Permafrost conditions in the Mediterranean region since the Last Glaciation

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1. Introduction

Permafrost is one of the elements of the cryosphere with the widest global distribution. The area occupied by permafrost in both hemispheres varies according to different authors between 22.10^6 km^2 (17% of the Earth's exposed land surface; Gruber, 2012) and 36.10^6 km^2 (24%; French, 2007). Terrestrial ecosystems in many ice-free areas at high latitudes and in high mountains are conditioned by the presence of permafrost. Additionally, permafrost has an influence on the socioeconomic activities of these areas, especially those related to the construction and maintenance of all kinds of equipment and infrastructures (Nelson et al., 2001, 2002; Bommer et al., 2010). However, as a research topic, permafrost has received much less global attention than other elements of the cryosphere, such as glaciers, which are highly visible elements in the landscape and carry important hydrological and global sea level consequences (e.g. Favier et al., 2014; Smith et al., 2017), and sea ice due to its influence on ocean-atmospheric interactions (e.g. Curran et al., 2003; Robinson et al., 2012) and geopolitical implications across the Arctic (Østerud and Hønneland, 2014).

As a result of the International Polar Year 2007-2008, there has been an important development in the study of the thermal state and current distribution of permafrost (Lewkowicz, 2010) and the active layer (Oliva et al., 2017), along with their geoecological implications (Jorgenson et al., 2001; Jorgenson and Osterkamp, 2005; Genxu et al., 2006). Most studies have focused mainly on the Arctic (e.g. Mackay, 2000; Brown et al., 2001; Lantuit et al., 2012; Fritz et al., 2017), but also on Antarctica (e.g. Vieira et al., 2010; Bockheim et al., 2013; Guglielmin et al., 2014; Oliva and Ruiz-Fernández, 2015) and mountainous areas with increased population and tourism pressure (e.g. Alps, Haeberli and Beniston, 1998; Boeckli et al., 2012). Permafrost degradation as the consequence of recent global warming is already posing risks and problems to certain socio-economic activities in many areas (Nelson et al., 2001, 2002; Ravanel, et al., 2017). Besides, during the last decade, scientists have pointed out the importance of permafrost not only at the local and regional level, but also at global scale examining the role of permafrost thawing in the carbon cycle and greenhouse gas fixation, as well as its contribution to recent and future climatic evolution (e.g. Hugelius et al., 2014; Zubrzycki et al., 2014; Schuur et al., 2015).

From a geographical perspective, the thermal state and distribution of permafrost have been much less studied in environments such as mountain ranges in the mid-latitudes. This is the case of the high mountains of the Mediterranean basin, where knowledge about the thermal dynamics, depth and extent of permafrost is uneven across the region.

The Mediterranean region is a transition zone between the subtropical climatic area and temperate regions more characteristic of the middle latitudes (Woodward, 2009). This region has its own climatic characteristics, conditioned by the presence of a nearly closed sea (the Mediterranean), and a complex and mountainous landscape defined by the existence of large peripheral mountain ranges, rugged coasts, and many islands and archipelagos. Most research on cold processes in the Mediterranean mountains has focused on the Quaternary glaciations (Hughes et al., 2006a, 2007, 2013; Hughes and Woodward, 2008), while permafrost and periglacial processes have been less examined (Hughes and Woodward, 2009). Thus, knowledge of the distribution of permafrost in this basin in the past is limited to generic maps that represent the area occupied by permafrost in the Northern Hemisphere during the Last Glacial Maximum (LGM; 26.5-20/19 ka (Clark et al., 2009); 27.5-23.3 ka (Hughes and Gibbard, 2015), as a complement to glacial extent (Brown et al., 2001, Vandenberghe et al., 2014; Kitover et al., 2016). However, the Mediterranean region is loosely represented in these maps since permafrost areas are barely included, except for certain areas such as the Alps and the SE coast of France. Besides, knowledge of the distribution of permafrost following the LGM and throughout the Holocene is limited and geographically sparse.

As far as the current dynamics are concerned, the existing studies on permafrost and related periglacial morphodynamics mainly concentrate on the southern slope of the Alps and the Pyrenees (Guglielmin et al., 1994; Hoelzle, 1996; Dramis et al., 1995; Ikeda and Matsuoka, 2002; Lugon et al., 2004; Julián and Chueca, 2007; Serrano et al., 2009, 2010; Colucci et al., 2016a), and to a much lesser extent on other massifs, such as the Sierra Nevada (Gómez-Ortiz et al., 2001, 2014; Oliva et al., 2016b) and the Atlas (Vieira et al, 2017).

In consequence, permafrost studies in the Mediterranean region are spatially limited and temporally dispersed. Thus, it is essential to encourage detailed reflection on the state of knowledge of permafrost in this region, and identify the areas and periods in which research on this topic needs to be intensified. Accordingly, the general aim of this article is to fill this gap by looking into the spatial distribution of permafrost in the mountains of the Mediterranean region since the Last Glaciation, which is defined by the last cold stage (Weichselian/Würmian; marine isotope stage (MIS) 5d-2), to the present day. To meet this aim, the following specific objectives will be addressed:

- Reconstructing the spatio-temporal evolution of permafrost in this region since the Last Glaciation.
- Identifying the Mediterranean areas where permafrost has been studied.
- Determining the best-known permafrost stages and those with knowledge gaps.
- Pointing out future lines of work to the scientific community centred on the study of permafrost in the Mediterranean mountains.
- Discussing the distribution of the spatio-temporal pattern of the areas affected by permafrost and its associated geomorphological processes within the climatic and environmental evolution recognised on the European continent since the Last Glaciation.

2. Regional setting

Extending over a surface of ca. 2.5.10^6 km^2 (Jeftic et al., 1989), the Mediterranean constitutes an almost enclosed basin between southern Europe (north), Anatolia (east), northern Africa (south) and the Iberian Peninsula (west), only connected to the Atlantic Ocean through the Strait of Gibraltar (ca. 15 km wide). Etymologically, Mediterranean derives from the Latin '*medius*' (middle) and '*terra*' (land), highlighting one of the main characteristics of the Mediterranean region: the presence of mountains next to the sea (Figure 1). Several volcanoes and mountain ranges with elevations exceeding 2500-3000 m asl stretch a few km off the coastline (Table 1).

Figure 1

Table 1

The Mediterranean basin is located in a boundary position between the mid-latitude westerlies and the influence of continental and subtropical high-pressure systems. The existence of multiple peninsulas, islands, rugged coastlines and mountain ranges determines a broad spectrum of climate regimes across the region, both in terms of temperatures and precipitation (Figure 1). This results in a wide range of hydrological, edaphic, geomorphological and biological processes prevailing in the Mediterranean region.

A broad variety of marine and terrestrial records preserve evidence of the nature and magnitude of Quaternary climate oscillations driving environmental changes in the Mediterranean basin (Woodward, 2009). Climate changes also determined large social transformations transitioning from the prehistoric inhabitants to the first sedentary agricultural communities that settled in the fertile soils of the Mediterranean lowlands (Mithen, 2004). The topographic configuration of the Mediterranean and its relief, together with relatively warm climate conditions during the Holocene, favoured the development of early civilizations, agriculture flourishing and trade and cultural exchange. Human-induced activities, together with shifts on ecosystem dynamics promoted by climate fluctuations, substantially affected the Mediterranean landscape, particularly since the Mid-Holocene, accelerating over the last few centuries.

Quaternary climate variability has followed different spatio-temporal patterns across the Mediterranean basin, which resulted in changing cold-climate geomorphological processes occurring in the mountains of the region. Glacial and periglacial activity has shaped the highest environments during the Late Pleistocene (Hughes and Woodward, 2016). The timing and spatial extent of glacial processes was highly conditioned by the combination of cold and moisture regimes, with significant spatio-temporal differences between the areas (Hughes et al, 2006a,b; Hughes and Woodward, 2008). Periglacial activity followed a similar pattern, expanding down-valleys during glaciations and shifting to higher elevations during interglacial periods (Oliva et al., 2016b). Though periglacial processes in Mediterranean mountains are currently mostly related to seasonal frost, environments affected by permafrost conditions in the past were significantly larger as suggested by a wide range of records that will be examined in detail in this research.

3. Methodology

This paper presents a detailed analysis of the published scientific literature on permafrost evolution in the Mediterranean region since the Last Glaciation. To better examine spatial and temporal patterns of permafrost conditions within the region, the Mediterranean basin has been subdivided in the following areas: Iberian Peninsula, southern Alps, Italian Peninsula, Balkan Peninsula, Anatolian Peninsula, northern Africa and Mediterranean islands (Figure 1).

For each of the areas we have conducted a detailed review of all the existing knowledge on permafrost conditions focusing on five main periods: Last Glaciation, deglaciation, Holocene, LIA and present-day. We establish the chronological boundary between the Last Glaciation and the deglaciation at 19-20 ka, after the LGM, when Clark et al. (2009) proposed the onset of deglaciation in the Northern Hemisphere, including also the Mediterranean region. We recognise that glaciers reached their maxima at different times during the Last Glaciation and glaciers in some areas retreated to smaller positions earlier than others, but the widespread deglaciation occurred after 19-20 ka in all parts of the Mediterranean mountains. The deglaciation extends until the Holocene and is divided considering different colder/warmer periods, sometimes characterised by phases of glacier advance/retreat, namely the Oldest Dryas (OD; 17.5-14.7 ka, stadial GS-2.1a), Bølling-Allerød (BA; 14.7-12.9 ka, interstadial GI-1) and Younger Dryas (YD; 12.9-11.7 ka, stadial 1 - GS1). The Holocene is subdivided following Walker et al. (2012) who established the limit between Early-Middle Holocene at 8.2 ka BP and the boundary between the Middle-Late Holocene at 4.2 ka BP. The LIA is defined as a single period within the Late Holocene encompassing from the 14th to 19th centuries, followed by the post-LIA warming. The characterization of current permafrost distribution and associated geomorphic processes in Mediterranean mountains allows comparing present-day dynamics with past permafrost conditions.

All these data have been summarized in a table including the location of the massifs, topographic characteristics (elevation, aspect), geomorphic evidence, available chronology (if existing) and associated references for each region. The distribution of permafrost-related features in each of the areas for each of the study periods has been mapped in GIS environment.

Rock glaciers constitute the most frequent and reliable information on palaeo-permafrost occurrence in Mediterranean mountains (Oliva et al., 2016b). Several fossil rock glaciers have been dated by analyzing the exposure of surfaces to cosmogenic radiation, usually by the presence of ^{36}Cl or ^{10}Be isotopes in the rock surface (i.e. Çiner et al., 2017). Two approaches have been generally used for inferring the age of rock glaciers: (i) their maximum age of formation has been often determined from samples collected from polished bedrock outcrops in which the rock glacier lies, (ii) their minimum age of stabilization has been also established based on samples taken from the surfaces of boulders distributed from the front of the rock glacier to the roots in the slope. Therefore, following this approach the dated samples should yield younger ages from the rock bedrock outcrops towards the roots of the rock glacier, which in fact has happened in most cases. However, there are still many unsolved issues concerning the application of surface exposure dating on rock glaciers, mainly related to: (i) the timing between boulders (i.e. rock glacier) stabilization and permafrost disappearance, (ii) the paleoclimatic significance of rock glacier activity in very active geomorphological settings experiencing paraglacial readjustment (Ballantyne et al., 2009; Palacios et al., 2016, 2017b; Andrés et al., 2018), (iii) the problem of excessive inherited nuclides yielding too old ages (Ivy-Ochs et al., 2009; Çiner et al., 2017) contrary to the moraines where glacial carving removes the cosmogenic inheritance, and last but not least (iv) the problem related to toppling and/or erosion of boulders that result in too young ages (e.g. Moran et al., 2016).

4. Geomorphological and sedimentological evidence

4.1 Iberian Peninsula

The Iberian Peninsula constitutes the SW tip of the Eurasian continent between the Atlantic Ocean and the Mediterranean Sea. Relatively flat terrain is mostly located in the central plateaus and coastal plains, separated by several mountain ranges aligned W-E with the highest peaks exceeding 2000 m: NW ranges, Cantabrian Mountains, Pyrenees, Central Range, the Iberian Range and the Betic Range (Figure 2). The rough terrain together with the high elevations have conditioned the magnitude and extent of glacial and periglacial processes during the Quaternary, including substantial variations of the spatial domain of permafrost through time. Sierra Nevada, in the Betic Range (Mulhacén, 3478 m), and the Maladeta massif, in the Pyrenees (Aneto, 3404 m) encompass the highest elevations in Western Europe outside the Alps.

A wide range of lithologies is found across Iberia. While some mountain regions show a relatively homogeneous composition - such as the NW ranges, Central Iberian Range and Iberian Range with prevailing metamorphic and crystalline rocks - or the Cantabrian Mountains with widespread thick Mesozoic carbonates -, others include a variety of rocks – such as the Pyrenees and the Betic Range with alternating Palaeozoic crystalline rocks and Mesozoic carbonates.

Quaternary periglacial activity has been more or less intense and extensive in the Iberian Peninsula in response to prevailing climate conditions. The intensity and duration of the cold determined if periglacial processes were conditioned by the presence of permafrost or seasonal frost conditions. This is also reflected in the widespread geomorphological and sedimentological features of periglacial origin in mid and high mountain environments in Iberia, but also at lower elevations such as in the NW corner and Cantabrian coast where periglacial activity almost reached sea level.

Last Glaciation

Although the chronology of periglacial activity in the Iberian Peninsula during the Last Glaciation is still uncertain (Oliva et al., 2016a), it is likely to consider that its extent and intensity followed periods of glacial advance and retreat. In this sense, the local Maximum Ice Extent (MIE) occurred asynchronously across Iberian mountain ranges: while the combination of cold and moisture was more effective for glacier development in NW Iberia between 35 and 45 ka cal BP (Valcárcel, 1998; Jiménez and Farias, 2002; Moreno et al., 2010; Serrano et al., 2012, 2013, 2015; Jiménez et al., 2013; Rodríguez-Rodríguez et al., 2014, 2016; Nieuwendam et al., 2016; Ruiz-Fernández et al., 2016), in the Central and some sectors of the Eastern Pyrenees it occurred prior to the LGM, probably at approximately 60 ka cal BP (García-Ruiz et al., 2003, 2010, 2013; Lewis et al., 2009; Delmas, 2015), in Sierra Nevada it took place around 30-32 ka (Gómez-Ortiz et al., 2012a, 2015; Oliva et al., 2014; Palacios et al., 2016) and in the rest of high mountain ranges, also in the Eastern Pyrenees, was slightly earlier or almost synchronous with the Last Glacial Maximum (LGM) (Pallàs et al., 2006; Domínguez-Villar et al., 2013; Hughes et al., 2013; Pedraza et al., 2013; Carrasco et al., 2015; Palacios et al., 2015). During the coldest stages of the Last Glaciation (e.g. MIS 2) glaciers extended across the highest mountain ranges in Iberia and periglacial processes expanded to the lowlands; warmer phases favoured periglacial activity only in mountain environments in parallel to glacier shrinkage.

Periglacial processes in mid and low environments were mainly driven by seasonal frost conditions, though some researchers have found evidence of ice wedge development in Pleistocene fluvial terraces between 200 and 1000 m in the Tajo and Duero basins, central Iberia, which may be related to very intense periglacial conditions but probably not associated with a permafrost regime (Badorrey et al., 1970; Asensio-Amor and González-Martín, 1974; Serrano et al., 2010a). Taking into account the presence of block streams and rock glaciers, the non-glaciated slopes of Iberian mountains were affected by permafrost conditions above 700-1300 m in northern mountain ranges (Figure 2), significantly higher in lower latitude mountain ranges, such as the Central Range (1800 m) and Sierra Nevada (2500 m) (Table 2). Sediments of periglacial origin, such as stratified and head deposits, could have been also produced under permafrost conditions, namely in the highest slopes and during the coldest stages. The presence of meter-sized patterned ground features, such as the sorted circles existing across the summit plateaus of most of the mountain ranges (Pyrenees, Sierra Nevada, Cantabrian Mountains), suggests also very intense periglacial conditions in these relatively flat areas where wind redistribution did not favour snow accumulation.

Figure 2

Deglaciation

As in many other glaciated environments in the Northern Hemisphere (Clark et al., 2009), a rapid deglaciation process started in Iberian mountains around 19-20 ka (Hughes and Woodward, 2008; García-Ruiz et al., 2010) at the same time as in mountain areas bordering the NE Atlantic to the north (Hughes et al., 2016). The terrain exposed by retreating glaciers was subject to paraglacial activity and intense periglacial processes under permafrost conditions. Many rock glaciers, protalus lobes and few block streams formed in recently deglaciated environments, though most of them became gradually inactive as temperatures kept raising, namely those located at lower altitudes (Table 2).

Table 2

However, the deglaciation was not a linear process and glacial expansion occurred in the highest mountain ranges during two major cold periods before the Holocene, namely the OD and YD. Following each of these cold stages, rock glaciers developed inside the cirques and highest slopes. As temperatures rose, rock glaciers became inactive at lower elevations with activity only at higher locations; in the case of the Pyrenees, rock glaciers formed during the OD at elevations above 2250 m with widespread permafrost above 2490 m, while during the YD rock glaciers developed above 2350 m and permafrost was limited to elevations above 2525 m (Oliva et al., 2016a). Many rock glaciers in the Cantabrian Mountains have been assigned to this period (Redondo et al., 2010; Gómez-Villar et al., 2011; Pellitero et al., 2011; García-Ruiz et al., 2016a,b), as well as in the Iberian Range, where several rock glaciers were also related to this period in the Cebollera Sierra (Ortigosa, 1986), and in the Demanda Sierra (García-Ruiz et al., 1979; Fernández-Fernández et al., 2017). Surface exposure dating suggests that rock glaciers were active in the Eastern Pyrenees between 15 and 10.5 ka (Palacios et al., 2015; García-Ruiz et al., 2016a,b) and in the Sierra Nevada from 12.8 to 6.4 ka (Gómez-Ortiz et al., 2012a; Palacios et al., 2016). Patterned ground probably related to permafrost conditions may have been active during the coldest stages of the OD and YD in the high lands of the highest mountain ranges.

Holocene

The onset of the Holocene saw the last massive deglaciation in Iberian mountains that was parallel to a progressive shift of periglacial conditions to higher elevations (Oliva et al., 2011). It is unlikely that the atmospheric temperatures recorded in Iberian mountains during the Holocene favoured the existence of widespread permafrost regimes (Oliva et al., 2016b), with the only exception of the highest massifs in the Pyrenees (Serrano, 1998). In the other highest mountain ranges permafrost conditions during the Holocene may have been marginal and only related to favourable geomorphological settings.

