowline. This is confirmed by the existence of permafrost-related landforms that are inactive under current climate regime (Fig. 9), such as metre-sized stone circles developed in relatively flat summit surfaces where the intense wind did not favour snow and ice accumulation (Oliva et al. 2018b). Present-day analogues of these patterned ground features exist in polar regions, where permafrost is widespread with MAAT \sim –6 °C (French 2007). It is thus likely to associate the formation of these present-day inactive landforms to the occurrence of permafrost. At the summit level, in palaeonunataks protruding the former glaciers, permafrost must have been widespread. In these areas, under very intense periglacial conditions, large block fields and tors also developed in favourable lithological settings. Other features typically associated with permafrost landscapes are those developed in glacial-free slopes, such as rock glaciers, which also formed during the Last Glacial Cycle at a broad range of elevations (\sim 1,000 to

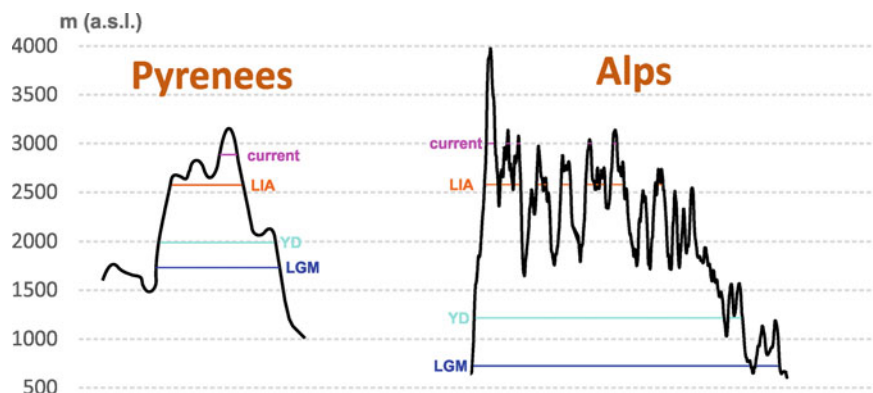


Fig. 9 Altitudinal changes of permafrost distribution during different periods of the last climate cycle in the Pyrenees and Alps

2,500 m; Fig. 9), or block streams—some several km long and hundreds of metres wide—distributed between 700 and 1800 m of altitude.

At lower elevations, seasonal frost was widespread and affected the low-mid mountain slopes <1,000–1,500 m in most mountain ranges, including those close to the Mediterranean coast. In Iberia, some stratified periglacial deposits have been dated of pre-LGM age in NW mountains, whereas others coincided with the LGM in the Central Pyrenees (chapter “[The Iberian Peninsula](#)”). In these areas, the existence of stratified slope deposits (or *grèzes litées*) has been used to reconstruct the occurrence of periglacial conditions during the Last Glacial Cycle (Fig. 10). In addition, the lowlands in more continental areas (e.g. at 300–800 m) were also affected by intense seasonal frost dynamics, such as in Central Iberia (chapter “[The Iberian Peninsula](#)”), or the Balkan region (chapter “[The Balkans \(Without Carpathians\)](#)”), although geomorphic evidence of that period in these areas is scarce and mostly documented based on the existence of stratified screes.

The North European Plain building the northern part of Central Europe was repeatedly covered by the southern margin of the European Ice Sheet (Batchelor et al. 2019; Palacios et al. 2022), with the most extensive advances reaching the northern forelands of Central European Variscan Ranges and Northern Carpathians during the Elsterian, MIS 12 (Nývlt et al. 2011). However, the largest advance took place during the Donian (MIS 16) and Saalian (MIS 6) glaciations in the Eastern European Plain, where the ice sheet front crossed by extensive frontal lobes the latitude of 50°N and advanced to Southern Russia and Ukraine, respectively (Ehlers et al. 2011). However, the European Ice Sheet was of a limited extent during the LGM (MIS 2) with its maximum extent occurring during the Leszno Phase, dated at 20.7 ± 0.8 ka (Tylmann et al. 2019; Marks et al. 2022) near the current state boundary of Germany and Poland and vast areas of Central Europe were glacier-free apart from mountain glaciers. The most extensive mountain glaciation in Central Europe covered during the LGM the Alps, where ~55% of the mountain system was glacierised (Ivy-Ochs

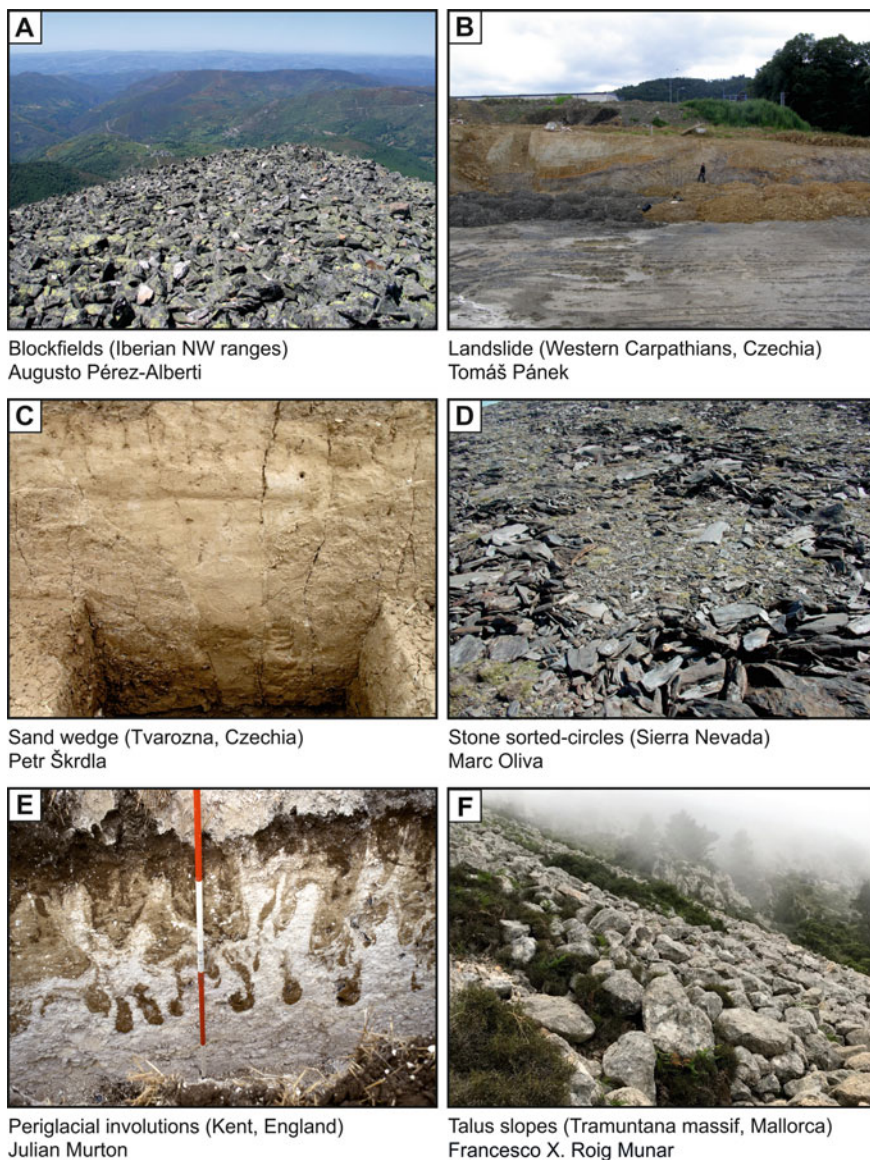


Fig. 10 Examples of periglacial features formed during the Last Glacial Cycle in Europe

et al. 2022; Chapter 8). Mountain valley glaciers were also present in Carpathians (Urdea et al. 2022; Zasadni et al. 2022), or in the highest parts of Central European Variscan Ranges (Mentlík et al. 2013; Engel et al. 2014; Hofmann et al. 2020). The vast glacier-free corridor between the European Ice Sheet in the north and the Alps ice cap in the south (with only minor mountain glaciers) represented a periglacial zone

with negative MAAT and highly dynamic periglacial processes during the LGM. The area was underlain by a continuous permafrost (Vandenberghe et al. 2014; Lindgren et al. 2016) and numerous periglacial features indicative of its presence (e.g. ice wedge pseudomorphs, relic sand wedges). Due to gradual long-term cooling leading to peak (pleni)glacial conditions, the permafrost thickness in low-lying areas of Central Europe could reach 200–250 m, according to cryogenic deformations of the rocks detected in deep boreholes and model calculations (Šafanda and Kubík 1992; Czudek 2005). This was also the reason why the North European Plain and adjoining mountain ranges became a classical area of the study of past periglacial features since the early twentieth century, as is well described in chapters “[The Central European Variscan Ranges](#)”–“[The North European Plain](#)”. The periglacial features are mostly connected with their origin during Pleistocene cold phases and vast majority of them would stem from the Last Glacial.