The Pyrenees, Cantabrian Mountains and Sierra Nevada were the only environments encompassing glaciers during the coldest stages of the present-day interglacial. Apart from the LIA glaciation, Gellatly et al. (1992) dated a cold phase in the Troumouse cirque (French Pyrenees) between 4.6 and 5.1 ka cal BP and García-Ruiz et al. (2014) reported evidence of glacier activity in the Pyrenees during the Neoglacial (5.1, 3.5 ka) and Dark Ages (1.4-1.2 ka) periods. Based on denudation rates, debris supply from cirque walls and flow displacement rates, Serrano et al. (2006, 2011) suggested that some active rock glaciers today in the Pyrenees (i.e. Argualas, Bastampé, Besiberri and Guerreys) probably developed between 3.4 and 6.2 ka; therefore, during part of the Mid Holocene permafrost environments may have existed in the highest mountains. This stage of rock glacier development seems to correlate with Neoglacial cold phases favouring glacial expansion in the Troumouse cirque (Gellatly et al., 1995) and Monte Perdido massif (García-Ruiz et al., 2014). The morphology and internal stratigraphy of the rock glaciers, some of which reshaped by LIA glacial advance, suggests that at least seven currently active rock glaciers originated during the Mid Holocene (Serrano et al. 2002, 2010b, 2011). Ice caves in the Central Pyrenees also started forming during the Mid Holocene, namely between 6 and 2 ka cal BP (Sancho et al., 2016).

In Sierra Nevada, Oliva and Gómez-Ortiz (2012) also identified glacial advances during the Neoglacial period (2.8-2.7 ka cal BP) and Dark Ages (1.4-1.2 ka cal BP). These glacial conditions were accompanied by an extent of periglacial activity, enhanced erosion processes and decrease of vegetation cover (Oliva-Urcía et al., 2013), with permafrost conditions only in the highest parts of the glacial cirques (Oliva and Gómez-Ortiz, 2012). The existence of permafrost was confirmed by surface exposure dating in some areas such as in Sierra Nevada, where rock glaciers were active until the Mid Holocene in favourable topographic environments inside the YD glaciated cirques (Palacios et al., 2016). In some cases, the development of protalus lobes during the Holocene may be also indicative of permafrost conditions (Table 2).

LIA

In Iberia, colder than present-day climate conditions accompanied by fluctuating precipitation prevailed during the LIA (Barriendos, 1997; Rodrigo, 1999). These conditions led to glacial advance in the Pyrenees, as well as the appearance of small glacial spots in the Cantabrian Mountains and Sierra Nevada (González-Trueba, 2006, González-Trueba et al., 2008; Gómez-Ortiz et al., 2006, 2009; Oliva and Gómez-Ortiz, 2012). Periglacial conditions expanded down-valleys and more extensive snow fields remained during the summer season. Permafrost conditions reactivated and new protalus lobes and rock glaciers formed in the Pyrenees above 2560 m (Serrano et al., 2001, 2011; Fernandes et al., 2017). Several currently active rock glaciers are linked with LIA lateral moraines, suggesting their origin following the LIA in historical times. All of them are relatively small and influenced by glaciers, which indicates the elevation boundary of permafrost conditions above 2560 m (Serrano, 1998; Serrano et al., 2004, 2011; González-García, 2014). With the exception of the surroundings of the glaciated areas and highest periglacial environments, it is unlikely that permafrost existed in other areas.

Present-day

Temperature increase in Iberian mountains since the last decades of the XIX century has been quantified in about 1 ºC (González-Trueba et al., 2008; Oliva and Gómez-Ortiz, 2012). The 0 ºC isotherm lies today close to the top of the highest peaks in the Cantabrian Mountains (2500 m; González-Trueba et al., 2007) and Sierra Nevada (3400 m; Oliva et al., 2016b), which explains the inexistence of widespread permafrost conditions. In these massifs, the bottom of the northern cirques that held glaciers during the LIA encompasses buried ice and permafrost covered by debris left by paraglacial dynamics (Serrano et al., 2011b; Gómez-Ortiz et al., 2004; Ruiz-Fernández et al., 2016; Pisabarro et al. 2016). In the Veleta cirque, in Sierra Nevada, a rock glacier formed during post-LIA deglaciation shows multiple subsidence and collapses as a result of the accelerated melting of the frozen body existing in its interior (Gómez-Ortiz et al., 2014).

In the Pyrenees, BTS measurements, geophysical surveying and geomorphological observations made in several massifs indicate that mountain permafrost regime occurs today above 2630 m in northern aspects and 2800 m in southern ones (Serrano et al., 1999, 2001, 2002, 2006, 2009, 2011a; González-García et al., 2014). Possible permafrost conditions have been estimated above 2400 m in northern slopes and above 2650 m in southern ones, with high variability depending on the massifs. While possible permafrost in the Infierno massif is located above 2450 m, in the Maladeta it exists above 2760 m and in the Posets above 2800 m (Serrano et al., 2001, 2009; González-García, 2014). The same variability is detected in the probable permafrost areas, located generally above 2700 m, oscillating between 2630 m in north faces and above 2800 m in southern ones. While in the Infierno massif the probable permafrost lies above 2650, in the Maladeta it sits above 2890 m and in the Posets above 2950 m. Three types of high mountain permafrost have been differentiated: (i) climatic origin, as suggested where MAAT are around -1/-2 ºC; (ii) topoclimatic, located in lower environments where cirque walls condition low incident radiation; and (iii) morphodynamic, where buried ice existing below the debris cover favours permafrost aggradation (González-García, 2014). In recently deglaciated environments there are landforms linked to frozen bodies, such as protalus lobes and frost mounds, as well as other, such as patterned ground features and solifluction lobes located always around 2700-3000 m, which may be also associated with permafrost (Serrano et al., 2000, 2001; Feuillet, 2010; Feuillet and Mercier, 2012; González-García, 2014; González-García et al., 2017).

Both in the Cantabrian Mountains (Gómez-Lende et al., 2014) and the Pyrenees (Bartolomé et al., 2015) there are also other landforms including subsurficial frozen features, which are an inheritance from the LIA. Ice caves show bedrock recording below freezing mean annual ground temperatures in contact with the perennial ice, and therefore may be defined as permafrost. In the case of the Cantabrian Mountains, the organic remnants trapped in the ice provided ages between 200 and 600 cal yr BP (Gómez-Lende, 2015), whereas in the Pyrenees ranged between 200 and 1200 cal yr BP (Bartolomé et al., 2015; Leunda et al., 2015; Sancho et al., 2016).

Figure 3

4.2 Southern Alps

The Southern Alps constitute the south side of the European Alps and include only the southern slope of the highest summit of the European Alps like the Monte Bianco (4810 m), Monte Rosa (4634 m) but also entirely some other peaks around 4000 m (like Gran Paradiso 4061 m or Ortles 3905 m). The Southern Alps are crossed by large valleys mainly W-E oriented (Aosta Valley, Valtellina) or N-S (Ossola Valley, ValCamonica, Adige Valley; Piave Valley or Tagliamento Valley). The Southern Alps can be divided into four main sectors: the Southern sector or Maritime Alps located between Italy and France south of Maddalena Pass; the western sector located between Maddalena Pass and the Ossola Valley; the central sector between the Ossola Valley and Adige Valley and the Eastern sector eastward of Adige Valley. These sectors reflect different local climate and different prevailing lithologies.

The southern and western sectors are characterized mainly by prevailing metamorphic and crystalline rocks while in the eastern sector Mesozoic carbonates prevail. Within the central sector the southern areas are dominated by Mesozoic carbonates whereas the northern areas by metamorphic and crystalline rocks. The differences in climate of these sectors are mainly related to the variation in the precipitation regime that reflects the main orientation of the mountain belt. In general, the outer parts of the mountain belt (southern and the more eastern areas of the Eastern sector) are the wettest while the northern (and inner) areas of the central sector are the driest.

Last Glaciation

There is still considerable uncertainty about the timing of the last glacial stadials recorded in the Alpine end-moraines at about 18-21 ka cal BP before the onset of glacial termination (Ivy-Ochs et al., 2008), although the LGM culminations in the southern Alps may span from 26 to 21 ka cal BP (Monegato et al., 2007, 2017; Ravazzi et al., 2012; Mozzi et al., 2013; Rossato et al., 2013; Federici et al., 2017). Paleo-temperatures between 30 and 17 ka BP have been recently reconstructed in NE Italy through the analyses of chironomids of lacustrine sediments series (Lake della Costa, Samartin et al., 2016). These authors associated the lack of chironomids remains around 24 ka and 18 ka cal BP to very cold conditions correlated with Heinrich events (HE; Heinrich, 1988) HE-2 (24.2 ± 3.8 ka cal BP) and HE-1 (16.9 ± 3.3 ka cal BP) (Hemming, 2004). In these events possibly a weakening of the Mediterranean low-pressure systems occurred with a higher aridity in the Mediterranean (Tzedakis et al., 2004; Kwiecien et al., 2009). On the other hand, a southward shift of the storm track between 26.5-23.5 ka cal BP has also been suggested, together with a change in the seasonal distribution of precipitation, predominantly occurring between spring and autumn (Luetscher et al., 2015).

Evidence of active periglacial processes at low elevations as well as in the Po plain are almost unknown. Nevertheless, two stratified scree deposits have been described in the Maritime Alps (Pappalardo, 1999) with stratified screes, partially cemented, on the southern slopes of M. Cornizzolo (1242 m) and on M. Barro (923 m) very close to Lake Como. Possible cryoturbations on the loess cover at Bagaggera (310 m), only few km south of Lake Como, have also been detected (Guglielmin, unpublished data). An interesting indicator of periglacial or cryotic conditions during the Late Pleistocene and LGM is the distribution of loess. According to Cremaschi et al. (2015) between 70 and 35 ka cal BP large loess deposits were deposited along the northern, western and southern margins of the Po plain. In the Julian Alps, between Italy and Slovenia, four rock glaciers have their front at 1076 m on average (Colucci et al., 2016a), which is very close to the estimated ELA during the LGM (1200 m, Colucci et al., 2014). This also suggests a temperature drop of at least 7.1-7.6 °C, which is in accordance with LGM temperature reconstructions for this sector of the Alps (Kuhlemann et al., 2008).

The presence of relict block streams and blockfields is also reported (Figure 4), and the largest area covered by such features (140 km^2) is the Ultramafic Lanzo Complex a short distance from Turin, at an elevation of 1000 to 1650 m. These landforms could be related to permafrost conditions during the LGM (Fioraso and Spagnolo, 2009; Paro, 2011). Other block streams and blockfields are located in the internal margin of Piedmont between Sangone valley (50 km S Turin) and Oropa (150 km NE of Turin) (Paro, 2011).

Figure 4

Deglaciation

The onset of glacial retreat in the southern Alps is thought to have occurred at 20.8 ± 1.5 ka (Gianotti et al., 2008), almost simultaneously with the Fennoscandinavian Ice Sheet (FIS) (A.L. Hughes et al., 2016). Downwasting of the large piedmont glaciers back into the Alpine valleys during the Late Glacial promoted widespread paraglacial activity, although several glacial readvances took place between the OD and the YD (Ivy-Ochs et al., 2008, 2009). Remnants of moraines of the YD glacier advance have been morphologically associated to rock glaciers from the lowermost belt (Frauenfeder et al., 2001; Ivy-Ochs et al., 2009).

Rock glaciers constitute widespread periglacial landforms in the southern Alps, both in the form of active and inactive landforms (Table 3). At present, climatic conditions favourable for their activity range between 2400 and 2900 m (Guglielmin and Smiraglia, 1997; Seppi et al., 2012; Scotti et al., 2013; Colucci et al., 2016a). Relict forms are generally 400-500 m lower than the active ones, corresponding to a temperature drop of ca. 2.6-3.3 °C (Frauenfelder et al., 2001). Also for this reason, despite the low number of dated rock glaciers in the southern Alps, YD or early Holocene ages have been suggested for most of them. The mean altitude of the front of the YD rock glaciers in the Alps is found at the lowest elevations in the SE fringe of this mountain range, with fronts distributed at 1777 m (Colucci et al., 2016a).

Table 3

In the SW Alps (France), more than 1000 relict rock glaciers have been identified by Bornet et al. (2014). The mean altitude of their fronts is around 2260 m, whereas a few of them (7) reach altitudes lower than 1800 m, with the lowest one lying at 1440 m. It is probable that most of those relict rock glaciers are of YD age, though no absolute dating is yet available. A YD age is in good agreement with the general scheme of Late Glacial rock glaciers formation further North (45°N) proposed by Cossart et al. (2010). Nevertheless, the authors emphasize the poor climatic significance of those permafrost-related landforms, because they developed at higher altitude than the local YD moraine and disconnected from talus slopes, therefore their extent has probably been limited by the availability of sediment sources. These conclusions are consistent with several geoelectrical resistivity soundings carried out on rock glaciers in different sectors, from the Col du Lautaret (near Briançon village) to the Mercantour area (data partly reviewed by Ribolini and Fabre, 2006). Overall, looking at the minimum altitude of the relict rock glacier fronts in the southern Alps (Guglielmin and Smiraglia, 1997; Seppi et al., 2012; Scotti et al., 2013; Bornet et al., 2014; Colucci et al., 2016a), there is a clear lowering trend from the western Alps (2260-2340 m) to the central Alps (2170-2281 m) and the southeastern Alps (1778 m) (Figure 5). Although the Maritime Alps do not follow this scheme, this overall pattern in rock glacier elevations follows the ELA depression observed in the eastern Alps compared to the western Alps at the LGM peak (Kuhlemann et al., 2008).

Figure 5

Holocene

According to Finsinger and Tinner (2006), during the Early Holocene two cooling phases (8.8-7.3 and 6.1-5.2 ka cal BP) occurred in the western Alps. The cold event at 8.2 ka cal BP has also been documented in the eastern Alps with ca. 3°C below the early Holocene thermal maximum (Ilyashuk et al., 2011), which has been reconstructed between 9 and 5 ka BP with an estimated mean July temperature 1-2°C warmer than the recent pre-industrial period (Samartin et al., 2017). Since 4.5 ka BP, with the exception of the end of the 20th century, a progressive cooling trend occurred (Ilyashuk et al., 2011). During the Holocene glaciers showed several advance stages that were not synchronous within the Alps, although most of the glaciers at the onset of the Holocene recorded a rapid shrinkage to a size generally smaller than their late 20th century size until 3.3 ka cal BP (Ivy-Ochs et al., 2009). Subsequently, glaciers showed a widespread re-advance especially between 3.0 and 2.6 ka cal BP and between 1.4 and 1.2 ka cal BP with the final major last re-advance during the LIA between 0.45 and 0.1 ka cal BP (Ivy-Ochs et al., 2009).

With regards to the chronology of Holocene periglacial activity in the southern Alps, there are only a few available ages for periglacial and permafrost landforms. A cooler and dryer period ca. 4.5 ka cal BP must have favoured permafrost aggradation and rock glaciers development in the southern Alps (Ilyashuk et al., 2011). The re-calibraton with OxCal 4.3.1 (Bronk Ramsey, 2009) of the available radiocarbon ages of inactive rock glaciers (Calderoni et al., 1998; Dramis et al., 2003) in the central and western Italian Alps using the IntCal 13 dataset, suggests that the period of maximum activity occurred between 2720 and 2850 cal yr BP with some older exceptions (all younger than 5900 cal yr BP). On the other hand, based on the existence of a paleosoil buried by the rock glacier front, Calderoni et al. (1998) and Scapozza et al. (2010) showed evidence of their activity between 929 and 1374 cal yr BP, a similar age to that reported by a Salix sp. leaf trapped within the ice of the Foscagno active rock glacier dated at 919-961 cal yr BP (Stenni et al., 2007). Recently, Krainer et al. (2015) found that the age of ice within an active rock glacier in the central Italian Alps ranged between 8960 and 2240 cal yr BP at 23.5 and 2.8 m depth, respectively, showing evidence that the ice preserved within the rock glaciers can be useful paleoclimate archives. The investigation in fact shows that frozen material, which had existed during most of the Holocene, is now thawing; this is strong evidence that temperatures during the past about 9 ka have never been as high as they are today.

LIA and present-day

Widespread wetter and cooler condition over the Alps promoted a well-documented glacier advance during the LIA (~1400-1860 AD; Ivy-Ochs et al., 2009). After the LIA, the warming trend in the Alps has been about twice the global trend over the time period 1906-2005 (1.4°C; Brunetti et al., 2009) and in the SE Alps temperature at 2200 m increased by about 1.7°C from 1851 to 2012 (Colucci and Guglielmin, 2015) leading to a 96% reduction in volume of glacial bodies (Colucci and Žebre, 2016). In the French western Maritime Alps, geophysical investigations provided resistivity values coherent with permafrost existence at about 2550 m in the Mercantour sector (Evin and Fabre, 1990). In the Ubaye-Queyras sector, rock glaciers below 2400 m lack internal ice, while above 2500 m permafrost starts to be common (e.g. Marinet rock glacier; Evin and Fabre, 1990; Evin et al., 1990; Assier et al., 1996; Smiraglia et al., 1996; Ribolini et al., 2010). This indicates that 2500 m represents here the modern limit for the existence of active rock glaciers. In the Lautaret and Mt. Viso areas, active rock glaciers were found at 2400-2450 m (Evin, 1991; Francou and Reynaud, 1992; Bodin et al., 2009). According to Bornet et al. (2014) more than 50% of the active and inactive rock glaciers (following Barsch, 1996) in the SW French Alps, from a total number of 780 inventoried landforms, are located above 2600 m and 51% of them are lying on N and NW facing slopes. They cover 2% of the total surface (50 km²) and mostly develop between 2100 and 3000 m. Most of the active rock glaciers in the southern French Alps are smaller than 0.1 km², whereas the relict ones are generally larger (up to a few km²). More than 80% of the active and inactive rock glaciers are probably located below the regional mean annual 0°C-isotherm, raising the question of their stability under climate change (Bodin et al., 2015). This probable imbalance with the present climatic conditions may be responsible for the fast flow observed on some very active rock glaciers (Delaloye et al., 2008), and may also have led to the collapse of the Bérard rock glacier during summer 2006 (Bodin et al., 2017).

In the western Italian Alps active rock glaciers reach with their fronts a mean elevation of 2647 m although a few active rock glaciers appear sporadically between 2000 m and 2300 m. In the Central Italian Alps the mean elevation reached by the active rock glaciers is lower (2526 m) but ice was found within rock glaciers locally also at very low elevation in ventilated talus (1000 m; Guglielmin, unpublished data). As well known the lower altitudinal limit of permafrost based on the elevation of the active rock glaciers, although investigated through geophysical methods, can be not representative of the real regional permafrost; in fact, the lower limit can decrease up to 200 m due to the local density-driven air flows (Balch, 1900). Several rock glaciers have developed since the LIA in the Adamello-Presanella area (eastern Italian Alps) where the current mean elevation of the fronts of active rock glaciers is 2527 m, around 400 m above the limit of the relict ones and where patches of sporadic permafrost can reach an elevation of 2200 m (Baroni et al., 2004). Only a few active rock glaciers occur in the western Alps within LIA deglaciated areas; this is the case of an active rock glacier in the Dolomites (SE Alps) at about 2400 m forming in an area previously occupied by the tongue of a LIA glacier and now dominated by periglacial processes. This study provides an example of how periglacial processes are replacing glacial processes under present climate change during the paraglacial stage (Seppi et al., 2014). Recently, possible sporadic permafrost in an area occupied by two protalus rampart have been deduced based on BTS measurements in the Carnic Prealps at 2258 m (Colucci et al., 2016a). Nine active protalus and pronival ramparts, located in front of permanent snow/firn bodies and ice patches, have been also reported in the Julian Alps and in the Kamnik-Savinja Alps at an average altitude of 2158 m (Colucci, 2016).

Although in the Alps permafrost is limited to the highest and driest areas, ground ice is also reported from several caves (Figure 6). Ice caves are generally widespread in high elevated karstic environments of the Alps, mainly in the central and eastern parts. The real distribution and size of ground ice in Alpine caves is still very uncertain, although recently in the southeastern Italian Alps 1100 ice caves have been reported. Here, they are generally located between 1500 and 2200 m and their distribution is well-correlated with MAAT and altitude, being significantly more frequent for MAAT lower than 2 °C and 5°C; therefore, they are placed within the periglacial domain characterized by high precipitations (Colucci et al., 2016b). In Slovenia, 551 ice caves have been reported by Mihevc (2008). According to the Slovenian Cave Registry (2014), approximately 520 cryo caves, of which ca. 100 ice caves, can be found in the Slovenian Alps. Most of them have the entrance between 1000 and 2400 m (Mihevc, 2008). The radiocarbon age of an insect found in an ice core extracted in 2003 from the Abisso sul Margine dell'alto Bregai (central Alps) reported an age of 185 ± 30 cal yr BP (Citterio et al., 2005). In the cave *Ledena jama pri Planini Viševnik* (eastern Alps), an ice core was recently extracted and larch trapped in the ice yielded a radiocarbon age of 300 ± 45 cal yr BP (Staut et al., 2016). In the light of accelerated climate change, other important paleoclimate archives exploitable from ice caves are represented by *in-situ* coarse cryogenic cave carbonates, recently discovered for the first time in a vertical ice outcrop of a cave in the southeastern Alps (Colucci et al., 2017).

Figure 6

4.3 Italian Peninsula

The Italian Peninsula lies between the Tyrrhenian Sea on the west, the Ionian Sea on the south, and the Adriatic Sea on the east. It includes large plains of northern Italy (Po Plain and Venetian Plain) and extends over more than 1200 km crossing 8° of latitudes between 38 and 46 ºN. The Apennine Mountains run along the entire peninsula and culminate in the Gran Sasso d'Italia (2912) and La Majella (2793 m). Geologically, the Apennines are quite homogeneous with a clear dominance of Mesozoic carbonates although sandstones, shales and flysch are also found, especially in the northern Apennines.

In the Italian Peninsula, from the Late Pliocene/Early Pleistocene transition the climate was characterized by rapid succession between xeric cool (glacials) and humid warm (interglacials) phases (Bertini, 2010). These alternating climatic conditions favoured the existence of "warm" steppes or coniferous forests alternated with subtropical to warm-temperate deciduous forests. During these periods, the paleogeography of the Italian coasts changed significantly due to sea level variations; indeed, the coastline of Adriatic Sea moved hundreds of km southward during the LGM and did not reach the current position until ca. 7 ka (Lambeck et al., 2011; Anzidei et al., 2014).

Last Glaciation

In the Apennines valley glaciers were only present on the highest mountains, with the maximum glacial expansion dated at 27.2 ± 0.9 ka cal BP in the Campo Imperatore valley and a rapid glacial retreat at around 21 ka cal BP (Giraudi and Frezzotti, 1997). Subsequently, lake level oscillations of Fucino Lake showed an aridity period between 20 and 17 ka cal BP and between 15 ka cal BP and the Early Holocene (Giraudi, 1998). More recently, at Campo Felice (central Apeninnes) Giraudi (2012) suggested that the maximum glacial advance was dated earlier than in other areas (33-27 ka cal BP) driven by a relatively milder and wetter period. Kuhlemann et al. (2008) modelled the air circulation and the ELA of the glaciers in the Mediterranean area indicating possible conditions for glaciers occurrence also on Mt. Pollino in Calabria and on Mt. Etna in Sicily.

Several landforms were related to periglacial or permafrost conditions during the LGM or during the deglaciation process, particularly during the OD in Italy (Figure 7). In the northern Apennines only a couple of relict rock glaciers described by Chelli and Tellini (2002) were associated to

LGM period on the northern slopes of the highest mountains of this belt between Mt. Cusna (2121 m) and Mt. Cimone (2165 m). Rock glaciers from this period are more frequent in the central Apennines, where 38 relict rock glaciers were identified by Giraudi (2002) almost all located along the eastern side of the central Apennines. They are mainly present on the Gran Sasso Massif (2912 m), the Maiella Massif (2793 m) and Mt. Velino (2486 m). In the southern Apennines only one relict rock glacier has been detected on the northern slope of Mt. Pollino (2267 m).

Figure 7

Block streams from this stage are located around Mt. Beigua (1287 m), few km far from the Ligurian Sea. These landforms are located between 900 and 1100 m on all the slopes and are characterized by an open-work texture in the upper 1.5 m thickness with angular or subangular blocks and frequent vertical dipping, or imbrication of the boulders (Firpo et al., 2006). Remarkably, none has any rock cliff or free face at their head and they show generally the smallest blocks in their frontal part. These landforms are all shaped on the same bedrock (serpentine and metabasalts) and they could have formed at a maximum age of 33 ka (Firpo et al., 2006). Moreover, in the upper parts of the Mt. Beigua thors and blockfields are also present, suggesting intense periglacial conditions (Firpo et al., 2006).

In the Apennines stratified scree deposits (gréeze litée) are the most widespread and studied periglacial sedimentary archives (Table 4). They were found in different localities of the Northern Apennines, such as Corniglio and Bratica Valley (Chelli and Tellini, 2002), Monte Prampa (Bernini et al., 1978), Alpi Apuane (Federici, 1981) and, above all, in the central Apennines (Coltorti and Dramis, 1988; Coltorti et al., 1983) and in different valleys in the Marche region. In the central Apennines stratified screes were described in the late 1970's mainly along the Adriatic side of Umbria-Marche Apennines (Coltorti et al., 1979, Castiglioni et al., 1979; Boenzi, 1980), where these deposits are particularly widespread and connected with outcrops of limestone and marly limestone bedrock (Coltorti et al., 1983). These deposits are documented at different elevations from the coast to more than 2000 m (Mt. Conero) generally at the foot of rectified slopes or down-slope ancient nivation hollows. Features identified as stratified scree were also found southward in Abruzzo (Maielama valley; Frezzotti and Giraudi, 1992; Sulmona valley, Miccadei et al., 1999), Molise (Coltorti, 1983; Scarciglia, 2000) and Puglia regions (Boenzi et al., 1977). In some cases, they were also found close to the coast of the western side of the Apennines on the Palmaria Island, close to La Spezia (Chelli and Pappalardo, 2006) and on the Sorrento Peninsula (Brancaccio, 1968). The southernmost stratified screes were found in Calabria (Mt. Pollino and Mt. Sila down to 500 m; Boenzi and Palmentola, 1975; Praia a Mare and Mormanno down to 700 m; Robustelli and Scarciglia, 2006). In most of the cases, they were probably associated to seasonal frost conditions, with no permafrost regime.

Table 4

Finally, cryoturbations have been also documented on soils of Mt. Beigua at 650 m (Rellini et al., 2014) and at even lower elevations in the Cilento hills (at 30 m asl; Scarciglia et al., 2003) where severe winter frost was hypothesized during the Last Glaciation. Only in one case, in the Sila Mountain at 1350 m, a fossil sand wedge, an ice wedge cast and some other cryogenic features within the soils were described by Dimase (2006).

Holocene

According to Giraudi et al. (2011), the onset of the Holocene in the Italian Peninsula and in Sicily was characterized by dramatic environmental and climatic changes. Between ca. 9.5 and ca. 6-5.5 ka cal BP the climate was generally warmer and wetter than present, although in some records the wettest period seems to have occurred later at ca. 6-5 ka cal BP. Subsequently, a progressive cooling with higher precipitation variability started at 4.5 ka cal BP, with wet periods recorded at 2.8-2.6, 1.4-1.2 and 0.5-0.3 ka cal BP and an aridity peak between 3.7 and 3.2 ka cal BP (Calderoni et al., 1998; Giraudi, 1998). No dated evidence of glacial advances has been documented in the Italian Peninsula during the Holocene except for Gran Sasso area, where the Calderone glacier extends now over a few ha (Hughes et al., 2006a). Here, the glacier is thought to have persisted in this cirque throughout the Early Holocene, although the cirque was probably ice-free between ca. 4.3 and 3.9 ka cal BP (Giraudi, 2004; Hughes, 2006a). Subsequently this glacier experienced several phases of expansion, namely at ca.: 2.9- 2.7, 1.4-1.3 and 0.64-0.58 ka cal BP (Giraudi, 2003, 2004; Hughes, 2006a).

The ages for periglacial landforms developed during the Holocene are rare and limited to few rock glaciers in the central-eastern Apennines. According to these ages, it is possible that permafrost existed during the Early Holocene at elevations higher than 1900-2000 m and started to degrade at ca. 7 ka cal BP (Dramis et al., 2003). Nevertheless, a possible new phase of permafrost aggradation may have occurred at high elevations during the aridity phase that took place between 3.7 and 3.2 ka cal BP (Dramis et al., 2003).

LIA and present-day

It is reasonable to hypothesize that colder conditions prevailing during the LIA favoured the development of some sporadic permafrost at elevations above 2200 m in the central Apennine (Mt. Gran Sasso, Maiella, Mt. Velino). A still active rock glacier is located on the north face of Mt. Amaro, with its front reaching 2522 m (Figure 8) (La Maiella massif, Dramis and Kotarba, 1992). As documented by BTS measurements, at elevations higher than 2400 m in the same massif (Bisci et al., 2003) and locally lower in Mt. Velino permafrost occurrence was possible, until at least 15 years ago (Guglielmin, unpublished data).

Figure 8

4.4 Balkan Peninsula

The Balkan Peninsula (also referred to as Balkans) is situated in south-eastern Europe and is bounded by the Adriatic, Ionian, Aegean and Black Seas. Situated between the Eurasian and African lithospheric plates, the Balkan Peninsula constitutes a mosaic of several mountain systems: Dinaric, Balkan, Macedonia–Rhodopes and Pindus, separated by internal tectonic depressions and major valleys. This internal topographic variability, coupled with general orientation of the mountain ranges, NNW to SSE Dinaric and Pindus and W to E for Bulgarian Mountains, generates a wide diversity of local geographical conditions, reflected in the presence and evolution of permafrost through time. The highest peak of the Balkans is Musala (2925 m), found in the Rila Mountain.

The geology of the Balkan Peninsula is very diverse. For instance, the Dinaric Alps are dominated by up to 8000 m thick Mesozoic carbonates (Jurasic limestone, Triasic dolomite and Werfen sandstones and schists, and Wengen sandstones). The Šar Mountains and Mount Korab on the eastern Albania border consist of polymetamorphic metasedimentary sequence with Hercynian granitoid intrusions in its central part. The major Bulgarian mountains are composed largely of Precambrian and Palaeozoic crystalline rocks (crystalline schist, gneiss, amphibolites, metamorphosed sedimentary rocks etc.) intruded by the Rila-West Rodophian batholith (Sinnyovsky, 2015), in some areas also by crystalline limestone and marble. In Greece the Pindus Mountains and Olympos Massif are dominated by Mezozoic (Lower Jurassic - Upper Cretaceous) sedimentary rock, especially limestone, dolomite and breccia, with ophiolithic rocks also present in some areas, such as at Mount Smolikas and Mount Vasilitsa.

The Balkan Peninsula is one of the Mediterranean regions where information on present and past permafrost conditions is very general and has a speculative character (King and Akermann, 1993; Dobiński, 2005). There is evidence that "isolated patches of permafrost" (Brown et al., 2001) are still present in the high mountain environment of the Balkan Peninsula. However, apart from ice caves (Bočić et al., 2014; Buzjak et al., 2014; Kern et al., 2006a) no other active permafrost features have been recognised in the study area. Relict permafrost features in the Balkan Peninsula have been recognised, although only recorded in detail in the Pindus, Prokletije, Rila and Pirin Mountains, where rock glacier inventories have been compiled (Palmentola et al., 1995; Hughes et al., 2003; Palmentola and Stamatopoulos, 2004; Dimitrov and Gikov, 2012; Gikov and Dimitrov, 2011; Magori et al., 2017). Sand wedges and cryoturbations present in the Pannonian Basin indicate a deep seasonal frost and perhaps isolated patches of sporadic permafrost (Ruszkiczay-Rüdiger and Kern, 2015) in the SW lowlands of the Pannonian Basin and in Getic Basin (Romanian Plain) (Frenzel et al., 1992) during the Pleistocene. However, there is no evidence of continuous permafrost as suggested by some previous studies (e.g., Poser, 1948; Velichko, 1982). At the LGM only the highest mountains of the Central Balkans were located in the discontinuous permafrost zone (Van Vliet-Lanoë and Hallegouët, 2001). Hence, it follows that rock glaciers are the main source of paleo-environmental information to infer past permafrost conditions in low-latitude areas of the Balkan Peninsula, while the ice caves are the best indicators for evaluating the present permafrost distribution in this area. Other elements of periglacial morphogenesis associated with permafrost conditions, such as large and medium size patterned ground and block streams, are very little-studied in the Balkan Peninsula. As is other parts of the Mediterranean region, thick Pleistocene stratified scree deposits are present across the Balkans - especially in the uplands - and attest to active frost action under cold climate conditions. These sediments can be well preserved in limestone caves and rockshelters (e.g. Bailey and Woodward, 1997; Morley and Woodward, 2011), but establishing their paleoclimatic significance is not always straighforward (Woodward and Goldberg, 2001).

Last Glaciation and Deglaciation

The Last Glaciation in the Balkan Peninsula was characterised by the presence of ice fields and valley glaciers over some of the highest mountain massifs (e.g., Milivojević et al., 2008; Hughes et al., 2010, 2011; Kuhlemann et al., 2009, 2013; Žebre and Stepišnik, 2014; Sinnyovsky, 2015; Žebre et al., 2016). Late Pleistocene glaciers reached their maximum extents before the LGM in Pindus Mountains and this is recorded indirectly in the fluvial record downstream where glaciofluvial sediments have been dated at 25-30 ka (Lewin et al., 1991; Woodward et al. 1995; 2008; Hughes et al 2006b). In the Peloponnese, pre-LGM moraines have recently been dated to 30-40 ka using ^{36}Cl exposure dating (Pope et al., 2017). In the Rila Mountains, Bulgaria, it is most likely that two glacial advances occurred, one at the beginning (25-23 ka) and one at the end (16-18 ka) of the global LGM (Kuhlemann et al., 2013). However, the exact timing of the local LGM over the Balkan Peninsula is still a matter of debate with some of the best evidence coming from the fluvial record in NW Greece (Woodward et al., 2008). In the Dinaric Alps, relatively high moisture (Hughes et al., 2010; Žebre and Stepišnik, 2014) was more favourable for glacier development in the highest massifs, meanwhile the lowlands were likely characterised by periglacial environment (Table 5). Almost no research on permafrost features from the Pleistocene period has been presented until now from this area. According to the interpretation by van Vliet-Lanoe et al. (2004) and Ruszkiczay-Rüdiger and Kern (2015), the Dinaric Alps area (at least its NW part) probably hosted discontinuous permafrost, bounded by a seasonal frost or even isolated patches of sporadic permafrost in the Pannonian basin to the NE. In fact, ice wedges and cryoturbations have been found in till in the Lovćen Mountain (Liedtke, 1962), supporting the idea of the permafrost presence in the Dinaric Alps. The presence of tors on the highest peaks indicates intensive periglacial processes in the ice-free areas above the local Equilibrium Line Altitude (ELA).

Table 5

In contrast, in the continental, more arid part of the Balkan Peninsula relict rock glaciers, and also block streams and patterned ground are widespread (Figure 9). Early periglacial studies from the Bulgarian mountains revealed a broad presence of scree deposits, block streams and blockfields (Stoyanov and Gachev, 2012). In addition, 27 and 55 rock glaciers were identified in Rila (Gikov and Dimitrov, 2011) and Pirin Mountains (Dimitrov and Gikov, 2012), respectively, though Magori et al. (2017) recently found new field evidence of the existence of 122 units in both massifs. Rock glaciers are distributed at elevations between 2080 and 2600 m, with a mean altitude of 2340 m in Rila Mountains and 2390 m in Pirin Mountains (Magori et al., 2017). They are considered to be from the end of Pleistocene and early Holocene (Dimitrov and Gikov, 2012; Gikov and Dimitrov, 2011; Kuhlemann et al., 2013), although the spatial relation between different altitudinal groups of landforms suggest other timing interpretation. In the Šar Mountains several rock glaciers, most of them in direct spatial connection with latero-frontal moraines, were recognized above 1700 m by Kuhlemann et al. (2009). Several generations of rock glaciers appear on the northern side of the ridge Bistra (2651 m) – Jezerski (2604 m) between 1750 and 2550 m. A similar situation is found on the north and north-western slope of Mount Korab, where relict rock glaciers are present at altitudes between 1480 and 2600 m. Relict rock glaciers are widespread also in the Pindus and other mountains in Greece. These features have been identified on Mount Parnassus (Pechoux, 1970) in central Greece, and Mount Tymphi (Hughes et al., 2003), Mount Smolikas (Hughes et al., 2006c) and Mount Lakmos-Peristeri (Palmentola and Stamatopoulos, 2004) in northwest Greece. The relict rock glaciers of the Pindus Mountains have wide-ranging elevations and occupy cirques between 1330 and 2300 m (Hughes et al., 2003; Palmentola and Stamatopoulos, 2004; Hughes et al., 2006a) and are likely to represent different generations of rock glacier. Hughes et al. (2003) argued that the rock glaciers of Mount Tymphi must belong to the LGM and formed as climate became colder and drier forcing cirque glaciers to retreat and become overwhelmed by debris. Further research is necessary to test the ages of the rock glaciers in the Balkans.

Figure 9

Besides rock glaciers other permafrost features can also be identified in the inland part of the Balkans (Figure 10). On the Vitosha Mountain, relict stone circles, tors and large blockfields were identified at the elevation range of 1850-1900 m. Some of the largest block streams in Europe, called Zlatnite Mostove (Golden Bridges) and Golyamata Gramada (Big Pile) are also present in this area. The first is 2.2 km long and 150 m wide, descending from 1800 to 1410 m, and the latter 1 km long and 300 m wide at the altitude between 1900 and 1550 m. Block streams are present in the Stara Planina (Balkan) Mountains, descending even down to 930 m in the northern slopes and 1130 m in the southern slopes. They can be found also in the granite part of Rila and Pirin Mountains where they descend down to 1500 m, and in Osogovo Mountain, where they reach 1700 m. Patterned ground and frost-shattered features were noted on Mount Tymphi by Hagedorn

(1969) and Hughes (2004). Separating active or recent periglacial activity from Pleistocene features is difficult for many of the landforms on the highest mountains, although the lower periglacial forms must be Pleistocene in age.

It is certainly true that some rock glaciers form during deglaciation when glacier retreat causes glaciers to become overloaded by debris and morphing into ice-cored rock glaciers (Morris and Olyphant, 1990). It is also true that rock glaciers are most likely to form when climate is cold and dry (Haeberli, 1985). The period characterised by "deglaciation" depends on the timing of the local LGM. In many parts of the world this predates the global LGM, with cold and dry conditions at the global LGM causing glacier retreat (Hughes et al., 2013).