The MAAT dropped by 6 to 12 °C during the LGM compared to current interglacial temperatures (Heyman et al. 2013; Pačes et al. 2017). In the non-mountain areas, MAAT ranged between –7 and –2 °C (Šafanda and Rajver 2001). Peak pleniglacial conditions also represented the driest periods of the Quaternary climatic cycles, as mean annual precipitation values in Central Europe declined by 50–60% and were estimated to be from 250 to 300 mm in lower lying areas (Hatté and Guiot 2005). Low temperatures and decreased precipitation amounts led to scarce vegetation cover, which persisted only in low-lying areas and especially along the south-facing slopes of rivers (Feurdean et al. 2014; Magyari et al. 2014). These climatic conditions favoured periglacial dynamics not only in mountain ranges of Central Europe, such as the Alps, and Carpathians, but also in lowlands of the North European Plain. Detailed descriptions of Last Glacial periglacial landforms and their time frame of formation is given in chapters “[The Alps](#)”–“[The North European Plain](#)”. For example, hillslope activity and accumulation of periglacial slope deposits during the LGM and slightly after was proven in some of the Central European Variscan ranges (Mt. Ślęza with ages from 17.1 ± 1.1 to 19.5 ± 1.3 ka; Waroszewski et al. 2020). However, even pre-LGM periglacial landsliding activity (MIS 3; 48 to 56 cal. ka BP) has been described and radiocarbon dated in the Carpathians (e.g.; Pánek et al. 2014; Jankovská et al. 2018). However, it should also be noted that pre-Last Glacial periglacial features could often be found in Mid-Pleistocene aeolian, glacial or fluvial deposits preserved outside the southern maximum extent of the European Ice Sheet during the LGM. Accordingly, pre-glacial tors were recently found buried by Elsterian glacial deposits at the southernmost margin of European Ice Sheet (Hanáček et al. 2021).

North European regions were extensively glaciated by European Ice Sheet (Batchelor et al. 2019; Palacios et al. 2022). In Iceland, the Pliocene–Pleistocene succession of volcanic rocks evidence the occurrence of many glaciations covering great part of the Island and the surrounding shelves. During the LGM, the Icelandic Ice Sheet (IIS) covered the entire island, reaching the shelf edge at 40 km off the coastline (chapter “[Iceland](#)”). In Great Britain and Scandinavia, the successive cold-based ice sheets allowed the preservation of periglacial landforms originated prior to the last glacial cycle. This was the case of the blockfields and tors formed in the plateau areas before the passing of the Late Devensian ice sheet (>30 ka) (Ballantyne and Stone

2015) in Great Britain. They coexisted with the formation of patterned ground during a number of episodes (Bateman et al. 2014; West 2015). During the Pleistocene, different climatic stages alternated, and during the cold ones, aeolian activity was especially intense, with the deposition of periglacial aeolian deposits such as loess and coversands (chapter “Great Britain and Ireland”). The influence of the air ice masses during the cold stages of the Pleistocene led also to significant changes of the ground thermal regime in the ice-free areas, from continuous/discontinuous permafrost to seasonally frozen ground. In these circumstances, is likely that permafrost developed under the ice sheets and at their vicinity, as deformed sediments wedges, polygonal structures (eastern England), remnants of frost mounds, thermokarst depressions and involutions have suggested (Waller et al. 2011). Conversely, in the flat lowland areas of the British Isles periglacial weathering occurred, together with river incision, formation of abundant dry valleys, fluvio-colluvial deposits and river terraces (chapter “Great Britain and Ireland”).

In Scandinavia, the situation was relatively similar. Ice sheets and smaller ice caps than those of the LGM covered successively the plateau areas, with some glacier-free parts likely being subaerially exposed to periglacial conditions during interglacials. Block fields formed under that morphoclimatic domain and were successively over-ridden by successive ice sheets whose cold-based regime allowed them to survive even to the LGM glaciation (chapter “Scandinavia”). The same occurred with tors, which benefited from shielding by the cold-based ice sheet and survived multiple glacial cycles, alike to what it must have occurred to more fragile sorted patterned ground and block streams appearing in the blockfield areas (Wilson et al. 2017). In some lowland areas, periglacial activity occurred in the form of ice-wedge casts (~290–130 ka) forming polygonal patterns that remained active until the YD (chapter “Scandinavia”).

4.2 Last Deglaciation

The deglaciation process of European Ice Sheet that started by 20.7 ka (Tylmann et al. 2019) promoted major environmental changes in Europe (Clark et al. 2009). Its shrinking towards the high latitudes was parallel to the deglaciation of most Northern Hemisphere mountain environments. Warmer climate conditions favoured also shifts in the intensity and spatial extent of periglacial processes in Europe, which became progressively confined to the highest mountain systems. Indeed, as glaciers retreated, periglacial dynamics occupied the recently glacier-free terrain with very active processes during the paraglacial phase (Ballantyne 2008).

However, the deglaciation was not a gradual process, but included an alternation of colder and warmer phases (chapter “Periglacial Landforms Across Europe”). During these major periods, the glacial evolution in European mountains followed a much more homogenous pattern than during the Last Glacial Cycle. Recent studies show interruptions during the long-term glacial retreat and even some phases of glacial re-expansion that took place in full synchrony within the Mediterranean region and with

the mountainous systems of central Europe, as well as with the European Ice Sheet. The onset of a rapid and massive retreat of this ice sheet occurred at 20.7 ka (Tylmann et al. 2019) and accelerated at 18 ka as it also occurred in most European mountains (Palacios et al. 2022). In the Alps, for example, some of the largest glaciers in the range had already lost ca. 80% of their mass by 18 ka (Ivy-Ochs 2015), whereas the Pyrenees could have been fully deglaciated at that time, as it was probably the case of most south European mountain ranges (Palacios et al. 2017a).

Within the last deglaciation, the first glacial advance of mountain glaciers and of the European Ice Sheet occurred during the Oldest Dryas (17.5–14.6 ka; Palacios et al. 2022). Glacial expansion was driven by a significant reduction of the temperature of the North Atlantic Ocean that favoured a decrease in the intensity of the Atlantic Meridional Overturning Circulation, with extreme seasonality of cold winters and mild summers (Denton et al. 2005). Glaciers advanced to elevations of only 400 m above the LGM moraines in the Alps (Ivy-Ochs 2015), but also advanced to positions close to the LGM limits in several mountains in the Iberian Peninsula (Oliva et al. 2019); substantial glacial advances were also recorded in other European mountains, such as in the Tatra Mountains, Apennines, Balkans and Anatolian mountains (see details in Palacios et al. 2022). In the Alps, MAATs have been reconstructed to be 0 °C lower than present-day, and precipitation was one-third less (Ivy-Ochs 2015). This colder climate must have favoured the expansion of periglacial conditions, including permafrost, in mountain areas. Some fossil periglacial phenomena in European mountains may have formed during this period, including blockfields, patterned ground features or rock glaciers (Fig. 11; Oliva et al. 2018a, b). However, there is a lack of absolute ages for periglacial landforms formed during this period.