Figure 10

In the Pindus Mountains, on Mount Smolikas (2632 m), a rock glacier occurs at much higher altitude (>2100 m) than on neighbouring Mount Tymphi (1800-2200 m) (Table 6). This led Hughes et al. (2006a) to argue for a YD age for the Smolikas rock glacier. Based on climate modelling, Hughes et al. (2006a) argued that this feature and the associated last ever glaciers on Mount Smolikas could not be Holocene in age. On Mount Tymphi Late-glacial rock glaciers are absent because the level of permafrost was above the cirque floors and consequently, rock glaciers only formed on higher mountains such as Mount Smolikas. On Mount Lakmos-Peristeri, just 25 km southeast of Mount Tymphi, Palmentola and Stamatopoulos (2004) argued for an OD age for the numerous relict rock glaciers in the cirques of these mountains. This was based on earlier suggestions made for similar relict rock glaciers in the Prokletije Mountains, where the study of 16 inactive rock glaciers located above 1700 m was presented by Palmentola et al. (1995). Formation of several relict rock glaciers in the inland Dinaric Alps of Bosnia and Herzegovina and Montenegro (never studied before, but clearly recognized from different satellite images), indicating past permafrost conditions at elevations above 1650 m, may also be related to one of the dry and cold periods of the late Pleistocene. This is in accordance with the recently compiled rock glacier inventory in the nearby SE Alps (Colucci et al., 2016a) where the formation of rock glaciers was related to the YD cold phase. However, in Greece Hughes et al. (2003) argued that similar elevation rock glaciers were older, forming during the LGM. The OD ELA in Šar Mountains was calculated to 2200-2350 m (Kuhlemann et al., 2009). Thus, it is likely that rock glaciers present in the altitudinal belt 1950-2060 m on the northern side of the ridge Bistra (2651 m) – Jezerski (2604 m) belong to this period. During YD, the ELA was placed at 2300-2400, so the rock glacier at 2260-2550 m could belong to this period. Relict rock glaciers on Mount Korab located above 2400 m appear to be fresher (without vegetation) and could belong also to one of these two periods. In the Rila Mountains, rock glaciers above 2200 m and below local ELA are likely to be formed in the deglaciation period. During the deglaciation period, the stadial moraines at 1700 m, 1850 m, and 2000 m point to gradual retreat of the glaciers in the Osogovo Mountain (Milevski, 2008). Embryonic rock glaciers at 1850 and 2030 m in the cirque below the Ruen Peak (2251 m) were likely active during this period. However, none of the rock glaciers in the Balkans have been dated and therefore the competing geochronological hypotheses proposed by Palmentola et al. (1995), Hughes et al. (2003) and Palmentola and Stamatopoulos (2004) remain to be tested.

Table 6

Holocene

Holocene cooling periods in 8.2 ka cal BP (Tonkov et al., 2016), 7.5 ka cal BP (Stefanova et al., 2006) and between 6.7 and 5 ka cal BP (Bozilova and Tonkov, 2000) based on pollen records were reported for the Rila and Pirin Mountains. It is likely that these cooling events triggered the formation of some newly rock glaciers, while the highest lying rock glaciers continued their existence. Rock glaciers in the Rila and Pirin Mountains situated above 2500 m were probably active also during Holocene. One example is the rock glacier on the northern slope of Polezhan Peak (2851 m) with the front at 2490 m where at least three rock glacier generations can be recognised. Also in the Šar Mountains, on the northern slopes of the Bistra (2651 m) – Jezerski (2604 m) ridge, rock glaciers above ~2300 m likely persisted in their active mode during the Holocene, as suggested by Kuhlemann et al. (2009).

Some of the cirque and valley glaciers that were present during the YD likely persisted to Holocene in some areas of the Balkans. Calcite cements in cirque moraines, which provide minimum ages for the moraines, yield early Holocene ages in central Montenegro (Hughes et al., 2011). On Mount Olympus, in the north-facing cirque of Megali Kazania, large moraines were considered to be Holocene Neo-glacial moraines by Smith et al. (1997). This possibility has recently been advocated by Styllas et al. (2015). This would seem plausible given the presence of small glaciers today further north in the Balkans. It is possible that this was also the case in other inland mountains in the Balkan Peninsula, although there is very little known about the Holocene glacial history. Periglacial forms related to these small glaciers, such as nivation hollows, are also poorly understood. However, given the warmer conditions at the Holocene optimum it is arguable that smaller glaciers and snow patches nowadays present in Pirin, Prokletije and Durmitor Mountains (Gachev et al., 2016) would have been less than today.

LIA

Owing to cooling conditions during the LIA there were numerous small cirque glaciers (Hughes, 2010) in the highest massifs of the Balkan Peninsula. There would have also been many more perennial snow fields. Nival processes would have therefore been a significant geomorphological agent across the mountains in the Balkan Peninsula at this time. Rock glaciers above 2350 m in the Rila and Pirin Mountains probably reactivated during the LIA period.

Present-day

Because of relatively low altitude and low latitude, the mountain permafrost in the Balkan Peninsula is nearly inexistent in recent climate. Permafrost may be present in some niche localities on the highest peaks, and this possibility was highlighted in Dobinski (2005) who suggested that permafrost conditions may exist above ~2700 m on Mount Olympus (2912 m), and above ~2350 m on Rila Mountain (2925 m). This idea is partially supported by Brown et al. (2001) modelling results, suggesting that sporadic permafrost is present only in the highest peaks of the southern Dinaric Alps, Šar Mountains, and Rila and Pirin Mountains. In fact, Milivojević et al. (2008) mapped two active rock glaciers in the central Prokletije Mountains. According to Hughes (2009) these features may not be actively moving because they lack the clear lobate form. However, they may be described as patches of sporadic permafrost. Perennial snow fields and modern nival forms are widespread across the highest mountains of the Balkan Peninsula. For example, Styllas et al. (2016) reported perennial snow and ice masses in the north-facing cirque of Megali Kazania (Olympus Mountain). Apart from this 16 glacierets and small cirque glaciers exist in the Durmitor, Prokletije and Pirin Mountains (Hughes, 2008; Djurović, 2013; Milivojević et al., 2008; Hughes, 2009; Gachev and Stoyanov, 2012; Gachev et al., 2016). Some of these ice masses are debris-covered and exhibit similar geomorphology to the buried ice patches of the Corral

Veleta site in the Sierra Nevada, Spain (Gómez-Ortiz et al., 2001). Therefore, the possible presence of patches of sporadic permafrost below these perennial snow patches can not be excluded.

Instead ice caves are permafrost phenomena in carbonate massifs of the Balkan Peninsula which often exist at elevations where outside MAAT is well above 0 °C (Colucci et al., 2016b) owing to karstic topography which functions as cold air trap and thus promote the survival of permanent snow and ice. Majority of caves containing permanent frozen materials, the so-called cryo caves (sensu Colucci et al., 2016b), occur at elevations above 800-1000 m in the Dinaric Alps (Bočić et al., 2014; Kern et al., 2006b; Zupan Hajna, 2016) and mountains in Macedonia (Temovski, 2016), although some cryo caves in Slovenia were reported from lower altitudes (Mihevc, 2008). In the Dinaric karst of Slovenia about 100 cryo caves (Slovenian Cave Registry, 2014) and in the northern Velebit Mountain in Croatia 150 cryo caves (Buzjak et al., 2016) were explored by speleologists. Moreover, active patterned ground likely related to seasonal frost heaving was reported from some Slovenian caves by Zupan Hajna (2007), Mihevc (2009) and Obu et al. (2018).

Nevertheless, periglacial activity is a significant geomorphological agent in the highest areas of the Balkan Peninsula. In the Rila and Pirin Mountains, periglacial processes are active above 1900-2000 m (Stoyanov and Gachev, 2012).

4.5 Anatolian Peninsula

Located between one of the world's most seismically active areas, the Anatolian Peninsula constitutes a relatively small orogenic plateau that is bounded by the Mediterranean, Aegean, Marmara and Black Seas. Despite its modest average elevations just ~1200 m and low overall exhumation, the Anatolian Peninsula is a first-order morphotectonic feature that has fundamentally impacted the geologic, geomorphic, and climatic evolution of the Eastern Mediterranean (Çiner et al., 2013). The tectonically active boundaries of the plateau delineate the Anatolian plate, which has been extruding toward the west with respect to Eurasia since the Miocene as the result of extension in the Aegean and the Arabia-Eurasia collision (Şengör and Yılmaz, 1981; Bozkurt, 2001). The Eastern Black Sea Mountains (also known as the Pontides) and the Taurus Mountain Range bound the northern and southern flanks of the plateau respectively and attain elevations more than 3000 m in places. Several paleoglacial valleys with moraines, occasional small glaciers and periglacial features are preserved in these mountains (Palgrave, 1872; Luis, 1944; Kurter, 1991; Çiner, 2003; Sarıkaya et al., 2011; Çalışkan et al., 2012; Sarıkaya and Tekeli, 2012; Bayrakdar et al., 2015; Yavaşlı et al., 2015). In the central parts, high volcanoes such as Ağrı (also known as Ararat, 5137 m), Süphan (4058 m) and Erciyes (3917 m) also contain active glaciers and periglacial features (Penther, 1905; Erinç, 1951; Messerli, 1964; Kurter, 1991; Kesici, 2005; Sarıkaya et al., 2009; Sarıkaya, 2012).

The geology of the Anatolia Peninsula is rather complicated. However, the Taurus Mountains along the Mediterranean coast of Turkey are mainly made up of platform carbonate rocks with occasional presence of ophiolitic rocks emplaced as trust sheets (Monod, 1977). On the other hand, the Eastern Black Sea Mountains along the NE Black Sea coast of Turkey contain mostly quartz bearing lithologies composed of plutonic and volcanic rocks. The volcanoes scattered in the central parts of Anatolian Peninsula are generally andesitic to rhyolitic in composition.

Late Pleistocene glacial deposits in Turkey are now well described and constitute one of the best-dated records of its kind in the world (Hughes and Woodward, 2017) (see a review of glacial landforms and TCN dating chronologies by Sarıkaya and Çiner, 2015, 2017). On the other hand, periglacial landforms are only scarcely described, often in local journals and in Turkish and lack quantitative age results. Here, we present an overview of the distribution of mountains with periglacial landforms based mainly on literature but also on field observations on periglacial processes and landforms formed since the Last Glaciation (Table 7).

Table 7

Last Glaciation

There are no age data on the periglacial landforms related to the Last Glaciation in Turkey. However, moraines in the Taurus Range, particularly in Mt. Sandıras (2295 m) (Sarıkaya et al., 2008), Mt. Akdağ (3016 m) (Sarıkaya et al., 2014), Mt. Dedegöl (2992 m) (Zahno et al., 2009; Çılğın, 2015; Köse et al., 2017), Mt. Aladağlar (3756 m) (Zreda et al., 2011), Mt. Bolkar (3524 m) (Çiner and Sarıkaya, 2017) and Mt. Geyikdağ (2877 m) (Çiner et al., 2015, 2017; Sarıkaya et al., 2017) yield TCN ages that vary between ~50 to 5 ka. To the NW of Turkey, on Başyayla Valley of Mt. Kaçkar, (3937 m) along the Eastern Black Sea Mountain Range, the MIE is reported to be around 56 ka (Reber et al., 2014). It is therefore reasonable to assume that the extent of glacier advances and retreats also controlled the intensity of periglacial activity as well as the development of periglacial landforms in these areas.

The synchronicity of the LGM in the eastern Mediterranean region is under current debate (e.g., Clark et al., 2009; Hughes and Woodward, 2008; Hughes et al., 2013). However, there seems to be a consensus on the rather synchronous LGM extent and timing (centred to around 21 ka) on the Taurus Mountains (e.g., Sarıkaya and Çiner, 2017), Eastern Black Sea Mountains (Akçar et al., 2007, 2008), Lesser Caucasus Mountains (Dede et al., 2017) and the individual mountains in the central Anatolia such as Mt. Uludağ (2543 m) (Akçar et al., 2014, 2017; Zahno et al., 2010) and Mt. Erciyes (Sarıkaya et al., 2009). It is therefore very likely that during the Last Glaciation and at LGM, periglacial processes below LGM snowline, approximately between 2400-2600 m in the Taurus Mountains and 2300-2500 m in the Eastern Black Sea Mountains (Erinç, 1952; Messerli, 1967; Çiner, 2004), were active and permafrost mainly controlled the development of some relict rock glaciers as well as patterned ground features (Figure 11).

Figure 11

Deglaciation

Following the LGM, at around 21 ka a rapid deglaciation is documented throughout the Anatolia Peninsula (e.g., Sarıkaya and Çiner, 2015). As in other periglacial alpine environments (Barsch, 1992; Knight and Harrison, 2008), mountains in Anatolia also experienced increasing rockfall and mass movements, leading to rock glacier instability, formation of outwash plains and sediment release to the rivers. This deglaciation was interrupted by two cold intervals; OD, which was rather long compared to the relatively shorter YD, where glaciers regained momentum and descended down valleys (Çiner et al., 2015; Sarıkaya and Çiner, 2015). These two cold stages were followed by rock glacier development, especially in the Eastern Black Sea Mountains (e.g., Akçar et al., 2007) and in Mt. Mercan (also known as Mt. Munzur; 3463 m) in central Anatolia (Bilgin, 1972). A relict rock glacier terminus that lies around 2100 m was recently TCN dated to

15.7 ± 1.3 ka in the Mt. Karçal (3932 m) of the Lesser Caucasus Mountains (Dede et al., 2017) (Figure 12).

Figure 12

Holocene

The Holocene record of periglacial conditions is not well known in Turkey. Although the existence of glaciers in southeastern Taurus Mountains (e.g., Mt Cilo, 4135 m) (İzbıkak, 1951; Erinç, 1953) and high volcanoes such as Mt. Ararat are known (Sarıkaya, 2012), only few glacial deposits were TCN dated. For instance, on the Erciyes Volcano in central Anatolia, reported Early Holocene (9.3 ± 0.5 ka) and Late Holocene (3.8 ± 0.4 ka) glacial advances suggesting also periglacial conditions surrounding this mountain (Sarıkaya et al., 2009). On the other hand, rock glaciers probably were extensively developed during the Holocene especially in the northern parts of the Anatolia Peninsula. However, the development of rock glaciers is rather restricted in the Taurus Range probably because of the well-developed karst that restrained surface water flow. Nevertheless, rock glaciers were previously described in Mt. Geyikdağ in the central Taurus Mountains (Arpat and Özgül, 1972; Çiner et al., 1999) and recently Çiner et al. (2017) TCN dated a rock glacier to < 6 ka (Fig. 2c). On the other hand, a study carried out on the southern slopes of Mt. Ararat (Avcı, 2007) claimed that during the glacial/periglacial climates of the past, mass movements gave rise to debris on the slopes and blocky colluvium in the valley floor.

LIA

With the exception of Mt. Ararat ice cap that covers around 5.6 km^2 (Sarıkaya, 2012) and Mt. Cilo Uludoruk glacier (<3 km long), only few small glaciers are present today in Turkey, where two third are located in the southeastern Taurus Mountains (Kurter, 1991; Çiner, 2004; Akçar and Schlüchter, 2005; Sarıkaya and Tekeli, 2014). However, glaciers were more numerous and much larger during the LIA in Turkey and in the Mediterranean mountains in general (Hughes, 2014, 2017).

For instance, the highest moraines in Mt. Uludağ were attributed to LIA (Erinç, 1952). More recently Zahno et al. (2010) TCN dated a boulder from the innermost moraine (sample TRU-12 in Zahno et al., 2010) with negligible cosmogenic nuclide concentration supporting a probable LIA advance in Mt Uludağ. Birman (1968) also tentatively proposed a recent age for the formation of this moraine.

In Mt. Aladağlar in the central Taurus Mountains a small glacier covered by rock debris (Altın, 2006; Gürgen et al., 2010) is probably a remnant of LIA glacier development. Even though at least 12 glaciers exist in the Eastern Black Sea Mountains, moraines associated with a LIA advance are reported to be absent in Kavron and Verçenik valleys of Mt. Kaçkar probably because of dry and cold climatic conditions (Doğu et al., 1993; Akçar et al., 2007, 2008; Bayrakdar and Özdemir, 2010; Reber et al., 2014). We can therefore presume that periglacial conditions surrounding these glacial environments were more extensive during LIA compared to today's conditions in these areas with probable development of rock glaciers and protalus lobes.

Present-day

Although no active glaciers exist in Mt. Uludağ in western Turkey, it is probably one of the best-described mountains concerning periglacial landforms. Studies carried out by Erinç (1949, 1957) identified two distinct periglacial levels on Mt. Uludağ. The first one lies between 1900-2300 m and is characterised by garland soils that developed in slopes ranging between 2 and 40°. The second level lies above 2300 m and is mainly represented by stone accumulations, 40-50 cm in diameter, stone stripes and stone circles preserved over quasi-flat (0-10°) surfaces (Figure 12) (Öztürk, 2012). Additionally, Türkeş and Öztürk (2008, 2011) indicated that while several alpine plant species are effective in the formation of garlands, only two types of *Festuca* sp. are effective in the formation of circles. Although much less developed compared to Mt. Uludağ, Bilgin (1960) also reported garland soils at the summit and NE facing slopes of Mt. Kazdağ (~1700 m), in the westernmost part of the Anatolia Peninsula, near the Aegean Sea. This mountain composed of granite and schist was too low to be glaciated during the Last Glaciation but relict block streams were encountered at 1350 m facing northeast suggesting past permafrost conditions (Bilgin, 1960).

Mt. Ilgaz in central Anatolia also shows some periglacial features (Erinç et al., 1961). While large features such as block flows and cyroplanation surfaces are classified as inactive, smaller stone circles (50 cm in diameter) and garlands are thought to be active. Garlands and polygonal soils above 2500 m near Yedigöller Lake on the north-facing slopes of Mt. Esence (also known as Mt. Keşiş; 3549 m) are present (Akkan and Tuncel, 1993). Stone rings at 2650 m on Mt. Mescid (3239 m) are also reported (Atalay, 1983).

On the Eastern Black Sea Range at Mt. Karagöl (3107 m), stone circles ranging from 25 cm to 1 m in diameter and ovoid depressions 1 to 1.5 m in diameter and few decimetres in depth are reported at ca. 1800 and 1900 m, respectively (de Planhol and Bilgin, 1964). Turoğlu (2009) also reported recent solifluction, frost creep and mass movements such as rock falls, talus, talus creeps, rock avalanches and rock flows on the same mountain. In the nearby Mt. Karadağ (3331 m) periglacial features are also known to exist (Bilgin, 1969; Gürgen, 2001). On Elevit and Hacıvanak glacial valleys of Mt. Göller (3328 m) active garlands, stone circles and solifluction terraces related to seasonal frost conditions are best observed on the south facing slopes (Çiçek et al., 2006).

Active rock glaciers are mostly concentrated on the northeastern Turkey and central Anatolia. In a recent study, Gorbunov (2012) detected several active, inactive and fossil rock glaciers in Turkey using satellite imagery. Among around 600 rock glaciers located mainly between 2800-3400 m, of which about 200 are reported to be active, with some reaching 1200-1300 m in length. In some places, fossil rock glaciers descend downvalleys to 2200-2300 m. For instance, in the upper parts of the northern tributary of the Başyayla Valley in the Eastern Black Sea Mountains and in the Mt. Karçal in the Lesser Caucasus Mountains several active rock glaciers were reported (Çalışkan, 2016; Gürgen and Yeşilyurt, 2012; Reber et al., 2014; Dede et al., 2015). In Kavron Valley intensive rock glacier activity is responsible for the destruction of LIA moraines (Akçar and Schlüchter, 2005). In the so-called ski valley of Erciyes Volcano a rock glacier is also described as being active (Sarıkaya et al., 2003; Ünal and Sarıkaya, 2013) (Figure 12b). On the other hand, according to Yeşilyurt and Doğan (2010) several debris-covered glaciers were mistakenly interpreted as rock glaciers on Mt. Munzur.

Mt. Dedegöl is the only site in western Taurus Mountains where rock glaciers are developed (Dellanoy and Maire, 1983). Although some are not active, others situated between 2500 and 2800 m are active with calcareous blocks of 0.5 to 1.5 m in diameter. In Mt. Aladağlar in the central Taurus Mountains, speleologists discovered a karstic cave at ~3400 m, containing 120 m thick ice along a shaft that was probably inherited from LIA (Bayarı et al., 2003; Klimchouk et al., 2006) (Figure 12). This is the only known ice occurrence within a cave system in Turkey. On the Yedigöller plateau of Mt. Aladağlar where the paleo-ice cap is known to have melted at the onset of the Holocene (Zreda et al., 2011), stone stripes are also reported at 3200 m (Figure 12). Rock glaciers also developed in front of the rapidly melting glaciers of Mt. İhtiyar Şahap (also known as Mt. Kavuşşahap; 3650 m) on the southeastern Taurus Mountains (Doğu, 2009; Yeşilyurt et al., 2018).

4.6 Northern Africa

Northern Africa mountainous regions are dominated by the Atlas, a series of ranges and plateaus, running eastwards from the Atlantic at ca. 31º N in coastal Morocco to 35 ºN in the Mediterranean in northern Tunisia, across almost 2000 km. In Morocco, the Atlas is divided in the Middle Atlas (Djebel Bou Naceur, 3340 m), the High Atlas (Djebel Toubkal, 4167 m) and the Anti-Atlas (Djebel Sirwa, 3304 m), not analysed in this paper. The three major massifs in the High Atlas are, from west to east, the Djebel Toubkal massif (4167 m), the Irhil M'Goun massif (4071 m) and the Djebel Ayachi (3751 m). Towards the east, the High Atlas extends to the Saharian Atlas, culminating in the Aurès Massif in Algeria at 35ºN (2324 m). North of the Atlas and close to the Mediterranean coast, lie the Rif (3445 m) and the Algerian Tell (2308 m) divided from the former by the Algerian high plateaus (1729 m). Further south, already in the interior Sahara Desert, lie two other large massifs, with a drier climate and with evidence of relict periglacial phenomena: the Hoggar (2981 m) in southern Algeria at 23ºN, and the Tibesti (3445 m) in northern Chad at 21ºN.

The major geological structures of the mountains in North Africa are fold-thrust belts formed by the collision between the African and the European plates (Gómez et al., 1977; Dewey et. al., 1989) along the plate boundary. The materials in the Rif-Tell Atlas are allochthonous, with the presence of flysch sequences and limestones covering African shield metamorphics and granitoids. The High Atlas – Saharan Atlas is an intracontinental fold-thrust belt in the foreland of the Rif (Arboleya et al., 2004), with autochthonous rocks. Mesozoic limestones prevail in the Middle Atlas, also with Cenozoic basaltic lava flows infilling pre-existing valleys (Arboleya et al., 2004). The High Atlas has three major morphostructures, with the western part showing limestone plateaus, the central zone showing Paleozoic magmatics from the African shield (granites, rhyolites, andesites and trachytes) and the eastern zone, a folded structure of carbonates (Joly, 1962; Hughes et al., 2004). The Saharan Atlas is composed by folded structures affecting mostly Cenozoic limestones.

The Hoggar massif is located in a cratonic setting, showing in its western part, middle-Proterozoic thick meta-sedimentary units and alkaline-peralkaline intrusives, magmatic complexes of basic to ultrabasic rocks and volcanoclastic deposits, andesites to dacites and calc-alkaline batholiths. The central Hoggar is mainly composed of gneisses and schist belts, while the eastern Hoggar shows mainly gneisses and granites (Bertrand and Kaby, 1978). The Tibesti massif is composed of Precambrian crystalline rocks, being formed by a core of intrusive and metamorphic rocks, which is bounded by Paleozoic and sedimentary sequences. These units are partially capped by Tertiary volcanics (Ghuma and Rogers, 1978, Permentier and Oppenheimer, 2007).

The mountains of North Africa mark strong climatic gradients between the Atlantic in the west, the Mediterranean in the north, and the Sahara influence in the south. This impacts essentially the precipitation amounts, varying from over 2,000 mm in the Rif mountains to145 mm in the Hoggar and 11 mm in the Tibesti, and are characterized by regimes, from the Mediterranean type, in the north, to semi-arid and arid regimes as one moves towards the Sahara. Air temperatures show both the effects of continentality and latitude, with striking impacts on the estimated altitudes of the 0 ºC MAAT, which range from slightly above 3000 m in the Rif to about 4900 m in the Hoggar, much higher than the maximum elevations in these ranges (Table 2).

Periglacial features in North African mountains have been described by several authors, but most through a classical naturalistic approach, lacking quantification, sedimentological studies and absolute dating of relict features (Table 8). Most features relate to evidence of frost shattering, talus slopes, stratified slope deposits and more sporadically, to rock glaciers, although lacking detailed observations. At least one case of the latter, in the High Atlas, has been recently reinterpreted as a catastrophic rock slope failure deposit, rather than as a rock glacier (Hughes et al., 2014), which calls for a need for encompassing detailed studies and reinterpretation. Present-day periglacial activity has been vaguely described by the presence of solifluction landforms and ground frost features at high altitude localities, but again the observations lack modern objective assessments and monitoring data are lacking almost everywhere. A recent study by Vieira et al. (2017) based on geomorphological evidence and ground surface temperature data indicates the possible presence of permafrost in the upper reaches of the High Atlas. Both relict and present-day periglacial features reflect the climatic gradient, with the evidence showing an increasing moisture content from SE to NW and a decrease of temperatures.

Table 8

Last Glaciation

No absolute age records exist from periglacial deposits or landforms in North Africa and the only chronological framework for the cold event derives from glacial evidence, which have been especially well-studied in the High Atlas (De Martonne, 1924; Dresch, 1949; Wiche, 1953; Messerli, 1967; Messerli and Winiger, 1992; Chardon and Riser, 1981; Hughes et al., 2004, 2011, 2018; Hannah et al., 2017). In the Djebel Toubkal region, Hughes et al. (2018) have identified three phases of glacier advance or sustained stabilisation, with ages of ca. 50 ka, 22 ka and 12 ka, pointing to a pre-LGM maximum and to the presence of glaciers during both the LGM and the YD. These relate to several valley glaciers that descended in some areas down to 1900 m, with typical glacier lengths of 2 to 9 km. Other ranges showing glacier evidence are the Rif (Mensching, 1960) and the Middle Atlas (Dresch and Raynal, 1953; Raynal et al., 1956; Awad, 1963) in Morocco, and the Tell Atlas (Barbier and Cailleux, 1950) and the Aurès mountains (Ballais, 1981) in Algeria, but with no known absolute age records.

Relict periglacial deposits in northern Africa mountains have been attributed by the different authors mostly to the Last Glaciation (Figure 13). The most frequent reference to periglacial features are probably slope deposits, which vary from head-type deposits to stratified slope deposits and grèzes in some limestone areas. They were identified in the Algerian Tell above 1400 m and in the Aurés Massif above 1800 m (Tihay, 1973), and are widespread in the High Atlas (Chardon and Riser, 1981). We have observed a good outcrop at the Middle Atlas southeast of the Djebel Bou Iblane at 2350 m. Other references to slope processes relate essentially to a diffuse concept of solifluction, which has to be accounted for with great care due to the difficulty relating to the identification of processes, but also to the variability of the interpretation of the concept. As such, relict solifluction features have been identified in the High and Middle Atlas (Awad, 1963) and Messerli (1973) associates frost action in the Hoggar and Tibesti to the presence of slope deposits and filling of hollows, which are present even below 2000 m. Nivation forms (presumably hollows), as a set of geomorphic processes, have been identified by Messerli (1973) in the Hoggar (above 2400 m), in the Tibesti (above 3000 m), by Tihay (1973) in the Aurés above 1800 m and in the Algerian Tell above 1500 m, by Marre and Quinif (1981) in the Algerian High Plateaux above 1500 m, and in the Rif by Mensching (1960). Rock glaciers, which are the single landform identified which is an indicator of the presence of permafrost, have been reported for the Algerian Tell (possible) above 1100 m by Tihay (1973) (an altitude that seems excessively low), for the Middle Atlas between 2100 and 2500 m (Awad, 1963) and for the Rif by Mensching (1960). Nivation forms would have been widespread marginal to glaciated areas of the High Atlas during the Last Glaciation. Patterned ground can be observed on plateau surfaces in the High Atlas, especially between 3000 and 3600 m over the extensive high plateaus of Iouzagner (3502 m) and Tazaghart (3980 m). Hannah et al. (2017) described large polygonal and linear patterned ground (10-20 m wide and up to 1000 m long) as well as extensive covering of thick regolith, blockfields, blockstreams and smaller-scale stone stripes on these high plateaus. The plateaus were covered in ice during the most extensive glacial phase of the Last Glaciation (before ~50 ka based on dating in Hughes et al. 2018) and the periglacial features formed afterwards and are probably still active today given the high altitude of the plateau.

Figure 13

Deglaciation

Few observations exist for the deglaciation phase. However, the rock glaciers reported for the M'Goun Massif by Wiche (1953) which are located inside the valleys well-within the glacier limit can be included in this phase. These features are probably derived from frontal moraines in the cirques, as suggested by Hughes et al. (2006). It is also possible that the rock glaciers described by Awad (1963) for the Middle Atlas, close to the Djebel Bou Naceur, as well as some of the High Atlas rock glaciers described by Dresch (1941) and Wiche (1953) also correspond to this phase.

Present-day

Observations on present-day periglacial processes and landforms are scarce and lack quantitative data. The lack of a common methodology by the different authors limits a regional analysis. Messerli (1973) shows that the Tibesti and Hoggar lack present-day frost activity, a fact that they attribute more to the dryness than to the lack of sub-zero temperatures. At the Aurés Massif, Tihay (1973) indicates that frost action occurs above 2300 m, while Ballais (1981) mentions solifluction between 1300 and 1800 m. For the Algerian Tell, Tihay (1973) lowers these limits, putting frost action above 1500 m. In the Middle Atlas, Dresch and Raynal (1952) and Raynal et al. (1956) indicate active frost shattering. Above 2700 m a marginal periglacial zone starts to be evident in the eastern part of the Bou Naceur massif, with vegetation crescents and very shallow solifluction.

The most consistent reports on present-day periglacial activity in North Africa are from the Djebel Toubkal massif in the High Atlas, the highest peak in north Africa. Periglacial features are widespread (Hughes et al., 2011) and were described for the Central High Atlas by Couvreur (1966), who reported active solifluction above 2200 m (Figure 14). For the western High Atlas, Chardon and Riser (1981) indicate the limit of frost activity to be around 2500 m and considered that frost action dominates the morphogenesis above 3000 m. Chardon and Riser (1981), seemingly were the first to interpret a lobate feature in the Irhzer Ikhibi south at 3800 m as an active rock glacier. Vieira et al. (2017) installed a series of ground surface temperature and air temperature dataloggers during one full year, from 3210 to 4160 m altitude. The authors analysed the ground temperature regimes and snow cover and identified a hot season from late-May to late-September and a long cold season from mid-October to mid-April. Freeze-thaw regimes were analysed and the most important finding was the possible presence of permafrost at a location close to the rock glacier identified by Chardon and Riser (1981). This interpretation was based on the very low temperatures (c. -5.8 ºC) measured at the ground surface beneath a stable snow pack that lasted from mid-December until late-March. Other lobate rock debris features are found at similar altitudes in other cirques nearby and it is possible that these are also associated with sporadic patches of permafrost (Hughes, 2018). Permanent snowfields were present in the High Atlas in the mid-20th century and were probably widespread in the LIA (Hughes, 2018).

Figure 14

4.7 Mediterranean islands

The Mediterranean Sea encompasses several archipelagos and some major islands, some of which exceed 10,000 km^2 (Sicily, Sardinia, Corsica, Crete). Some of the islands are very mountainous, with elevations exceeding 2000 m in Sicily (Etna, 3350 m), Corsica (Monte Cinto, 2706 m), Crete (Lefka Ori, 2452 m). Lithologies are highly variable, with abundant limestones and granites and also volcanic rocks in the case of the highest elevation within the limits of the Mediterranean Sea, the Etna. While past glacial activity in Mediterranean islands has received some attention (e.g. Hughes et al., 2006a, Hughes and Woodward, 2016), periglacial processes have been less examined.

Last Glaciation

Glaciation has affected the mountains of Corsica, and the active volcano of Mt. Etna in Sicily, and the mountains of Crete. In the highest mountains in Corsica, Kuhlemann et al. (2008) dated *roches moutonnées* at 18 ka BP. In Crete, Bathrellos et al. (2014) found 17 glacial cirques between 1870 and 2360 m distributed around the two main mountains groups: Mt. Lefka Ori (2453 m) and Mt. Idi (2456 m). However, the ages of the Cretan glaciations remain untested and could potentially be older than the Late Pleistocene as in parts of mainland Greece (cf. Hughes et al., 2007).

Past periglacial activity has been documented in several Mediterranean islands associated in most of the cases to seasonal frost regime. In Corsica, despite the quite extensive glaciation during the LGM periglacial landforms are not common if we exclude granite weathering features like tafoni, weathering pits or grooves that are extensively widespread in the island and that occur largely also in modern cryotic conditions in Antarctica (i.e Guglielmin et al., 2005; Strini et al., 2008). Kuhlemann et al. (2009) recorded blockfields and tors above 2200 m and describe them as possible periglacial landforms, even reporting less common similar features at lower altitudes.

Landforms associated to intense periglacial conditions are also observed in Sardinia as indicated by Ginesu (1990) who described several blockfields and block streams composed by basalts in the centre of the island (Pranu Mannu plateau) at an altitude of about 600 m. Ginesu and Sias (2006) suggested a periglacial origin also for other block accumulations on different lithologies in the higher reliefs of Sardinia (Gennargentu, Limbara, Perdasdefogu). More recently, Ginesu et al. (2014) interpreted some block deposits found on Asinara island as block streams. In this case, block streams are composed by granites subrounded blocks and located close to the coast and even below the present sea level (4 m depth) in Cala Arena. Block streams and block slopes close to Tyrrenian coast have been described in the Pisani Mountains between 400 and 800 m on quarzites by Casarosa and Pappalardo (2006).

Poser (1957) found patterned ground and solifluction lobes above 1800 m in Crete, suggesting that the lower limit of periglacial activity in this island was represented by boulder pavements extending down to elevations of 800 m. Cemented stratified scree of unknown age are present on the southern slope of Mt. Idi (Guglielmin, unpublished), probably related to the coldest phases of the Last Glaciation.

In the Balearic Islands there is evidence of periglacial activity during the Last Glaciation, with scree deposits located at elevations above 1200 m in the limestone Tramuntana massif (Rosselló, 1977) though no evidence of permafrost conditions was detected at this altitude.

Deglaciation

The temperature increase recorded following the LGM conditioned the deglaciation of Mediterranean mountains, with only the possible existence of minor glaciers during the OD and YD in the highest mountains in Corsica. No evidence of permafrost conditions has been reported for this stage, though periglacial processes driven by seasonal frost may have existed during the coldest stages in the highest mountains.

Holocene

It is very unlikely that climate conditions prevailing during the Holocene allowed the existence of permafrost even in the high lands of the major islands.

LIA and present-day

LIA cold-climate conditions probably reactivated periglacial processes in the highest mountains in Sardinia and Sicily. In fact, evidence of permanent ice deposits in lava tubes has been found at 2043 m on the north flank of Mt Etna volcano (*Grotta del Gelo,* Cave of Frost), in Sicily. It is probably the southernmost European ice cave (Marino, 1992; Hughes and Woodward, 2009; Scoto et al., 2016). The lava tube formed during the historic long-lasting eruption of 1614-24 and despite its geological setting and latitude, after about twenty years from the last phases of the eruption (coinciding with the Minimum Maunder, 1645-1715), subterranean freezing inside the cave started to take place. Beside the development of seasonal ice formations (seasonal lake ice, ice stalactites, stalagmites and columns generally located close to the entrance), perennial ground ice is present in the deepest zone. The ice extension reaches about 240 m^2, and the volume is estimated at about 220-260 m^3 (Scoto et al., 2016). On Mt. Etna, at elevations above 2900 m, nivo-aeolian deposits were found with frozen ground beneath that can persist for more than 2 years, and therefore can be considered as current permafrost (Guglielmin, unpublished).

5. Discussion

Climate variability during the Quaternary has conditioned the spatial distribution of glacial and periglacial processes in the Mediterranean region, and therefore the area under permafrost conditions (Figure 15). Glacial stages favoured a substantial expansion of glaciers in the currently still glaciated mountain ranges and the formation of glaciers in presently deglaciated mountain environments (Woodward, 2009). The glaciation in the Mediterranean mountains has been long debated and several reviews have focused on the calendar and geography of the maximum glacial expansion in the different mountain regions (i.e. Allen et al., 1999; Hughes et al., 2006a, 2008, 2013; Hughes and Woodward, 2016). The temperature increase recorded during interglacial phases - such as the Holocene - conditioned the complete disappearance or substantial retreat of glaciers and the migration of permafrost and periglacial processes to higher elevations (i.e. Oliva et al., 2016b).

Figure 15

5.1 Last Glaciation

The calendar of the maximum glacial expansion of the Last Glaciation shows a diachronous pattern among the massifs, which must be also framed with the dating method used in each study (Hughes et al., 2013). Whereas radiocarbon, U-series and OSL indicate an early glacial advance that occurred several thousand to tens of thousands of years earlier than the global LGM in the Sierra Nevada, Cantabrian Mountains, Central Pyrenees, Italian Apennines and Pindus Mountains, cosmogenic exposure ages suggest a local MIE in the Iberian Central Range, Maritime Alps and Anatolia mountains (almost) synchronous to the LGM (Hughes et al., 2008).

The FIS extended over northern Europe until latitude 50ºN in its southern fringe (Hughes et al., 2015; Stroeven et al., 2016). Therefore, the Mediterranean region was not extensively glaciated by comparison, with glaciers concentrated in mountain environments in the form of small ice caps and mountain glaciers filling the valleys. The ELA decrease in Mediterranean mountains during the LGM compared to present-day was about 800 and 1200 m, which is translated in a minimum temperature difference of ca. 6-10 ºC (Allen et al., 2008; Kuhlemann et al., 2008). The glaciated environments in the western and central Mediterranean region was associated with ELAs above 1500-2000 m in most of mountain ranges (Hughes et al., 2006a, 2008), slightly increasing towards the easternmost fringe in mountains of Turkish and Lebanon where the ELA was located around 2500-3000 m (Messerli, 1967).