Similarly, in Central Europe the development of the periglacial realm is linked to the retreat of European Ice Sheet, which had completely withdrawn from North European Plain after 16.5 ka (the age of Gardno deglaciation phase in Southern Baltic Sea sector; Tylmann and Uścińowicz 2022). The deglaciation of the highest Central European ranges (the Alps and Carpathians, as well as Central European Variscan ranges) took also place just after the local LGM, as the maximum advance of mountain glaciers took place mostly by 22 ka (Ivy-Ochs et al. 2022; Zasadni et al. 2022). Relict periglacial landforms are widespread in the Alps and Carpathians, as they are related either to glacier-free conditions and hence periglacial weathering during the LGM above the highest glacierised areas, or to periglacial landscape modification of deglaciated areas in the paraglacial phase during the Late Pleniglacial, Late Glacial and Holocene periods (chapters “The Alps”, “The Carpathians”). However, the currently available geochronological data are by far sufficient to address the timing of the periglacial landforms formation especially for the time period preceding the Younger Dryas, i.e. 18–13 ka. Besides, numerous periglacial features experienced multiple stage development during the colder phases of the Late Glacial, the Early Holocene and may have been activated also during the LIA in higher mountainous areas of Central Europe.

In the Alps, the relationship between glacier recession and formation of permafrost-related landforms was analysed by Kellerer-Pirklbauer and Kaufmann (2018). They also proposed a mass-balance approach of rock glacier evolution, as

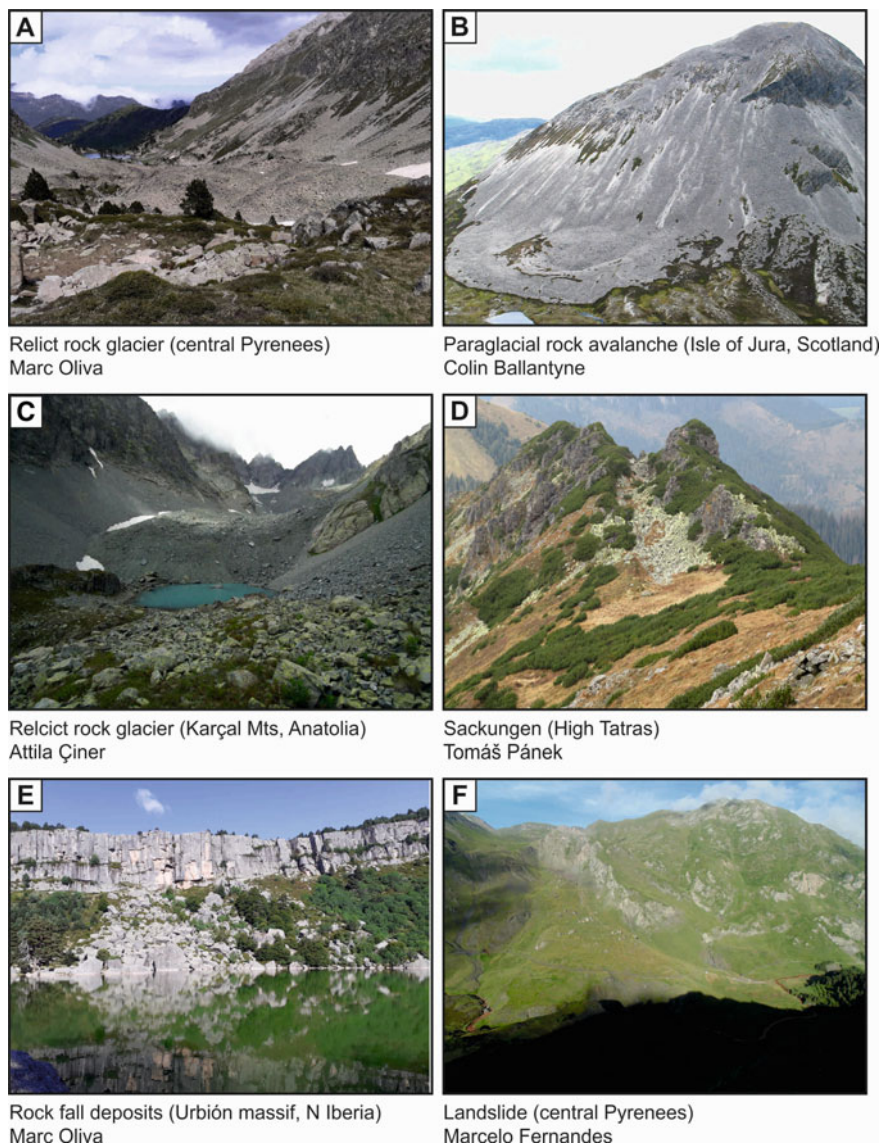


Fig. 11 Examples of periglacial features formed during the last deglaciation in Europe

most of them evolved significantly during the paraglacial phase after glacier retreat. Most of the currently relict rock glaciers of the Alps were formed shortly after, or even simultaneously with the pre-Bølling retreat of glaciers and stabilised around 15.7 ka (Steinemann et al. 2020). The beginning of the paraglacial slope activity in the high mountains of Central Europe is connected with the early deglaciation phase (starting from Older Dryas) as evidenced by CRE dating of deep-seated rock-slope

failures (sackungen) in Tatra Mts., the Western Carpathians by Pánek et al. (2017). In the Carpathians, the geochronological data on Older Dryas rock glacier activity are missing. However, Urdea (1998) was the first to claim that the formation of rock glaciers in the Southern Carpathians took place already during the Older Dryas. Besides, periglacial hillslope accumulations of Oldest Dryas and Older Dryas ages are described from Central European Variscan ranges not only based on their stratigraphic position in relation to LGM loess accumulation and Laacher See tephra unit, but also based on OSL dating (Mt. Ślęża with ages from 17.1 ± 1.1 to 19.5 ± 1.3 ka; Waroszewski et al. 2020). In the lower-lying North European Plain, frost wedges and fissures are common features proving the presence of permafrost during the cold stages of the Late Glacial (Kolstrup 2004; Buylaert et al. 2009). During the Late Pleniglacial and colder phases of the Late Glacial, aeolian processes were facilitated by periglacial climate leading to the deposition of cover sands in the North European Plain and the formation of the European Sand Belt stretching from Belgium and the Netherlands in the west, across Germany, southern Denmark, and Poland into Russia in the east (Zeeberg 1998; Kasse 2002; Koster 2009, Chapter 11). Different wind directions in Central Europe were connected with disappearance of northern katabatic winds flowing from the centre of European Ice Sheet and their replacement by westerlies during the Late Pleniglacial to Late Glacial periods (Holuša et al. 2022).

The retreat of the local ice sheets in Great Britain and Scandinavia after the end of the LGM was followed by a similar response. The unloading of the steep walls, glacio-isostatic rebound, frost wedging and the degradation of permafrost triggered the operation of the paraglacial processes in the form of rock slope failures of different types (rockfalls, large-scale slope deformation, catastrophic events, etc.). The connection of these events with the deglaciation is evident given that they took place just after the onset of the deglaciation; in fact, in Great Britain (chapter “[Great Britain and Ireland](#)”) and Scandinavia (chapter “[Scandinavia](#)”), rock slope failures clustered during the early stages of the deglaciation; lags up to 1.6–2.0 ka were reported from Great Britain and Scandinavia.