The southern ice-free land surrounding the large FIS was underlain by permafrost, which extended across the lowlands in much of central Europe as well as the northern fringe of the Mediterranean region, in areas such as central France and the northern Balkan region (Brown, 2001; Vandenberghe et al., 2014). Dry and cold climate conditions prevailing in these permafrost areas during the LGM promoted intense aeolian activity, with the deposition of thick loess deposits in non-glaciated environments from central-northern Europe (Antoine et al., 2009) and northern Italy (Cremaschi et al., 2015). In the Mediterranean region, the presence of ice wedges and cryoturbations features in some basins, such as in the Pannonian basin (van Vliet- Lanoe et al., 2004; Ruszkiczay-Rüdiger and Kern, 2015) or in some basins in the Spanish Meseta (Badorrey et al., 1970; Asensio-Amor and González-Martín, 1974; Serrano et al., 2010a) has been related to deep seasonal frost conditions or even isolated patches of sporadic permafrost. Despite few data in areas surrounding the valley glaciers in the southern Alps, environments at the foot of the northern Apennines and even at the Po plain should have been affected by permafrost conditions as suggested by loess deposits and the existence of some cryoturbation features (Cremaschi et al., 2015). In the Mediterranean mountains, immediately below the glaciated environments was the periglacial belt affected by permafrost regime at high elevations and seasonal frost at the foot of the mountains and high-altitude plateaus (e.g. 600-1200 m).

Permafrost was also widespread in non-glaciated environments above the snowline as revealed by the present-day distribution of inactive permafrost-derived features. Rock glaciers formed during the Last Glaciation have been described in several Mediterranean mountains at a wide range of altitudes (e.g. Figure 3 and Table 5). Block streams, some km long and hundreds of meters wide, also developed during this stage at elevations between 700 and 1800 m. In relatively flat summit areas where wind action did not favour snow and ice accumulation, meter-sized stone circles developed in several massifs except for the Italian Peninsula where there are no evidences of large stone circles in the LGM ice-free areas. Similar patterned ground features are observed today in present-day polar environments where permafrost conditions are widespread with mean annual temperatures below -6 ºC (French, 2007). It is therefore feasible to relate the formation of these currently inactive features to permafrost occurrence. Concurrently, very intense periglacial conditions favoured the development of large blockfields and tors, which are remnants of original surfaces in the highest lands. The thermal regime of the ground in the nunataks standing out the glaciated slopes must have been also characterized by permanent frozen conditions well below 0 ºC.

5.2 Deglaciation

During the most important phases of the last deglaciation, the evolution of the glaciers of the different Mediterranean mountains shows a much more homogenous pattern than during the LGM. The latest studies indicate interruptions during the deglaciation process and even some important glacial advances, in full synchrony within the Mediterranean region and with the mountainous systems of central Europe, as well as with the FIS. The onset of a massive retreat of the FIS is dated at 19 ka and accelerated at 18 ka (Marks, 2015; Stroeven et al., 2015; Toucanne et al., 2015; A.L. Hughes et al., 2016; Petrini et al., 2018), in parallel to most European mountains. For example, some of the largest glaciers in the Alps had lost ca. 80% of their mass at 18 ka (Ivy-Ochs et al., 2004, Ivy-Ochs, 2015) and glaciers could almost have disappeared completely in the Pyrenees at that time, as probably occurred in many other Mediterranean mountains (Palacios et al., 2017a).

The first glacial advance within the deglaciation took place during the OD, when advances have been detected in numerous sectors of the FIS (Marks, 2015; Stroeven et al., 2015; Toucanne et al., 2015; A.L. Hughes et al., 2016). In the Alps, glaciers expanded along the valley bottoms with their fronts reaching elevations only 400 m above the LGM moraines (Kerschner and Ivy-Ochs, 2008; Ivy-Ochs et al., 2009; Ivy-Ochs, 2015). Similar glacial advances occurred during this period in the Tatra Mountains (Makos, 2015) as well as Mediterranean glaciers that also experienced important advances, as in the Iberian Peninsula (Palacios et al., 2017a), Southern Alps (Federici et al., 2012), Apennines (Giraudi, 2015), Balkan Peninsula (Kuhlemann et al., 2013) and Anatolian Peninsula (Sarıkaya et al., 2008, Akçar et al., 2014). A significant drop of the North Atlantic temperature induced a significant reduction of the meridional overturning circulation (Bard et al., 2000; McManus et al., 2004), with extreme seasonality of cold winters and mild summers (Denton et al., 2005; Williams et al., 2012) favouring glacial expansion during this period. Similar conditions have also been detected in southern Europe and in the Mediterranean region (Fletcher et al., 2010). For example, glaciological models indicate that during the OD the MAAT in the Alps was 10 ºC lower than present-day, and precipitation was one-third less (Ivy-Ochs, 2015). Undoubtedly, these conditions would have favoured a substantial expansion of permafrost conditions in mountain environments. Many fossil periglacial landforms existing in Mediterranean regions, such as patterned ground, protalus lobes or rock glaciers, could have originated at this time. But these climatic conditions changed drastically during the BO, when the environmental conditions shifted abruptly and become very similar to present in the Mediterranean region (Fletcher et al., 2010). Consequently, there was a massive glacial retreat in Europe, including the FIS and British-Irish Ice Sheet (Marks, 2015; Stroeven et al., 2015; Toucanne et al., 2015; Hughes et al., 2016), central European mountains (Ivy-Ochs, 2015; Makos, 2015) and Mediterranean mountains (Sarıkaya et al., 2008; Federici et al., 2012; Kuhlemann et al., 2013; Akçar et al., 2014; Giraudi, 2015; Palacios et al., 2017a).

The recent application of absolute dating methods, fundamentally cosmogenic, to fossil rock glaciers is allowing a better comprehension of the timing of their formation and stabilization, remaining therefore exposed to cosmogenic radiation. In many European mountains, this method is showing that numerous rock glaciers developed on polished glacial surfaces that were deglaciated just at the end of OD. In many cases, the fronts of these rock glaciers were stabilized soon after formation, although their roots remained active during thousands of years (Oliva et al., 2016b, Palacios et al., 2017a,b). This fact shows evidence that most fossil rock glaciers in different Mediterranean mountain ranges developed under paraglacial conditions (Ballantyne, 2002; Mercier, 2008; Oliva et al., 2016b). They occupied the formerly glaciated cirques, extending over very active geomorphological areas at the end of the OD. Subsequently, as temperatures increased during the OD permafrost conditions migrate to higher elevations and rock glaciers became relict. This process has been mostly described in many Iberian mountain environments (Rodríguez-Rodríguez, 2016; Fernández-Fernández, 2017; Palacios et al., 2017a,b), and should be tested in the future in other Mediterranean ranges. Noteworthy, a similar pattern has been also reported in other mountains, such as in the British Isles (Ballantyne et al., 2009) and the Alps (Hippolyte et al., 2009).

During the YD, in full agreement with the last major advance of the FIS (Greenwood et al., 2015; A.L. Hughes et al., 2016; Stroeven et al., 2015), Mediterranean glaciers advanced again for the last time during the Pleistocene, but only in the highest massifs. Glaciers rarely exceeded the limits of the cirques in the Pyrenees (García-Ruiz et al., 2016a,b) and were moderately longer in the southern Alps (Ivy-Ochs et al., 2008, 2009). This glacial advance was related to a cold and arid period during the YD in the Mediterranean region that ended abruptly at 11.8 ka cal BP (Fletcher et al., 2010). In the Alps, where glaciers had virtually disappeared during the BO, they formed again during the YD and progressed considerably (Ivy-Ochs, 2015) due to 3.5-5 ºC lower MAAT than present and precipitation was up to 30% less (Kerschner and Ivy-Ochs, 2008). A similar pattern was also detected in the Tatra Mountains (Makos, 2015).

As occurred at the end of the OD, the end of the YD favoured the rapid retreat of glaciers and, in some cases, the reactivation of paraglacial processes in the walls of the cirques and the formation of rock glaciers and protalus lobes. Again, their fronts became inactive soon after formation, while the roots could have remained active even until the warmest Holocene periods (García-Ruiz et al., 2016a,b; Oliva et al 2016b). Many generations of Mediterranean rock glaciers have been considered to form during or at the end of the YD, as reported in many Iberian mountains (Palacios et al., 2015, 2016; Andrés et al., 2018) and southern Alps (Colucci et al., 2016a), though in other high mountain Mediterranean regions this fact still needs to be confirmed.

Therefore, the cold periods interrupting the long-term deglaciation process showed both glacial and periglacial evidence in the Mediterranean mountains, mainly with the formation of permafrost-related features associated with paraglacial dynamics. The existence of these inactive landforms suggests the minimum altitude for the presence of permafrost during these stages.

5.3 Holocene

The northern Atlantic region has been subjected to significant climate shifts during the Holocene, as revealed by marine sediment records (Witak et al., 2015) and Greenland ice cores (Masson-Delmotte et al., 2005). Climate variability has been also significant across the Mediterranean region, where Holocene temperature oscillations of the order of ca. ± 2 ºC have caused significant disruptions in early civilizations (Mayewski et al., 2004). During the Holocene, temperature and moisture shifts have also affected the type and intensity of cold-climate geomorphological processes prevailing in Mediterranean mountains.

The onset of the Holocene saw an accelerated shrinking of the glaciers, which disappeared until nowadays in many massifs during the Early Holocene (Gómez-Ortiz et al., 2012a; García-Ruiz et al., 2016a,b; Palacios et al., 2016). The formerly glaciated environments became occupied by periglacial dynamics, which expanded gradually upvalleys (Oliva et al., 2016b). In many mid-altitude mountain ranges – with highest peaks around 2000 m in the central and western Mediterranean and 2500-3000 m in the eastern part of the region – the periglacial belt disappeared during the Holocene, and significantly shrunk in the highest ranges, particularly during the warmest stages, such as the Holocene Warm Period (HWP) and the Medieval Climate Anomaly (MCA). In the southern Alps, even during warmer Holocene stages some currently active rock glaciers were still active (Calderoni et al., 1998; Stenni et al., 2007; Scapozza et al., 2010); indeed, the major part of the dated rock glaciers reported an age between 2720 and 2850 cal yr BP with some older exceptions (all younger than 5900 cal yr BP; Calderoni et al., 1998; Guglielmin et al., 2001; Dramis et al, 2003). During the warmest phases, only the environments above 2500-3000 m included a periglacial belt mostly related to seasonal frost conditions, with permafrost regime limited to the highest ranges in the Maritime Alps and the highest mountains in Anatolia.

Most of the active periglacial features during the YD became gradually inactive, and only reactivated during the coldest stages of the Holocene. Permafrost-related features showed activity until the Early Holocene in many of the highest Mediterranean massifs (Kuhlemann et al., 2009; Gómez-Ortiz et al., 2012a; Palacios et al., 2016, 2017a,b). Some authors proposed the initial formation of currently active rock glaciers synchronously to the HWP, such as in the central Taurus Mountains (Çiner et al., 2017) and the Pyrenees at ca. 6 ka (Serrano et al., 2010c). Periglacial landforms related to seasonal frost conditions, such as solifluction landforms located today in the present-day periglacial belt of Sierra Nevada at elevations between 2500 and 3000 m, were inactive during the HWP until 5 ka cal BP but reactivated later during cold and wet phases, namely at 5-4, 3.6-3.4, 3-2.8, 2.5-2.3, 1.8-1.6, 0.85-0.7, 0.4-0.15 ka cal BP (Oliva et al., 2011). Therefore, 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 permafrost conditions.

5.4 Little Ice Age

The LIA has been defined as the coldest most recent period where many glaciers reached their largest volume over the last 10,000 years (Bradley and Jones, 1992), and, thus, it has been widely employed in mid-latitude mountain environments as the reference for Holocene cold stages (Grove, 2004; Oliva et al., 2018). In the Mediterranean region, colder and (generally) wetter than present-day climate conditions prevailing during the LIA also favoured the presence of larger and more numerous glaciers (Hughes, 2014). Consequently, the spatial domain of periglacial dynamics expanded down-valleys and cryogenic processes reappeared in some areas where seasonal frost activity was limited during the MCA and is no longer active at present.

During the LIA permafrost features such as rock glaciers and protalus lobes formed and/or reactivated in the highest slopes of the main massifs, above 2350 m in some areas of the Balkan region and 2560 m in the Pyrenees (Tables 6 and 2). In the southern Alps, especially in the eastern part several rock glaciers developed since the LIA around 2400-2500 m (Baroni et al., 2004; Seppi et al., 2014) while only a few active rock glaciers formed in the western Alps and central Alps.

Ice caves, namely natural caves formed in bedrock containing perennial accumulations of ice, are considered as sporadic permafrost phenomena (Holmlund et al., 2005; Luetscher et al., 2005; Hausmann and Behm, 2011; Luetscher et al., 2013). As part of the cryosphere, ice caves occurrence is closely linked to cold climates, even if they do also exist in different kind of environments, often at an altitude with an outside mean annual air temperature well-above 0°C (Holmlund et al., 2005; Stoffel et al., 2009; Obleitner and Spötl, 2012; Colucci et al., 2016b). Ice caves existing in several high karstic mountains have been examined for paleoenvironmental purposes. In most cases these ice caves are several hundreds of meters below the glaciated environments during the LIA, reaching elevations even below 1000 m in the Dinaric Alps (Kern et al., 2006b; Bočić et al., 2014). The dating of the organic remnants preserved in the ice suggests that it accumulated during the coldest centuries of the LIA (Bayarı et al., 2003; Kern et al., 2006b; Klimchouk et al., 2006; Bočić et al., 2014; Gómez-Lende, 2015; Bartolomé et al., 2015; Temovski, 2016; Zupan-Hajna, 2016; Sancho et al., 2016), and even during the MCA and HWP (Sancho et al., 2016). Interestingly, the frozen ground in contact with the ice inside the caves has been also described as permafrost (Gómez-Lende, 2015; Colucci et al., 2016b), which in many cases still persist in environments with MAAT well above 0 °C due to specific microtopographic conditions (very low solar radiation, existence of stable temperature inversion or density driven air flows and chimney effects).

In all Mediterranean massifs, long-lasting and perennial snow-patches were more extensive than today, which also enhanced nival processes that left some small well-preserved landforms such as protalus ramparts at the foot of slopes.

5.5 Present-day

In the western Mediterranean region, the temperature increase since the last cold stages of the LIA has been estimated at ca. 1 ºC (González-Trueba, 2006; González-Trueba et al., 2008; Oliva and Gómez-Ortiz, 2012) and ca. 2 ºC since the Minimum Maunder (Oliva et al., 2018). A similar increase has been inferred for other parts of the Mediterranean, such as in the Balkans (Repapis and Philandras, 1988 Xoplaki et al., 2001; Hughes, 2010). This temperature increase has led to glacial retreat/disappearance (Zumbühl et al., 2008), enhanced paraglacial activity (Cossart and Fort, 2008), shift of periglacial processes to higher elevations (Oliva et al., 2011), geoecological changes (Cannone et al., 2007, 2008; Pauli et al., 2012; García-Ruiz et al., 2015; Camarero et al., 2016), as well as degradation of alpine permafrost in mid-latitude high mountain environments, e.g. the Alps (Harris et al., 2003; Gruber et al., 2004; Lugon et al., 2004; Zenklusen-Mutter et al., 2010).

A wide range of approaches have been used to infer the current altitudinal limit of permafrost conditions, such as geomorphological techniques, geophysical surveying and monitoring activities (BTS measurements, boreholes, terrain deformation). Today, permafrost conditions generally increase in elevation towards the eastern part of the region and from north to south. In the western and central Mediterranean, permanently frozen ground is rarely found below 2500 m, discontinuous permafrost is generally detected between 2500 and 2800/3000 m and continuous permafrost is distributed in ice-free environments above this level. An alpine permafrost belt is detected above 2630 m in northern aspects and 2800 m in southern ones in the Pyrenees (Serrano et al., 1999, 2001, 2002, 2006, 2009, 2011a; González-García et al., 2014), above 2400 m in the Southern Alps (Bodin et al., 2009), above ~2350 m on Rila Mountain and ~2700 m on Mount Olympus (Dobinski, 2005) and above 2800-3400 m in NE Turkey and central Anatolia (Gorbunov, 2012). No permafrost belt is found in the highest mountains in southern Europe (i.e. Sierra Nevada) and northern Africa (i.e. Atlas) where permanent frozen conditions are only found in form of isolated patches at the highest elevations at 3000-3100 m (Oliva et al., 2016a) and 3800 m (Vieira et al., 2017), respectively. Certain climate conditions (i.e. reduced snow cover) can favour the presence of permafrost patches at relatively low elevations in the Central Apennines on La Majella and M. Velino massifs (ca. 2400 m) or by lithological conditions (i.e. volcanic sediments), as detected in the highest active European volcano (Mt. Etna) at elevations above 2900 m.

Scientific papers and international reports coincide in anticipating that the Mediterranean basin will be one of the areas on Earth where annual and seasonal climate variations will be more pronounced, with significant warming and reduced rainfall (e.g. IPCC, 2013). These climate conditions would lead to a migration of permafrost conditions to upper elevations, even disappearing in those areas where isolated patches persist but undergo a rapid degradation process (e.g. Atlas, Sierra Nevada Cantabrian Mountains), possibly triggering large rock avalanches or other mass wasting processes with important socio-economic impacts.

6. Conclusions

The present and past distribution of cold-climate geomorphological processes in the Mediterranean region is conditioned by the topography and wide spectrum of microclimatic conditions prevailing in the region. Both glacial and periglacial processes (including permafrost conditions) since the Last Glaciation have been almost exclusively restricted to mountain environments.

Since the Last Glaciation there has been a long-term decrease in the area occupied by permafrost in the Mediterranean mountains. The existence of abundant inactive permafrost-derived landforms (rock glaciers, block streams, patterned ground features, ice wedges) formed during the last Pleistocene glacial cycle suggests that ice-free slopes in mid-altitude mountain environments and high summit plateaus together with the lowlands of the northernmost fringe of the Mediterranean region could have been underlain by permafrost conditions to a greater or lesser extent during that stage. The thermal increase following the LGM meant that, after that time, periglacial conditions were limited to mountain areas. The gradual upward shift of the snowline in all Mediterranean massifs conditioned a massive deglaciation that was only interrupted by brief periods of glacial readvance. Paraglacial dynamics during the deglaciation process favoured the development of most of the rock glaciers distributed in the majority of the massifs, almost all of which are inactive under present-day climate conditions. Therefore, their formation is primarily associated with adjustment of the cirque walls to a new morphodynamic setting and not strictly related to a climatic origin. Warmer temperatures during the onset of the Holocene saw a shrinking of periglacial activity to the highest elevations. Many rock glaciers and protalus lobes became inactive during the Early Holocene as temperatures rose. Permafrost disappeared from most of the massifs and only reappeared in some mountains during the coldest stages of the Holocene, such as the LIA. Since then, post-LIA warming led to the spatial confinement of continuous/discontinuous permafrost conditions in the highest mountain areas, such as some areas in the Pyrenees, Southern Alps, Apennines and Anatolian mountains, or as isolated patches in north-facing cirques that were glaciated during the LIA in the Sierra Nevada, Atlas Mountains and the Balkan region.

In contrast to neighbouring mountain environments where there has been a richer analysis of permafrost history (i.e. Alps), permafrost research in the Mediterranean basin still has some gaps:

- The improvement of the glacial chronology that has taken place over the last decade in many Mediterranean massifs has not been paralleled by a strengthening of the chronology of periglacial activity, and therefore of permafrost evolution in the Mediterranean region.
- Data about periglacial dynamics and spatial domains of permafrost are substantial for certain periods (i.e. the last deglaciation), though current knowledge is still poor for others (i.e. Holocene).

- Knowledge of permafrost evolution in some mountain environments is still limited, particularly in North Africa ranges, Apennines and Mediterranean islands.
- Future studies should focus on precise descriptions of landforms and quantitative age proxy data, especially on rock glaciers as indicators of present and past permafrost conditions.

Finally, permafrost scientists in the Mediterranean region should focus on these gaps in order to better understand the spatio-temporal evolution of permafrost conditions in the region. A better characterization of the evolution of permafrost - a key component of the cryosphere in landscape dynamics in mid-latitude high mountain ranges – may be also helpful to anticipate the future geoecological response towards the changing climate scenarios forecasted in these highly sensitive mountain ecosystems.

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Figure captions

Figure 1. Location of the different study areas examined in this paper within the Mediterranean region (top), together with mean annual temperatures and altitudes where the 0 ºC isotherm is located (middle), and annual precipitations with values for each of the mountain ranges considered in this study (bottom). Annual precipitation and mean annual temperatures for the period 1960-1990 was obtained from Hijmans et al. (2005), spatial resolution 30 arc-seconds, modified after WorldClim 1.4 (www.worldclim.org).

Figure 2. Distribution of permafrost-related features since the Last Glaciation in the Iberian Peninsula. The symbols of the legend are the same for all the other graphs.

Figure 3. Examples of permafrost-related landforms generated during different phases in different massifs of the Iberian Peninsula.

Figure 4. Distribution of permafrost-related features since the Last Glaciation in the southern Alps.

Figure 5. Minimum mean altitude of the relict rock glacier fronts in the southern side of the Alps, including data from (1) Bornet et al. (2014); 2) Seppi et al. (2012); (3) Colucci et al. (2016a).

Figure 6. Examples of periglacial and permafrost-related landforms generated during different phases in the southern Alps.

Figure 7. Distribution of permafrost-related features since the Last Glaciation in the Italian Peninsula.

Figure 8. Examples of permafrost-related landforms generated during different phases in the Apennines and Italian Peninsula.

Figure 9. Distribution of permafrost-related features since the Last Glaciation in the Balkan Peninsula.

Figure 10. Examples of permafrost-related landforms generated during different phases in different massifs of the Balkan Peninsula.

Figure 11. Distribution of permafrost-related features since the Last Glaciation in the Anatolia Peninsula.

Figure 12. Examples of permafrost-related landforms generated during different phases in different massifs of the Anatolia Peninsula.

Figure 13. Distribution of permafrost-related features since the Last Glaciation in northern Africa.

Figure 14. Examples of permafrost-related landforms generated during different phases in northern Africa.

Figure 15. Geomorphological sketch of the formation of different generations of permafrost-related features in Mediterranean mountains since the Last Glaciation.

Abstract

The relatively warm climate conditions prevailing today in the Mediterranean region limit cold geomorphological processes only to the highest mountain environments. However, climate variability during the Late Pleistocene and Holocene has led to significant spatio-temporal variations of the glacial and periglacial domain in these mountains, including permafrost conditions. Here, we examine the distribution and evolution of permafrost in the Mediterranean region considering five time periods: Last Glaciation, deglaciation, Holocene, Little Ice Age (LIA) and present-day. The distribution of inactive permafrost-derived features as well as sedimentary records indicates that the elevation limit of permafrost during the Last Glaciation was between 1000 m and even 2000 m lower than present. Permafrost was also widespread in non-glaciated slopes above the snowline forming rock glaciers and block streams, as well as meter-sized stone circles in relatively flat summit areas. As in most of the Northern Hemisphere, the onset of deglaciation in the Mediterranean region started around 19-20 ka. The ice-free terrain left by retreating glaciers was subject to paraglacial activity and intense periglacial processes under permafrost conditions. Many rock glaciers, protalus lobes and block streams formed in these recently deglaciated environments, though most of them became gradually inactive as temperatures kept rising, especially those at lower altitudes. Following the Younger Dryas glacial advance, the Early Holocene saw the last massive deglaciation in Mediterranean mountains accompanied by a progressive shift of permafrost conditions to higher elevations. It is unlikely that air temperatures recorded in Mediterranean mountains during the Holocene favoured the existence of widespread permafrost regimes, with the only exception of the highest massifs exceeding 2500-3000 m. LIA colder climate promoted a minor glacial advance and the spatial expansion of permafrost, with the development of new protalus lobes and rock glaciers in the highest massifs. Finally, post-LIA warming has led to glacial retreat/disappearance, enhanced paraglacial activity, shift of periglacial processes to higher elevations, degradation of alpine permafrost along with geoecological changes.

Key words: Mediterranean region, permafrost, Last Glacial Maximum, deglaciation, Holocene.

Permafrost conditions in the Mediterranean region since the Last Glaciation

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1. Introduction

Permafrost is one of the elements of the cryosphere with the widest global distribution. The area occupied by permafrost in both hemispheres varies according to different authors between 22.10^6 km^2 (17% of the Earth's exposed land surface; Gruber, 2012) and 36.10^6 km^2 (24%; French, 2007). Terrestrial ecosystems in many ice-free areas at high latitudes and in high mountains are conditioned by the presence of permafrost. Additionally, permafrost has an influence on the socioeconomic activities of these areas, especially those related to the construction and maintenance of all kinds of equipment and infrastructures (Nelson et al., 2001, 2002; Bommer et al., 2010). However, as a research topic, permafrost has received much less global attention than other elements of the cryosphere, such as glaciers, which are highly visible elements in the landscape and carry important hydrological and global sea level consequences (e.g. Favier et al., 2014; Smith et al., 2017), and sea ice due to its influence on ocean-atmospheric interactions (e.g. Curran et al., 2003; Robinson et al., 2012) and geopolitical implications across the Arctic (Østerud and Hønneland, 2014).

As a result of the International Polar Year 2007-2008, there has been an important development in the study of the thermal state and current distribution of permafrost (Lewkowicz, 2010) and the active layer (Oliva et al., 2017), along with their geoecological implications (Jorgenson et al., 2001; Jorgenson and Osterkamp, 2005; Genxu et al., 2006). Most studies have focused mainly on the Arctic (e.g. Mackay, 2000; Brown et al., 2001; Lantuit et al., 2012; Fritz et al., 2017), but also on Antarctica (e.g. Vieira et al., 2010; Bockheim et al., 2013; Guglielmin et al., 2014; Oliva and Ruiz-Fernández, 2015) and mountainous areas with increased population and tourism pressure (e.g. Alps, Haeberli and Beniston, 1998; Boeckli et al., 2012). Permafrost degradation as the consequence of recent global warming is already posing risks and problems to certain socio-economic activities in many areas (Nelson et al., 2001, 2002; Ravanel, et al., 2017). Besides, during the last decade, scientists have pointed out the importance of permafrost not only at the local and regional level, but also at global scale examining the role of permafrost thawing in the carbon cycle and greenhouse gas fixation, as well as its contribution to recent and future climatic evolution (e.g. Hugelius et al., 2014; Zubrzycki et al., 2014; Schuur et al., 2015).

From a geographical perspective, the thermal state and distribution of permafrost have been much less studied in environments such as mountain ranges in the mid-latitudes. This is the case of the high mountains of the Mediterranean basin, where knowledge about the thermal dynamics, depth and extent of permafrost is uneven across the region.

The Mediterranean region is a transition zone between the subtropical climatic area and temperate regions more characteristic of the middle latitudes (Woodward, 2009). This region has its own climatic characteristics, conditioned by the presence of a nearly closed sea (the Mediterranean), and a complex and mountainous landscape defined by the existence of large peripheral mountain ranges, rugged coasts, and many islands and archipelagos. Most research on cold processes in the Mediterranean mountains has focused on the Quaternary glaciations (Hughes et al., 2006a, 2007, 2013; Hughes and Woodward, 2008), while permafrost and periglacial processes have been less examined (Hughes and Woodward, 2009). Thus, knowledge of the distribution of permafrost in this basin in the past is limited to generic maps that represent the area occupied by permafrost in the Northern Hemisphere during the Last Glacial Maximum (LGM; 26.5-20/19 ka (Clark et al., 2009); 27.5-23.3 ka (Hughes and Gibbard, 2015), as a complement to glacial extent (Brown et al., 2001, Vandenberghe et al., 2014; Kitover et al., 2016). However, the Mediterranean region is loosely represented in these maps since permafrost areas are barely included, except for certain areas such as the Alps and the SE coast of France. Besides, knowledge of the distribution of permafrost following the LGM and throughout the Holocene is limited and geographically sparse.

As far as the current dynamics are concerned, the existing studies on permafrost and related periglacial morphodynamics mainly concentrate on the southern slope of the Alps and the Pyrenees (Guglielmin et al., 1994; Hoelzle, 1996; Dramis et al., 1995; Ikeda and Matsuoka, 2002; Lugon et al., 2004; Julián and Chueca, 2007; Serrano et al., 2009, 2010; Colucci et al., 2016a), and to a much lesser extent on other massifs, such as the Sierra Nevada (Gómez-Ortiz et al., 2001, 2014; Oliva et al., 2016b) and the Atlas (Vieira et al, 2017).

In consequence, permafrost studies in the Mediterranean region are spatially limited and temporally dispersed. Thus, it is essential to encourage detailed reflection on the state of knowledge of permafrost in this region, and identify the areas and periods in which research on this topic needs to be intensified. Accordingly, the general aim of this article is to fill this gap by looking into the spatial distribution of permafrost in the mountains of the Mediterranean region since the Last Glaciation, which is defined by the last cold stage (Weichselian/Würmian; marine isotope stage (MIS) 5d-2), to the present day. To meet this aim, the following specific objectives will be addressed:

- Reconstructing the spatio-temporal evolution of permafrost in this region since the Last Glaciation.
- Identifying the Mediterranean areas where permafrost has been studied.
- Determining the best-known permafrost stages and those with knowledge gaps.
- Pointing out future lines of work to the scientific community centred on the study of permafrost in the Mediterranean mountains.
- Discussing the distribution of the spatio-temporal pattern of the areas affected by permafrost and its associated geomorphological processes within the climatic and environmental evolution recognised on the European continent since the Last Glaciation.

2. Regional setting

Extending over a surface of ca. 2.5.10^6 km^2 (Jeftic et al., 1989), the Mediterranean constitutes an almost enclosed basin between southern Europe (north), Anatolia (east), northern Africa (south) and the Iberian Peninsula (west), only connected to the Atlantic Ocean through the Strait of Gibraltar (ca. 15 km wide). Etymologically, Mediterranean derives from the Latin '*medius*' (middle) and '*terra*' (land), highlighting one of the main characteristics of the Mediterranean region: the presence of mountains next to the sea (Figure 1). Several volcanoes and mountain ranges with elevations exceeding 2500-3000 m asl stretch a few km off the coastline (Table 1).

Figure 1

Table 1

The Mediterranean basin is located in a boundary position between the mid-latitude westerlies and the influence of continental and subtropical high-pressure systems. The existence of multiple peninsulas, islands, rugged coastlines and mountain ranges determines a broad spectrum of climate regimes across the region, both in terms of temperatures and precipitation (Figure 1). This results in a wide range of hydrological, edaphic, geomorphological and biological processes prevailing in the Mediterranean region.

A broad variety of marine and terrestrial records preserve evidence of the nature and magnitude of Quaternary climate oscillations driving environmental changes in the Mediterranean basin (Woodward, 2009). Climate changes also determined large social transformations transitioning from the prehistoric inhabitants to the first sedentary agricultural communities that settled in the fertile soils of the Mediterranean lowlands (Mithen, 2004). The topographic configuration of the Mediterranean and its relief, together with relatively warm climate conditions during the Holocene, favoured the development of early civilizations, agriculture flourishing and trade and cultural exchange. Human-induced activities, together with shifts on ecosystem dynamics promoted by climate fluctuations, substantially affected the Mediterranean landscape, particularly since the Mid-Holocene, accelerating over the last few centuries.

Quaternary climate variability has followed different spatio-temporal patterns across the Mediterranean basin, which resulted in changing cold-climate geomorphological processes occurring in the mountains of the region. Glacial and periglacial activity has shaped the highest environments during the Late Pleistocene (Hughes and Woodward, 2016). The timing and spatial extent of glacial processes was highly conditioned by the combination of cold and moisture regimes, with significant spatio-temporal differences between the areas (Hughes et al, 2006a,b; Hughes and Woodward, 2008). Periglacial activity followed a similar pattern, expanding down-valleys during glaciations and shifting to higher elevations during interglacial periods (Oliva et al., 2016b). Though periglacial processes in Mediterranean mountains are currently mostly related to seasonal frost, environments affected by permafrost conditions in the past were significantly larger as suggested by a wide range of records that will be examined in detail in this research.

3. Methodology

This paper presents a detailed analysis of the published scientific literature on permafrost evolution in the Mediterranean region since the Last Glaciation. To better examine spatial and temporal patterns of permafrost conditions within the region, the Mediterranean basin has been subdivided in the following areas: Iberian Peninsula, southern Alps, Italian Peninsula, Balkan Peninsula, Anatolian Peninsula, northern Africa and Mediterranean islands (Figure 1).

For each of the areas we have conducted a detailed review of all the existing knowledge on permafrost conditions focusing on five main periods: Last Glaciation, deglaciation, Holocene, LIA and present-day. We establish the chronological boundary between the Last Glaciation and the deglaciation at 19-20 ka, after the LGM, when Clark et al. (2009) proposed the onset of deglaciation in the Northern Hemisphere, including also the Mediterranean region. We recognise that glaciers reached their maxima at different times during the Last Glaciation and glaciers in some areas retreated to smaller positions earlier than others, but the widespread deglaciation occurred after 19-20 ka in all parts of the Mediterranean mountains. The deglaciation extends until the Holocene and is divided considering different colder/warmer periods, sometimes characterised by phases of glacier advance/retreat, namely the Oldest Dryas (OD; 17.5-14.7 ka, stadial GS-2.1a), Bølling-Allerød (BA; 14.7-12.9 ka, interstadial GI-1) and Younger Dryas (YD; 12.9-11.7 ka, stadial 1 - GS1). The Holocene is subdivided following Walker et al. (2012) who established the limit between Early-Middle Holocene at 8.2 ka BP and the boundary between the Middle-Late Holocene at 4.2 ka BP. The LIA is defined as a single period within the Late Holocene encompassing from the 14th to 19th centuries, followed by the post-LIA warming. The characterization of current permafrost distribution and associated geomorphic processes in Mediterranean mountains allows comparing present-day dynamics with past permafrost conditions.

All these data have been summarized in a table including the location of the massifs, topographic characteristics (elevation, aspect), geomorphic evidence, available chronology (if existing) and associated references for each region. The distribution of permafrost-related features in each of the areas for each of the study periods has been mapped in GIS environment.

Rock glaciers constitute the most frequent and reliable information on palaeo-permafrost occurrence in Mediterranean mountains (Oliva et al., 2016b). Several fossil rock glaciers have been dated by analyzing the exposure of surfaces to cosmogenic radiation, usually by the presence of ^{36}Cl or ^{10}Be isotopes in the rock surface (i.e. Çiner et al., 2017). Two approaches have been generally used for inferring the age of rock glaciers: (i) their maximum age of formation has been often determined from samples collected from polished bedrock outcrops in which the rock glacier lies, (ii) their minimum age of stabilization has been also established based on samples taken from the surfaces of boulders distributed from the front of the rock glacier to the roots in the slope. Therefore, following this approach the dated samples should yield younger ages from the rock bedrock outcrops towards the roots of the rock glacier, which in fact has happened in most cases. However, there are still many unsolved issues concerning the application of surface exposure dating on rock glaciers, mainly related to: (i) the timing between boulders (i.e. rock glacier) stabilization and permafrost disappearance, (ii) the paleoclimatic significance of rock glacier activity in very active geomorphological settings experiencing paraglacial readjustment (Ballantyne et al., 2009; Palacios et al., 2016, 2017b; Andrés et al., 2018), (iii) the problem of excessive inherited nuclides yielding too old ages (Ivy-Ochs et al., 2009; Çiner et al., 2017) contrary to the moraines where glacial carving removes the cosmogenic inheritance, and last but not least (iv) the problem related to toppling and/or erosion of boulders that result in too young ages (e.g. Moran et al., 2016).

4. Geomorphological and sedimentological evidence

4.1 Iberian Peninsula

The Iberian Peninsula constitutes the SW tip of the Eurasian continent between the Atlantic Ocean and the Mediterranean Sea. Relatively flat terrain is mostly located in the central plateaus and coastal plains, separated by several mountain ranges aligned W-E with the highest peaks exceeding 2000 m: NW ranges, Cantabrian Mountains, Pyrenees, Central Range, the Iberian Range and the Betic Range (Figure 2). The rough terrain together with the high elevations have conditioned the magnitude and extent of glacial and periglacial processes during the Quaternary, including substantial variations of the spatial domain of permafrost through time. Sierra Nevada, in the Betic Range (Mulhacén, 3478 m), and the Maladeta massif, in the Pyrenees (Aneto, 3404 m) encompass the highest elevations in Western Europe outside the Alps.

A wide range of lithologies is found across Iberia. While some mountain regions show a relatively homogeneous composition - such as the NW ranges, Central Iberian Range and Iberian Range with prevailing metamorphic and crystalline rocks - or the Cantabrian Mountains with widespread thick Mesozoic carbonates -, others include a variety of rocks – such as the Pyrenees and the Betic Range with alternating Palaeozoic crystalline rocks and Mesozoic carbonates.

Quaternary periglacial activity has been more or less intense and extensive in the Iberian Peninsula in response to prevailing climate conditions. The intensity and duration of the cold determined if periglacial processes were conditioned by the presence of permafrost or seasonal frost conditions. This is also reflected in the widespread geomorphological and sedimentological features of periglacial origin in mid and high mountain environments in Iberia, but also at lower elevations such as in the NW corner and Cantabrian coast where periglacial activity almost reached sea level.

Last Glaciation

Although the chronology of periglacial activity in the Iberian Peninsula during the Last Glaciation is still uncertain (Oliva et al., 2016a), it is likely to consider that its extent and intensity followed periods of glacial advance and retreat. In this sense, the local Maximum Ice Extent (MIE) occurred asynchronously across Iberian mountain ranges: while the combination of cold and moisture was more effective for glacier development in NW Iberia between 35 and 45 ka cal BP (Valcárcel, 1998; Jiménez and Farias, 2002; Moreno et al., 2010; Serrano et al., 2012, 2013, 2015; Jiménez et al., 2013; Rodríguez-Rodríguez et al., 2014, 2016; Nieuwendam et al., 2016; Ruiz-Fernández et al., 2016), in the Central and some sectors of the Eastern Pyrenees it occurred prior to the LGM, probably at approximately 60 ka cal BP (García-Ruiz et al., 2003, 2010, 2013; Lewis et al., 2009; Delmas, 2015), in Sierra Nevada it took place around 30-32 ka (Gómez-Ortiz et al., 2012a, 2015; Oliva et al., 2014; Palacios et al., 2016) and in the rest of high mountain ranges, also in the Eastern Pyrenees, was slightly earlier or almost synchronous with the Last Glacial Maximum (LGM) (Pallàs et al., 2006; Domínguez-Villar et al., 2013; Hughes et al., 2013; Pedraza et al., 2013; Carrasco et al., 2015; Palacios et al., 2015). During the coldest stages of the Last Glaciation (e.g. MIS 2) glaciers extended across the highest mountain ranges in Iberia and periglacial processes expanded to the lowlands; warmer phases favoured periglacial activity only in mountain environments in parallel to glacier shrinkage.