In parallel, the retreat of the European Ice Sheet was followed by periglacial conditions prevailing in the glacier-free areas. In fact, an intense accumulation of aeolian deposits took place in Great Britain, with most of the loess deposition occurring between 25 and 10 ka (Stevens et al. 2020) and coversand being deposited especially between 26 and 15 ka (chapter “[Great Britain and Ireland](#)”). In Scandinavia, aeolian activity during the LGM was also inferred from ventifacts, and luminescence dating of ice-sheet limits (chapter “[Scandinavia](#)”). The Late Glacial glacier-free areas of Great Britain experienced periglacial conditions that determined the formation of sorted patterned ground features above former permafrost in frost-susceptible regolith (earth hummocks) and the occurrence of mass movements producing terraces and lobes of debris, especially in granite mountains, extending their influence beyond the slopes reaching valley floors in some upland areas. During the transition to the warm Bølling-Allerød interstadial, very few rock slope failures occurred in Great Britain, which evidences that permafrost degradation was not the main driver at this time. By this time, following local deglaciation, some (talus-derived) rock glaciers

developed at low elevations in Scandinavia, even remaining active until ~11.1 ka when permafrost conditions ceased (chapter “[Scandinavia](#)”).

In Iceland, the marine-based character of the IIS determined its abrupt collapse eustatic sea level rise associated to the Bølling-Allerød interstadial, rapidly retreating from the shelf (Pétursson et al. 2015; Chapter 14). The retreat of the IIS determined that the outermost areas of north-western, north-eastern and eastern Iceland were subaerially exposed, enabling permafrost aggradation. However, the degradation of this early permafrost occurred rapidly due to the warm conditions of the Bølling-Allerød period (Etzelmüller et al. 2020). The response of the slopes of north-western, northern and eastern Iceland to the deglaciation was similar to that observed in Great Britain and Scandinavia, with the occurrence of numerous rock slope failures, concentrated especially within the first millennia of the deglaciation process, around ~10 ka, when the uplift rates were the highest. In addition to the rock slope failures, there is also evidence of solifluction processes starting following ~12.2 ka BP in a number of locations throughout Iceland (chapter “[Iceland](#)”).

Subsequently, at the transition between the Oldest Dryas and the onset of the Bølling-Allerød (14.6–12.9 ka) there was a substantial temperature increase that promoted a massive glacial retreat across European mountains, including the European Ice Sheet (see details in Palacios et al. 2022). Indeed, the use of CRE dating on relict rock glaciers provides a better comprehension of the timing of their formation and stabilisation, showing evidence that most features stabilised soon after formation, although their upper parts remained active during several millennia. This is the case of many rock glaciers in Iberian Mountains, which formed during the paraglacial phase at the onset of the Bølling-Allerød at ~15–14.5 ka, and must be linked to a glacialogenic origin (Palacios et al. 2017a; Andrés et al. 2018). As temperatures increased during this period, permafrost conditions moved to higher altitudes and those rock glaciers became relict. A similar process for rapid paraglacial formation of rock glaciers has been also described in other mountain regions, including the Alps (Hippolyte et al. 2009), southern Scandinavia (Linge 2020) and the British Isles (Ballantyne et al. 2009) (see chapter “[The Alps, Great Britain and Ireland and Scandinavia](#)”). At lower elevations, stratified screes in valleys and slopes have been associated with this phase (García-Ruiz et al. 2001).

Temperatures dropped again during the Younger Dryas (12.9–11.7 ka), which favoured the last major advance of the European Ice Sheet as well as of most mountain glaciers in Europe. In moderately high mountain ranges, such as the Pyrenees, Sierra Nevada or the Apennines, glaciers rarely exceeded the limits of the cirques as this period was characterised by cold, but arid conditions in southern Europe (Oliva et al. 2019). In other higher ranges, such as in the Alps - where MAAT were 3.5–5 °C lower than present and precipitation was up to 30% less (Kerschner and Ivy-Ochs 2008)—glaciers had a larger expansion and the largest units reoccupied the valley floors (see details in Palacios et al. 2022). Similarly to what occurred during the latest stages of the Oldest Dryas, the end of the Younger Dryas promoted the rapid shrinking of mountain glaciers that enhanced the reactivation of paraglacial processes in the cirque walls, which in turn favoured the development of rock glaciers and protalus lobes (Oliva et al. 2018b). CRE ages from inactive rock glaciers distributed

across several Iberian mountain ranges showed evidence that their fronts became relict immediately after their formation, although their roots remained active until the Mid-Holocene, when the higher temperatures favoured the melting of the last permafrost patches located beneath them (Palacios et al. 2015; García-Ruiz et al. 2016; Oliva et al. 2016b; Andrés et al. 2018). Although many rock glaciers in the Mediterranean mountains, such as in the southern Alps (Colucci et al. 2016a), could have also developed during this stage and be thus associated to a paraglacial origin, further research needs to confirm this hypothesis. The Younger Dryas promoted a widespread down-valley expansion of periglacial activity in southern Europe. This very cold climate regime expanded the permafrost domain to lower elevations and must have favoured the development of block streams in the highest mountain ranges and reactivated patterned ground features across the snow-drifted plateaus (Oliva et al. 2016b).

Consequently, the cold phases interrupting the long-term deglaciation process (Oldest and Younger Dryas) left both glacial and periglacial evidence in the highest mountain ranges in southern Europe, mainly with the formation of permafrost-related features derived from paraglacial dynamics. In fact, the presence of these inactive landforms under present-day climate conditions is indicative of the minimum altitude for the past occurrence of permafrost during those phases.

Numerous relict periglacial landforms found in the Alps and Carpathians were active during the Younger Dryas. However, they were often reactivated and does not constitute newly developed landforms of Younger Dryas age. Many nowadays relict rock glaciers in the Alps were formed between Heinrich Stadial 1 and the Younger Dryas and stabilised during the Early Holocene (Charton et al. 2021). Their past presence was supported by modelling of their Younger Dryas distribution for example in Hohe Tauern (Avian and Kellerer-Pirklbauer 2012), or in Err-Julier area in Swiss Alps (Frauenfelder et al. 2001). In the Carpathians, present data on the formation of rock glaciers indicate that most of their relict forms were active during the Younger Dryas (12.9–11.5 ka BP; Kotarba 1991–1992; Kłapyta 2013; Onaca et al. 2017a; Uxa and Mida 2017; Popescu et al. 2017). In the Tatras and other parts of Carpathians, the stabilisation of the rock glaciers took place with the prominent Early Holocene warming associated with the Preboreal period (11.8–10.0 ka; Zasadni et al. 2020). Younger Dryas cooling also enhanced hillslope activity, as shown by OSL ages for the upper periglacial cover beds in the Taunus Mts and Ślęza Mt. spanning 8.7 ± 1.0 to 12.5 ± 1.4 ka (Hülle et al. 2009) and 12.0 ± 1.0 to 14.2 ± 1.1 ka (Waroszewski et al. 2020), respectively. Furthermore, ice loading of the lithosphere induced by the proximity of the European Ice Sheet margin during the LGM reactivated the Sudetic Marginal Fault (Štěpančková et al. 2010, 2022), which led to hillslope processes such as gelifluction (Fig. 9) and enhanced hillslope activity continued till the beginning of the Holocene (10.9 ± 0.1 cal. ka BP; Štěpančková et al. 2010). The Younger Dryas also represents the final phase of the formation of aeolian landforms (Kasse 1999, 2002; Schirmer 1999), even though aeolian activity might have continued till the beginning of the Holocene at some places of North European Plain according to recent OSL dates (Moska et al. 2022).

4.3 *Holocene*

Marine sedimentary records, Greenland ice cores and a large number of terrestrial proxies showed evidence of a significant temperature increase in the North Atlantic region at the onset of the Holocene (chapter “[Quaternary Climate Variability and Periglacial Dynamics](#)”). The European continent has recorded substantial climate changes during the Holocene, with temperature oscillations of the order of ca. $\pm 2^\circ\text{C}$ that have caused significant disruptions in human civilizations (Mayewski et al. 2004). Temperature and moisture changes have thus affected type and intensity of cold-climate geomorphological processes prevailing in Europe (Oliva et al. 2018b). The sharp temperature increase recorded during the Early Holocene promoted an accelerated shrinking of mountain glaciers, which caused their final disappearance in many massifs, as well a rapid retreat and the complete melting of the European Ice Sheet during the Early Holocene (Palacios et al. 2022). The recently deglaciated environments were gradually replaced by periglacial processes that also migrate to higher elevations. Most of the periglacial landforms that were active during the Younger Dryas became progressively inactive, and only reactivated during the coldest Holocene phases (e.g. LIA).

In general, during the Holocene periglacial processes in Southern and Central Europe have mostly been restricted to the high mountain areas, where glaciers only existed during the coldest phases (Oliva et al. 2018b). As occurred in the Alps (chapter “[The Alps](#)”), the Apennines (chapter “[Periglacial Landforms Across Europe](#)”) or Iberian mountains (chapter “[The Iberian Peninsula](#)”), the cold-climate conditions during the Holocene promoted the formation and/or expansion of small glaciers in the highest mountains and decrease of periglacial processes to lower elevations whereas warmer phases favoured a lower periglacial activity, which also migrated to higher altitudes and favoured the spread of vegetation cover through the periglacial belt. In northern Europe, periglacial activity was limited to the summits and slopes of the highest mountains and cols of the British Islands (chapter “[Great Britain and Ireland](#)”) and to the new deglaciated terrain after the retreat of the local ice sheets in Scandinavia (chapter “[Scandinavia](#)”) and Iceland (chapter “[Iceland](#)”).

In most mid-altitude mountain ranges in southern Europe—with maximum elevations around 2000 m in the central and western Mediterranean and 2500–3000 m in the eastern fringe of the basin—the periglacial belt disappeared during the Holocene and was significantly reduced in the highest mountains. During the warmest periods, such as the Holocene Thermal Maximum (HTM; Renssen et al. 2012) and the Medieval Warm Period, only the mountain regions exceeding 2500–3000 m included a periglacial environment dominated by seasonal frost conditions, with permafrost only confined to the highest sectors of the Maritime Alps, as well as in the highest summits in Anatolia. However, geochronological evidence of Holocene periglacial processes in southern Europe is mostly related to rock glaciers and solifluction lobes, as well as other landforms and deposits, for which few ages are available (i.e., talus screes) (Fig. 11).

Actually, permafrost-derived landforms were probably active in many mountain ranges during the Early Holocene in the highest massifs in southern Europe (Oliva et al. 2018b). Indeed, in the central Taurus Mountains (Çiner et al. 2017) and the Pyrenees (Serrano et al. 2010) there is evidence of the formation of some still active rock glaciers during the HTM at ~6 ka. Similarly, in the southern Alps, some currently active rock glaciers showed activity during Holocene warm phases (chapter “The Alps”), with evidence for their formation during dry periods: 8 ka, 3.7–3.2 ka, and possibly 1.2 ka cal BP (Dramis et al. 2003). In some limestone massifs, below the periglacial environment, some ice caves containing perennial accumulations of ice started accumulating ice during at the onset of the Neoglacial period at 5.5–5 ka (Serrano et al. 2018a); the frozen ground in contact with the ice inside the caves has been also described as permafrost (Colucci et al. 2016b); thus, permafrost may have aggraded since then in these areas, persisting until today despite the MAAT well above 0 °C due to specific microtopographic conditions (very low solar radiation, existence of stable temperature inversion, or chimney effects) (Oliva et al. 2018b). Other periglacial phenomena associated with seasonal frost, such as solifluction lobes distributed across the present-day periglacial belt of Sierra Nevada at elevations between 2500 and 3000 m, were also active during this warm phase until 5 ka cal BP. These features include sedimentological evidence of solifluction activity during Mid and Late Holocene cold and wet phases, namely at 5–4 ka, 3.6–3.4 ka, 3–2.8 ka, 2.5–2.3 ka, 1.8–1.6 ka, 0.85–0.7 ka, and 0.4–0.15 ka cal BP (Oliva et al. 2009). The intensification of arid conditions in southern Iberia by 4.2 ka cal BP resulted in a long-term increase in the elevation of active solifluction processes in this semiarid massif (Oliva et al. 2011). As for solifluction, the alternation between cold and warm phases during the Holocene accompanied also by changing precipitation regimes must have also influenced the intensity of periglacial processes and the spatial distribution of periglacial processes. Stratified scree deposits at mid-altitude regions of the Picos de Europa revealed a lithological and topographical control on their formation regardless the prevailing climatic conditions of the Early-Mid Holocene (Ruiz-Fernández et al. 2019) (Fig. 12).

As in Sierra Nevada, the Alps (Veit 1988), or the Apennines (Giraudi 2005), warmer periods favoured soil formation and colder phases enhanced solifluction processes in seasonal frost environments, with variable intensities depending on the moisture regime. In fact, a similar timing for solifluction chronology has been inferred in these three mountain regions. Many of the debris-mantled slopes and blockstreams in the Alps and Carpathians were formed during the Pleistocene with only minor reshaping during the Holocene, but their reactivation only occurred in highest location during the Holocene’s coldest period, such as the LIA (Kotarba 1995; Puřan 2015; řerban et al. 2019). This was accompanied by enhanced physical weathering and rock fall activity (Kotarba and Pech 2002).

With the onset of the Holocene, associated with permafrost degradation combined with the establishment of groundwater circulation, numerous paraglacial landslides, slope failures and rock avalanches in the Alps and Carpathians originated (Margielewski 2006; Krautblatter et al. 2010, 2012; Pánek et al. 2014, 2016). Similarly, Early Holocene hillslope activity was also described from Variscan Ranges

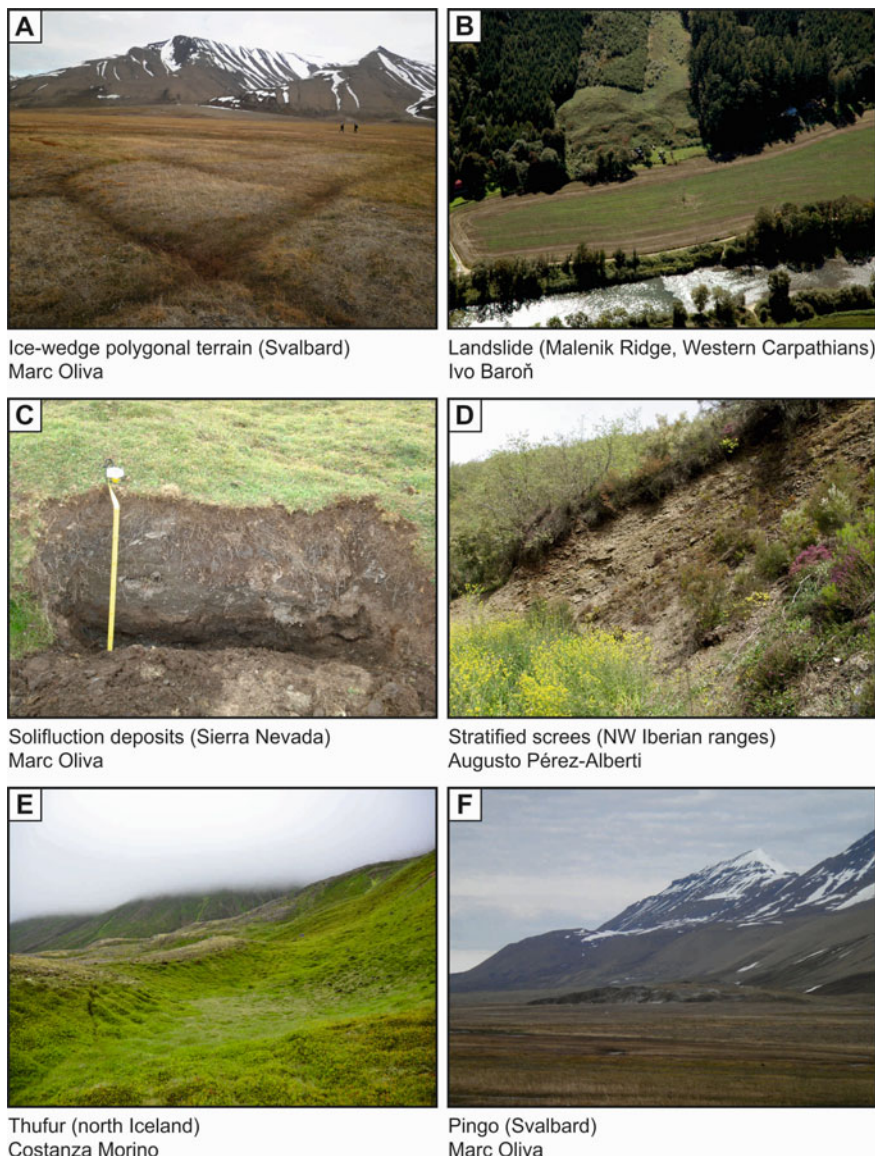


Fig. 12 Examples of periglacial features formed during the Holocene in Europe

of the Sudetes Mts. (e.g.; Štěpančíková et al. 2010). At some higher places of the Alps and Carpathians, rock glacier stabilisation did not take place earlier than the Early Holocene (Vespreameanu Stroe et al. 2012; Zasadni et al. 2020; Charton et al. 2021) and their activity increased during Holocene colder periods, especially the LIA (Urdea 1998; Avian and Kellerer-Pirklbauer 2012). This reactivation only occurred

in the highest massifs of the mountains. However, in some Carpathian ranges (such as Tatras) no rock glacial activity was detected during the LIA (Kędzia 2014). Special attention was paid on the modelling of mountain permafrost distribution and the distribution of rock glaciers and permafrost in European Alps in the past and today showing its larger distribution during Holocene cold periods, such as the LIA (e.g.; Avian and Kellerer-Pirklbauer 2012; Kellerer-Pirklbauer et al. 2012).

In Great Britain and Scandinavia, climatic conditions characterised by relatively warm summers prevailed during great part of the Early Holocene. Under these circumstances, Holocene frost weathering activity was restricted to granular disaggregation and flaking, especially in granitic and sandstone lithologies. In parallel, processes conducting to patterned ground sorting (circles, nets, stone stripes) were especially active in frost-susceptible soils in the mountains of the British Islands (chapter “[Great Britain and Ireland](#)”) and in till with water-saturated active layers, after the Early Holocene deglaciation in Scandinavia (chapter “[Scandinavia](#)”). In steep slopes inside the limits of the ice during the Loch Lomond Stadial/YD ice sheet, rock slope failure activity continued throughout the Holocene. Accordingly, there is evidence of rock slope deformation, permafrost creeping and rock glacier formation in Scandinavia starting at 11–10 ka in some places. In Iceland, the cirque glaciers of northern Iceland (Tröllaskagi Peninsula), retreated accordingly with the trend of the IIS outlets after culminating a main glacial advance at ~11 ka, triggering paraglacial processes at the steep slopes of the cirques and valleys. Consequently, cirque glaciers evolved to debris-covered glaciers and rock glaciers, which rapidly stabilised by 10–9 ka. Subsequently, new generations of rock glaciers formed, some of which are currently active albeit they remain static since 6.3 ka (chapter “[Iceland](#)”).

In Scandinavia, YD climate and the high debris supply linked to the paraglacial conditions prevailing at the Early Holocene allowed the development of pronival ramparts—that remained active throughout the Holocene—and the aggradation of fans, attaining a later peak around 9–8 ka. Climate was also prone for the occurrence of colluvial events (slope wash) and the formation of sorted circles and sorted stripes, dated at ~8–6 ka. After the retreat of the YD ice sheet, aeolian activity was intense in the new ice-free areas, with the formation and stabilisation of parabolic dunes at ~10.9–10.2 ka due to sand blasting from glacial deposits. However, some dunes were reactivated successively during the Holocene, especially after ~8.3 ka (chapter “[Scandinavia](#)”). Aeolian activity started also in Iceland as the IIS retreated at the highlands by ~10 ka, and subsequently, with the glacier disappearance during the Holocene Thermal Maximum (HTM). Consequently, glaciers left extensive ice-free surfaces covered by till. Such surfaces and the glacial rivers became important sources of aeolian sediments that contributed to the construction of sand-fields in the glacial margin areas and the accumulation of sand in subaerial Holocene lava flows and originating different types of dunes, especially in southern Iceland. On the other hand, aeolian activity contributed to the generation of deflation surfaces such as stone or gravel pavements (sandy lag gravel) and abrasion features such as sculpted rocks, micro-yardangs and ventifacts (chapter “[Iceland](#)”).

Throughout the Holocene, rockfall activity continued, with major events of rock-slides and rock avalanches during the Middle and Late Holocene, peaking at ~6–4 ka

and ~1.7 ka in Great Britain (chapter “[Great Britain and Ireland](#)”), an at 8.0–7.0, 5.4–4.0 and 3.0 ka in Scandinavia (chapter “[Scandinavia](#)”). In Great Britain, landslides and debris flows were also amongst the dominant processes reshaping mountain slopes (chapter “[Great Britain and Ireland](#)”). Warm climate conditions prevailing during the HTM, enhanced rock slope activity in Scandinavia, especially at the end of such period, due to permafrost degradation and its resulted minimum depth. The rise of the lower altitudinal level of discontinuous permafrost (1650–1700 m) led to the cessation of some circles / stripes activity, with stabilization ages between ~8 and ~7 ka. The climatic setting of this period, especially at the end (~7.0–6.0 ka), contributed to a decrease of debris flow activity as snow melt and the degradation of the permafrost active layer were more limited and less likely to occur. Conversely, significant avalanche activity occurred during the Mid- and Late Holocene, with the deposition of numerous fans (boulders dated at 7.4–2.3 ka, possibly linked to paraglacial processes and permafrost degradation during the HTM) and the formation of snow-avalanche impact ramparts (younger than 3.6 ka). However, after the HTM, a decline in rock fall activity was evident due to the exhaustion of permafrost, not enough for triggering large rock-slope failures (chapter “[Scandinavia](#)”).

In Great Britain, the climate underwent a significant cooling trend during the last 5 ka, where MAAT fluctuated in the range of ± 2 °C. Consequently, during, the Late Holocene cold climate conditions in Great Britain supported solifluction processes (above 600 m) and the movement of ploughing boulders, active since at least the Mid-Holocene. The same was found in Scandinavia, with solifluction lobe activity and development starting during the Mid-Holocene (phases at ~7.4–6.7 ka) and continuing after the HTM during the Late Holocene (4.2–3.4 ka, 2.6–2.1 ka and 1.5–0.5 ka) driven by millennial changes of moisture supply (chapter “[Scandinavia](#)”). Enhanced solifluction activity was also reported from several sites in central and western Iceland between 4 and 2 ka (chapter “[Iceland](#)”). Moreover, cooling trend starting after the HTM led to the formation of earth hummocks, some of which have been dated at between ~4.8 and ~3.1 ka in southern Norway and younger than ~4.1 ka in Finnish Lapland. Earth hummocks also formed in Iceland following the HTM, with different phases of formation at ~4.5 cal ka BP, or after 2.8–2.6 cal ka BP and prior to 1104 CE (chapter “[Iceland](#)”). The prevailing climate might have been cold enough to allow permafrost aggradation in the palsas of northern Finland and northern Pechoria by 3.0–2.5 ka, possibly starting between 5.0 and 4.0 ka (chapter “[Scandinavia](#)”). The cooling trend that followed the HTM and characterised the Neoglaciation would have led to the formation of the ground ice of the Icelandic palsas (Orravatsnústír), starting at 4.5 ka BP (chapter “[Iceland](#)”).

However, cryogenic processes were not the only ones that characterised the Mid- and Late Holocene periods. Evidence of increased aeolian activity since 6.0 ka was found in north-western Norway, culminating by around 1.7–0.7 ka. Strong winds and intense aeolian activity have been inferred also in Great Britain on the basis of features such as ventifacts, wind-patterned ground, wind stripes, gravelly/bouldery surfaces and vegetated aeolian sand deposits, accumulated in north-western Ireland especially during the Middle and Late Holocene (chapter “[Iceland](#)”).

4.4 Little Ice Age

From approximately 1300 to 1850 CE, the climate of Europe underwent the coldest most recent period, that is known as the LIA (Oliva et al. 2018a). During these centuries, which have been used as a reference for Holocene cold phases, geomorphological processes in Europe reacted accordingly to the lower temperatures: (i) glaciers that survived the Medieval Warm Period expanded to their largest volume over the last several millennia and several other small ice masses formed in the highest massifs (Bradley and Jones 1992), and (ii) the spatial domain of periglacial dynamics expanded down-valleys and cryogenic processes reappeared in some areas where seasonal frost activity was reduced during the preceding warm period and is no longer active at present.

In Europe, average temperatures during the coldest stages of the LIA recorded during the Minimum Maunder (1675–1715 CE) were ~ 0.5 – 1 °C colder than today, with changing precipitation patterns (Grove 2004; Luterbacher et al. 2006, 2016; Mann et al. 2009). In some regions of southern Europe, such as the Iberian and Balkan peninsulas, the difference of temperatures between the LIA and the present could have been even larger. Based on glacial and periglacial geomorphic evidence, temperatures during the last cold stages of the LIA were ~ 1 °C lower compared to present-day values, up to ~ 2 °C with respect to the Minimum Maunder (Oliva et al. 2018a).

Colder conditions favoured more extensive long-lasting and perennial snow-patches than today in the high mountains in southern Europe. In turn, nival processes were more intense, with increased avalanched activity (García-Hernández et al. 2017) that favoured the formation of (well-preserved) protalus ramparts at the foot of slopes as well as more active and widespread solifluction processes (Oliva et al. 2009). Colder conditions must have also favoured the reactivation of cryoturbation processes, including pattern ground phenomena in the highest sectors of the mountains (Fig. 13). Permafrost-related features, such as rock glaciers and protalus lobes formed and/or reactivated in the highest slopes of the main massifs, at ~ 2350 m in the Balkan region (chapter “[The Balkans \(Without Carpathians\)](#)”), ~ 2400 – 2500 m in the southern Alps (chapter “[The Alps](#)”) and ~ 2560 m in the Pyrenees (chapter “[The Iberian Peninsula](#)”). In addition, ice caves accumulated more ice during this period in several high karstic mountains in southern Europe, in areas located several hundreds of metres below the concurrently glaciated environments, even at ca. 1000 m in the Dinaric Alps (Kern et al. 2018).

As temperatures started rising during the second half of the twentieth century, glaciers start to retreat and many disappeared (Hughes 2014, 2018), paraglacial activity increased (Serrano et al. 2018b), periglacial processes migrate to higher elevations and alpine permafrost environments in mid-latitude high mountain started to degrade and reduce in extent (Oliva et al. 2016c). All these processes were parallel to substantial geological changes in the upper sectors of the highest ranges (Pauli et al. 2012). Pattern ground developed in the forelands of glaciated environments in the Pyrenees (Feuillet and Mercier 2012) as well as frost mounds that also formed near



Fig. 13 Examples of periglacial features formed during the Little Ice Age in Europe

the glacier margins on unconsolidated sediments (Serrano et al. 2018b). Proglacial areas and Little Ice Age moraines were also reshaped by debris flows, solifluction processes and alluvial dynamics, with the formation of several new lakes and lagoons (Serrano et al. 2018b). Shrinking glaciers favoured also rock avalanches and landslides that fed the talus slopes within the cirques, where, in some cases, even promoted

the development of incipient rock glaciers and protalus lobes that tend to become relict as temperatures keep increasing and buried ice masses and isolated patches of permafrost degrade (Gómez-Ortiz et al. 2014, 2019).

Glaciers in European Alps reached their LIA maximum extent in the fourteenth, seventeenth and nineteenth century AD with most of them reaching their maxima around 1850/1860 AD (Ivy-Ochs et al. 2009). Advances of local glaciers has also effect on periglacial dynamics such as physical weathering, debris flows and other hillslope processes, which account also for non-glaciated Central European mountain ranges of Central Europe during the LIA, such as Carpathians (Kotarba 1995; Kotarba and Pech 2002). In the Southern Carpathians, the formation of blockstreams was re-initiated above 2100 m (Puşan 2015; Şerban et al. 2019) and even rock glaciers re-activated above 2050 m (Urdea 1998), while this was not found for rock glaciers in the Western Carpathians (Kędzia 2014). Nevertheless, the climatic conditions of the LIA were not sufficient to change significantly the periglacial dynamics in the lower Variscan ranges of Central Europe, such as the Sudetes Mts. (chapter “[The Central European Variscan Ranges](#)”).

The LIA was the coldest period of the last 10 ka in Great Britain, with climate conditions that maintained perennial snow patch in some Scottish mountains. Its enhanced storminess brought intense wind erosion and deposition at the plateau areas (chapter “[Great Britain and Ireland](#)”). In Scandinavia, the lowering of the lower limit of discontinuous permafrost culminated during the LIA (~1750 CE), reaching ~1250 m a.s.l., leading to the aggradation of shallow permafrost, which led to a reduction of the frequency of rock slope failures and conversely, to the increase of the rock-fall talus accumulation (eighteenth century). The expansion of permafrost was also prevailing in Iceland, forming the youngest existing permafrost in palsa areas—although this landscape might have developed earlier, during the Medieval Warm Period (MWP)—so far and reaching the maximum Holocene extent (Etzelmüller et al. 2020) (chapter “[Iceland](#)”).

Climate of the LIA was also favourable for glacial advances—which in some cases reworked ice-cored moraines—and snow avalanches, some of which were dated to 0.6 ka in Scandinavia (chapter “[Scandinavia](#)”). In Iceland, the largest ice caps and smaller glaciers advanced, reaching their maximum Holocene extent and reworking older Neoglacial moraines in northern Iceland (Fernández-Fernández et al. 2019). In the surroundings of some ice caps at central Iceland, frost-crack macropolygons and likely ice-wedge nets developed between the sixteenth and nineteenth centuries, in the coldest decades of the LIA (chapter “[Iceland](#)”). Meanwhile, in most of the low-gradient vegetated surfaces, *thúfur* formed in accordance with the coldest and wettest periods. In other vegetated slopes, the loss of vegetation cover linked to the climate deterioration triggered solifluidal processes, with the reactivation of some lobes formed before the MWP. Conversely, the potential formation of some rock glaciers in the glacial cirques of northern Iceland (Tröllaskagi) has also been attributed to the LIA (chapter “[Iceland](#)”).

Once the climate conditions of the LIA ended, glacial retreat from the LIA positions started. Subsequently, patterned ground such as stone circles rapidly developed in some recently deglaciated forelands, and gully erosion and debris-flow activity

initiated in the steep slopes of the lateral moraines, contributing to their degradation (chapters “[Scandinavia](#)” and “[Iceland](#)”). Even there is evidence of rock glacier formation from moraines thank to increased debris supply and the incorporation of ice (interstitial ice or buried glacier ice) in Scandinavia (chapter “[Scandinavia](#)”).

5 The Interaction Between Periglacial Landscapes and Human Activities

The landscapes of Europe have been continuously transformed by human activities: from the local impacts in their surrounding environments carried out by hunter-gatherers since the last glacial cycle to the major changes imposed by present-day industrialised societies (Pinto-Correia et al. 2018). Obviously, these transformations followed different spatial and temporal patterns with variables impacts, but have almost affected the entire continent. Very few areas remain virtually intact and periglacial landscapes are not an exception as they have been also massively disturbed by human activities. This is the case of inherited periglacial landscapes in Central and Northern Europe above latitude 50 °N, where most of the infrastructures and settlements networks have been established on past permafrost terrain although current MAAT are well above 0 °C (Fig. 1). In these areas, periglacial landforms are almost no longer visible at the surface although periglacial deposits are preserved below and geotechnically condition the building processes, foundation of houses, construction of roads, etc. (Fig. 14).

Actually, inactive periglacial landforms are widespread across Europe, particularly in mountain regions where they tend to be better preserved because of generally lower anthropogenic pressure. However, present-day periglacial landscapes in many mountain areas have been also largely transformed by the construction of ski complexes, particularly in gently sloping glacial valleys and valley slopes currently affected by periglacial conditions. The initial conditioning of the ski slopes also involved the removal of (glacial erratic) blocks, dismantling of moraines, the use of glacial lagoons to produce artificial snow, soil disturbance, and other impacts across the entire ski area associated with the construction of the resorts, ropeway transport systems, high-altitude restaurants, huts, etc. As a result, this tourism massification induced a significant alteration of the natural ecosystem dynamics of the periglacial belt, together with severe impacts in some periglacial features abundant in these areas, such as solifluction landforms and even some larger scale landforms, such as rock glaciers (Fig. 14). In addition, snow grooming in ski resorts alters also the thickness and persistence of snow in the ground, which in turn affects the natural regime of ground temperatures and frost action dynamics.

Periglacial processes are still active in high mountain environments and in Northern Europe, particularly in high areas in Scandinavia and Iceland. Here, rapid mass movements (debris flows, mud flows, rock falls, etc.) within the periglacial domain frequently affect the road network, mountain tracks and trails and even



A Ski resort, Pas de la Casa (central Pyrenees)
Marc Oliva



B Road cutting a rock glacier (Swiss Alps)
José M. Fernández-Fernández



C Snow-avalanche defense structures (Swiss Alps)
José M. Fernández-Fernández



D Upslope treeline migration (central Pyrenees)
Marc Oliva



E Solifluction lobes in Sierra Nevada (S Iberia)
Marc Oliva



F Rock falls in permafrost terrain (Mt. Blanc massif)
Ludovic Ravel

Fig. 14 Examples of transformations of periglacial landscapes by human activities: **(A)** Ski resort built occupying an ancient glacial cirque currently affected by periglacial conditions, Pas de la Casa (central Pyrenees), **(B)** Rock glacier cut by the construction of the road in the Albula (Swiss Alps); **(C)** Defence structures against snow avalanches and debris flows (Swiss Alps); **(D)** Upslope migration of the treeline in the lower periglacial belt (central Pyrenees); **(E)** Solifluction lobes generated by the expansion of wetlands created by the spreading of irrigation channels in the Sierra Nevada (South Iberia); **(F)** Rock fall triggered by permafrost degradation at the foot of the Aiguille des Grands Montets (3285 m; Mt. Blanc massif, French Alps); almost affecting ski facilities

some settlements and mountain huts. They are frequent in spring when snowmelt favours slope processes, but they also occur in summer, particularly at high altitudes where permafrost degradation may also trigger slope failures. Slow mass wasting processes, namely solifluction, also affect widespread gentle and steep slopes in these regions, and some infrastructures such as ski resort facilities or electricity lines, must be constructed considering these processes. This is also valid for mountain regions in Central Europe, particularly the Alps and Carpathians. The Alps are under high anthropic pressure with tens of millions of inhabitants settling across the Alpine region, and therefore susceptible of being affected by periglacial processes. However, several Alpine countries have implemented effective and extensive monitoring systems in order to control potential natural hazards associated with permafrost thawing (Gruber and Haerberli 2007; Deline et al. 2011). However, rock falls during summer heat waves are becoming more widespread and a threatening risk for climbers and visitors of the highest massifs (Bodin et al. 2015; Magnin et al. 2017), and scientists are pushing stakeholders to implement new strategies to reduce their impacts on infrastructure in high mountain areas (Duvillard et al. 2021). Although to a lesser extent, mountain regions in Southern Europe are also affected by periglacial slope processes, which can damage ski facilities and huts (Serrano et al. 2018b). At the highest areas at ca. 3,000 m, some mountains still include permafrost, which is also undergoing a rapid degradation and favouring rock falls that can also affect climbers and mountaineers (Rico et al. 2021). Although not strictly periglacial, snow avalanches in mountain regions constitute also major hazards triggered in the periglacial belt; ski resorts and mountain villages are frequently affected by these processes and local administrations implement a wide range of measures to mitigate the risk (Fig. 14). Similarly, the alpine channels affected by snow avalanches can be also frequently used by debris flows in late spring and summer mobilising unconsolidated sediments from the periglacial zone to the valley floor, and may constitute also a potential hazard for mountain communities.

The strong relationship between human practices and geomorphic processes in some mountain ranges has also favoured an increase/decrease of the intensity of periglacial dynamics. Widespread grazing activities in steep hillslopes in the lower periglacial belt where soils are shallow and poorly developed have favoured shallow landslides and the formation of *terraces*. By contrast, in other areas such as in the Pyrenees, the accelerated abandonment of grazing activities have promoted a decrease of surface runoff and soil erosion (García-Ruiz et al. 2021). In Iceland, livestock—especially horse and cattle—accelerates the formation of *thúfur* (or earth hummocks; Schunke and Zoltai 1988) due to the frequent grazing in winter, when water saturation of the soil occurs: the animals put the feet in the depressions between the hummocks, and given the low soil cohesion, a gradual upwards push of the hummocks occurs (Arnalds 2015). Despite efforts on levelling terrain to remove hummocks, or other agricultural activities such as tilling and ploughing, these landforms may redevelop relatively fast, being a major problem for Icelandic farmers (Webb 1972). In parallel, the unvegetated periglacial belt is reducing as there has been an upslope migration of treeline over the last decades in response to both climate

warming and lower human pressure (Camarero et al. 2015). Human-induced deforestation since the Viking colonisation of Iceland (Church et al. 2007) lead to extensive deflation and damages to soils and vegetation cover (Brewington et al. 2015). Similarly, most of the Central European mountain ranges were deforested during the seventeenth-twentieth century (Cebecauer and Hofierka 2008; Kapustová et al. 2018) especially for the production of timber and charcoal, however, the deforestation started already during the Late Mediaeval period in some mountain ranges (Speranza et al. 2000). Deforestation lead to an increase of humidity in local soils and enhanced also periglacial processes in the summit areas of the mountains. The most intensive processes of soil erosion coincided with the early phases of the LIA in Central European mountains (Klimek et al. 2006). In the Iberian Central Mountain System, the increasing occurrence of wildfires have triggered an enhanced activity of debris flows (Sequeira et al. 2019), together with grazing, vegetation degradation and increases of frost processes (Vieira et al. 2003). In the Sierra Nevada, the bypassing the waters in glacial cirques and high valleys to transport water to the surrounding lowlands favoured the expansion of wetlands and promoted the formation of solifluction lobes in gentle slopes (Oliva 2009).

Despite these impacts, a long-term perspective allows us to be optimistic about the future preservation of the European periglacial landscapes. Since the creation of the first National Parks more than a century ago, many countries have preserved vast regions including periglacial landscapes. Currently, there are several environmental protection approaches (e.g. National Parks, Natural Parks, Geoparks, Geosites, Sites of Community Importance etc.) to protect glacial and periglacial landscapes, particularly in European mountain regions being visited by an increasing number of tourists (Debarbieux et al. 2014). In addition, over the last decades there has been an increase of the awareness of the uniqueness of the natural heritage of the periglacial landscapes. These measures for their protection must be expanded and improved, so that these features can be enjoyed and further studied by future generations.

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