Periglacial processes in mid and low environments were mainly driven by seasonal frost conditions, though some researchers have found evidence of ice wedge development in Pleistocene fluvial terraces between 200 and 1000 m in the Tajo and Duero basins, central Iberia, which may be related to very intense periglacial conditions but probably not associated with a permafrost regime (Badorrey et al., 1970; Asensio-Amor and González-Martín, 1974; Serrano et al., 2010a). Taking into account the presence of block streams and rock glaciers, the non-glaciated slopes of Iberian mountains were affected by permafrost conditions above 700-1300 m in northern mountain ranges (Figure 2), significantly higher in lower latitude mountain ranges, such as the Central Range (1800 m) and Sierra Nevada (2500 m) (Table 2). Sediments of periglacial origin, such as stratified and head deposits, could have been also produced under permafrost conditions, namely in the highest slopes and during the coldest stages. The presence of meter-sized patterned ground features, such as the sorted circles existing across the summit plateaus of most of the mountain ranges (Pyrenees, Sierra Nevada, Cantabrian Mountains), suggests also very intense periglacial conditions in these relatively flat areas where wind redistribution did not favour snow accumulation.

Figure 2

Deglaciation

As in many other glaciated environments in the Northern Hemisphere (Clark et al., 2009), a rapid deglaciation process started in Iberian mountains around 19-20 ka (Hughes and Woodward, 2008; García-Ruiz et al., 2010) at the same time as in mountain areas bordering the NE Atlantic to the north (Hughes et al., 2016). The terrain exposed by retreating glaciers was subject to paraglacial activity and intense periglacial processes under permafrost conditions. Many rock glaciers, protalus lobes and few block streams formed in recently deglaciated environments, though most of them became gradually inactive as temperatures kept raising, namely those located at lower altitudes (Table 2).

Table 2

However, the deglaciation was not a linear process and glacial expansion occurred in the highest mountain ranges during two major cold periods before the Holocene, namely the OD and YD. Following each of these cold stages, rock glaciers developed inside the cirques and highest slopes. As temperatures rose, rock glaciers became inactive at lower elevations with activity only at higher locations; in the case of the Pyrenees, rock glaciers formed during the OD at elevations above 2250 m with widespread permafrost above 2490 m, while during the YD rock glaciers developed above 2350 m and permafrost was limited to elevations above 2525 m (Oliva et al., 2016a). Many rock glaciers in the Cantabrian Mountains have been assigned to this period (Redondo et al., 2010; Gómez-Villar et al., 2011; Pellitero et al., 2011; García-Ruiz et al., 2016a,b), as well as in the Iberian Range, where several rock glaciers were also related to this period in the Cebollera Sierra (Ortigosa, 1986), and in the Demanda Sierra (García-Ruiz et al., 1979; Fernández-Fernández et al., 2017). Surface exposure dating suggests that rock glaciers were active in the Eastern Pyrenees between 15 and 10.5 ka (Palacios et al., 2015; García-Ruiz et al., 2016a,b) and in the Sierra Nevada from 12.8 to 6.4 ka (Gómez-Ortiz et al., 2012a; Palacios et al., 2016). Patterned ground probably related to permafrost conditions may have been active during the coldest stages of the OD and YD in the high lands of the highest mountain ranges.

Holocene

The onset of the Holocene saw the last massive deglaciation in Iberian mountains that was parallel to a progressive shift of periglacial conditions to higher elevations (Oliva et al., 2011). It is unlikely that the atmospheric temperatures recorded in Iberian mountains during the Holocene favoured the existence of widespread permafrost regimes (Oliva et al., 2016b), with the only exception of the highest massifs in the Pyrenees (Serrano, 1998). In the other highest mountain ranges permafrost conditions during the Holocene may have been marginal and only related to favourable geomorphological settings.

The Pyrenees, Cantabrian Mountains and Sierra Nevada were the only environments encompassing glaciers during the coldest stages of the present-day interglacial. Apart from the LIA glaciation, Gellatly et al. (1992) dated a cold phase in the Troumouse cirque (French Pyrenees) between 4.6 and 5.1 ka cal BP and García-Ruiz et al. (2014) reported evidence of glacier activity in the Pyrenees during the Neoglacial (5.1, 3.5 ka) and Dark Ages (1.4-1.2 ka) periods. Based on denudation rates, debris supply from cirque walls and flow displacement rates, Serrano et al. (2006, 2011) suggested that some active rock glaciers today in the Pyrenees (i.e. Argualas, Bastampé, Besiberri and Guerreys) probably developed between 3.4 and 6.2 ka; therefore, during part of the Mid Holocene permafrost environments may have existed in the highest mountains. This stage of rock glacier development seems to correlate with Neoglacial cold phases favouring glacial expansion in the Troumouse cirque (Gellatly et al., 1995) and Monte Perdido massif (García-Ruiz et al., 2014). The morphology and internal stratigraphy of the rock glaciers, some of which reshaped by LIA glacial advance, suggests that at least seven currently active rock glaciers originated during the Mid Holocene (Serrano et al. 2002, 2010b, 2011). Ice caves in the Central Pyrenees also started forming during the Mid Holocene, namely between 6 and 2 ka cal BP (Sancho et al., 2016).

In Sierra Nevada, Oliva and Gómez-Ortiz (2012) also identified glacial advances during the Neoglacial period (2.8-2.7 ka cal BP) and Dark Ages (1.4-1.2 ka cal BP). These glacial conditions were accompanied by an extent of periglacial activity, enhanced erosion processes and decrease of vegetation cover (Oliva-Urcía et al., 2013), with permafrost conditions only in the highest parts of the glacial cirques (Oliva and Gómez-Ortiz, 2012). The existence of permafrost was confirmed by surface exposure dating in some areas such as in Sierra Nevada, where rock glaciers were active until the Mid Holocene in favourable topographic environments inside the YD glaciated cirques (Palacios et al., 2016). In some cases, the development of protalus lobes during the Holocene may be also indicative of permafrost conditions (Table 2).

LIA

In Iberia, colder than present-day climate conditions accompanied by fluctuating precipitation prevailed during the LIA (Barriendos, 1997; Rodrigo, 1999). These conditions led to glacial advance in the Pyrenees, as well as the appearance of small glacial spots in the Cantabrian Mountains and Sierra Nevada (González-Trueba, 2006, González-Trueba et al., 2008; Gómez-Ortiz et al., 2006, 2009; Oliva and Gómez-Ortiz, 2012). Periglacial conditions expanded down-valleys and more extensive snow fields remained during the summer season. Permafrost conditions reactivated and new protalus lobes and rock glaciers formed in the Pyrenees above 2560 m (Serrano et al., 2001, 2011; Fernandes et al., 2017). Several currently active rock glaciers are linked with LIA lateral moraines, suggesting their origin following the LIA in historical times. All of them are relatively small and influenced by glaciers, which indicates the elevation boundary of permafrost conditions above 2560 m (Serrano, 1998; Serrano et al., 2004, 2011; González-García, 2014). With the exception of the surroundings of the glaciated areas and highest periglacial environments, it is unlikely that permafrost existed in other areas.

Present-day

Temperature increase in Iberian mountains since the last decades of the XIX century has been quantified in about 1 ºC (González-Trueba et al., 2008; Oliva and Gómez-Ortiz, 2012). The 0 ºC isotherm lies today close to the top of the highest peaks in the Cantabrian Mountains (2500 m; González-Trueba et al., 2007) and Sierra Nevada (3400 m; Oliva et al., 2016b), which explains the inexistence of widespread permafrost conditions. In these massifs, the bottom of the northern cirques that held glaciers during the LIA encompasses buried ice and permafrost covered by debris left by paraglacial dynamics (Serrano et al., 2011b; Gómez-Ortiz et al., 2004; Ruiz-Fernández et al., 2016; Pisabarro et al. 2016). In the Veleta cirque, in Sierra Nevada, a rock glacier formed during post-LIA deglaciation shows multiple subsidence and collapses as a result of the accelerated melting of the frozen body existing in its interior (Gómez-Ortiz et al., 2014).

In the Pyrenees, BTS measurements, geophysical surveying and geomorphological observations made in several massifs indicate that mountain permafrost regime occurs today above 2630 m in northern aspects and 2800 m in southern ones (Serrano et al., 1999, 2001, 2002, 2006, 2009, 2011a; González-García et al., 2014). Possible permafrost conditions have been estimated above 2400 m in northern slopes and above 2650 m in southern ones, with high variability depending on the massifs. While possible permafrost in the Infierno massif is located above 2450 m, in the Maladeta it exists above 2760 m and in the Posets above 2800 m (Serrano et al., 2001, 2009; González-García, 2014). The same variability is detected in the probable permafrost areas, located generally above 2700 m, oscillating between 2630 m in north faces and above 2800 m in southern ones. While in the Infierno massif the probable permafrost lies above 2650, in the Maladeta it sits above 2890 m and in the Posets above 2950 m. Three types of high mountain permafrost have been differentiated: (i) climatic origin, as suggested where MAAT are around -1/-2 ºC; (ii) topoclimatic, located in lower environments where cirque walls condition low incident radiation; and (iii) morphodynamic, where buried ice existing below the debris cover favours permafrost aggradation (González-García, 2014). In recently deglaciated environments there are landforms linked to frozen bodies, such as protalus lobes and frost mounds, as well as other, such as patterned ground features and solifluction lobes located always around 2700-3000 m, which may be also associated with permafrost (Serrano et al., 2000, 2001; Feuillet, 2010; Feuillet and Mercier, 2012; González-García, 2014; González-García et al., 2017).

Both in the Cantabrian Mountains (Gómez-Lende et al., 2014) and the Pyrenees (Bartolomé et al., 2015) there are also other landforms including subsurficial frozen features, which are an inheritance from the LIA. Ice caves show bedrock recording below freezing mean annual ground temperatures in contact with the perennial ice, and therefore may be defined as permafrost. In the case of the Cantabrian Mountains, the organic remnants trapped in the ice provided ages between 200 and 600 cal yr BP (Gómez-Lende, 2015), whereas in the Pyrenees ranged between 200 and 1200 cal yr BP (Bartolomé et al., 2015; Leunda et al., 2015; Sancho et al., 2016).

Figure 3

4.2 Southern Alps

The Southern Alps constitute the south side of the European Alps and include only the southern slope of the highest summit of the European Alps like the Monte Bianco (4810 m), Monte Rosa (4634 m) but also entirely some other peaks around 4000 m (like Gran Paradiso 4061 m or Ortles 3905 m). The Southern Alps are crossed by large valleys mainly W-E oriented (Aosta Valley, Valtellina) or N-S (Ossola Valley, ValCamonica, Adige Valley; Piave Valley or Tagliamento Valley). The Southern Alps can be divided into four main sectors: the Southern sector or Maritime Alps located between Italy and France south of Maddalena Pass; the western sector located between Maddalena Pass and the Ossola Valley; the central sector between the Ossola Valley and Adige Valley and the Eastern sector eastward of Adige Valley. These sectors reflect different local climate and different prevailing lithologies.

The southern and western sectors are characterized mainly by prevailing metamorphic and crystalline rocks while in the eastern sector Mesozoic carbonates prevail. Within the central sector the southern areas are dominated by Mesozoic carbonates whereas the northern areas by metamorphic and crystalline rocks. The differences in climate of these sectors are mainly related to the variation in the precipitation regime that reflects the main orientation of the mountain belt. In general, the outer parts of the mountain belt (southern and the more eastern areas of the Eastern sector) are the wettest while the northern (and inner) areas of the central sector are the driest.

Last Glaciation

There is still considerable uncertainty about the timing of the last glacial stadials recorded in the Alpine end-moraines at about 18-21 ka cal BP before the onset of glacial termination (Ivy-Ochs et al., 2008), although the LGM culminations in the southern Alps may span from 26 to 21 ka cal BP (Monegato et al., 2007, 2017; Ravazzi et al., 2012; Mozzi et al., 2013; Rossato et al., 2013; Federici et al., 2017). Paleo-temperatures between 30 and 17 ka BP have been recently reconstructed in NE Italy through the analyses of chironomids of lacustrine sediments series (Lake della Costa, Samartin et al., 2016). These authors associated the lack of chironomids remains around 24 ka and 18 ka cal BP to very cold conditions correlated with Heinrich events (HE; Heinrich, 1988) HE-2 (24.2 ± 3.8 ka cal BP) and HE-1 (16.9 ± 3.3 ka cal BP) (Hemming, 2004). In these events possibly a weakening of the Mediterranean low-pressure systems occurred with a higher aridity in the Mediterranean (Tzedakis et al., 2004; Kwiecien et al., 2009). On the other hand, a southward shift of the storm track between 26.5-23.5 ka cal BP has also been suggested, together with a change in the seasonal distribution of precipitation, predominantly occurring between spring and autumn (Luetscher et al., 2015).

Evidence of active periglacial processes at low elevations as well as in the Po plain are almost unknown. Nevertheless, two stratified scree deposits have been described in the Maritime Alps (Pappalardo, 1999) with stratified screes, partially cemented, on the southern slopes of M. Cornizzolo (1242 m) and on M. Barro (923 m) very close to Lake Como. Possible cryoturbations on the loess cover at Bagaggera (310 m), only few km south of Lake Como, have also been detected (Guglielmin, unpublished data). An interesting indicator of periglacial or cryotic conditions during the Late Pleistocene and LGM is the distribution of loess. According to Cremaschi et al. (2015) between 70 and 35 ka cal BP large loess deposits were deposited along the northern, western and southern margins of the Po plain. In the Julian Alps, between Italy and Slovenia, four rock glaciers have their front at 1076 m on average (Colucci et al., 2016a), which is very close to the estimated ELA during the LGM (1200 m, Colucci et al., 2014). This also suggests a temperature drop of at least 7.1-7.6 °C, which is in accordance with LGM temperature reconstructions for this sector of the Alps (Kuhlemann et al., 2008).

The presence of relict block streams and blockfields is also reported (Figure 4), and the largest area covered by such features (140 km^2) is the Ultramafic Lanzo Complex a short distance from Turin, at an elevation of 1000 to 1650 m. These landforms could be related to permafrost conditions during the LGM (Fioraso and Spagnolo, 2009; Paro, 2011). Other block streams and blockfields are located in the internal margin of Piedmont between Sangone valley (50 km S Turin) and Oropa (150 km NE of Turin) (Paro, 2011).

Figure 4

Deglaciation

The onset of glacial retreat in the southern Alps is thought to have occurred at 20.8 ± 1.5 ka (Gianotti et al., 2008), almost simultaneously with the Fennoscandinavian Ice Sheet (FIS) (A.L. Hughes et al., 2016). Downwasting of the large piedmont glaciers back into the Alpine valleys during the Late Glacial promoted widespread paraglacial activity, although several glacial readvances took place between the OD and the YD (Ivy-Ochs et al., 2008, 2009). Remnants of moraines of the YD glacier advance have been morphologically associated to rock glaciers from the lowermost belt (Frauenfeder et al., 2001; Ivy-Ochs et al., 2009).

Rock glaciers constitute widespread periglacial landforms in the southern Alps, both in the form of active and inactive landforms (Table 3). At present, climatic conditions favourable for their activity range between 2400 and 2900 m (Guglielmin and Smiraglia, 1997; Seppi et al., 2012; Scotti et al., 2013; Colucci et al., 2016a). Relict forms are generally 400-500 m lower than the active ones, corresponding to a temperature drop of ca. 2.6-3.3 °C (Frauenfelder et al., 2001). Also for this reason, despite the low number of dated rock glaciers in the southern Alps, YD or early Holocene ages have been suggested for most of them. The mean altitude of the front of the YD rock glaciers in the Alps is found at the lowest elevations in the SE fringe of this mountain range, with fronts distributed at 1777 m (Colucci et al., 2016a).

Table 3

In the SW Alps (France), more than 1000 relict rock glaciers have been identified by Bornet et al. (2014). The mean altitude of their fronts is around 2260 m, whereas a few of them (7) reach altitudes lower than 1800 m, with the lowest one lying at 1440 m. It is probable that most of those relict rock glaciers are of YD age, though no absolute dating is yet available. A YD age is in good agreement with the general scheme of Late Glacial rock glaciers formation further North (45°N) proposed by Cossart et al. (2010). Nevertheless, the authors emphasize the poor climatic significance of those permafrost-related landforms, because they developed at higher altitude than the local YD moraine and disconnected from talus slopes, therefore their extent has probably been limited by the availability of sediment sources. These conclusions are consistent with several geoelectrical resistivity soundings carried out on rock glaciers in different sectors, from the Col du Lautaret (near Briançon village) to the Mercantour area (data partly reviewed by Ribolini and Fabre, 2006). Overall, looking at the minimum altitude of the relict rock glacier fronts in the southern Alps (Guglielmin and Smiraglia, 1997; Seppi et al., 2012; Scotti et al., 2013; Bornet et al., 2014; Colucci et al., 2016a), there is a clear lowering trend from the western Alps (2260-2340 m) to the central Alps (2170-2281 m) and the southeastern Alps (1778 m) (Figure 5). Although the Maritime Alps do not follow this scheme, this overall pattern in rock glacier elevations follows the ELA depression observed in the eastern Alps compared to the western Alps at the LGM peak (Kuhlemann et al., 2008).

Figure 5

Holocene

According to Finsinger and Tinner (2006), during the Early Holocene two cooling phases (8.8-7.3 and 6.1-5.2 ka cal BP) occurred in the western Alps. The cold event at 8.2 ka cal BP has also been documented in the eastern Alps with ca. 3°C below the early Holocene thermal maximum (Ilyashuk et al., 2011), which has been reconstructed between 9 and 5 ka BP with an estimated mean July temperature 1-2°C warmer than the recent pre-industrial period (Samartin et al., 2017). Since 4.5 ka BP, with the exception of the end of the 20th century, a progressive cooling trend occurred (Ilyashuk et al., 2011). During the Holocene glaciers showed several advance stages that were not synchronous within the Alps, although most of the glaciers at the onset of the Holocene recorded a rapid shrinkage to a size generally smaller than their late 20th century size until 3.3 ka cal BP (Ivy-Ochs et al., 2009). Subsequently, glaciers showed a widespread re-advance especially between 3.0 and 2.6 ka cal BP and between 1.4 and 1.2 ka cal BP with the final major last re-advance during the LIA between 0.45 and 0.1 ka cal BP (Ivy-Ochs et al., 2009).

With regards to the chronology of Holocene periglacial activity in the southern Alps, there are only a few available ages for periglacial and permafrost landforms. A cooler and dryer period ca. 4.5 ka cal BP must have favoured permafrost aggradation and rock glaciers development in the southern Alps (Ilyashuk et al., 2011). The re-calibraton with OxCal 4.3.1 (Bronk Ramsey, 2009) of the available radiocarbon ages of inactive rock glaciers (Calderoni et al., 1998; Dramis et al., 2003) in the central and western Italian Alps using the IntCal 13 dataset, suggests that the period of maximum activity occurred between 2720 and 2850 cal yr BP with some older exceptions (all younger than 5900 cal yr BP). On the other hand, based on the existence of a paleosoil buried by the rock glacier front, Calderoni et al. (1998) and Scapozza et al. (2010) showed evidence of their activity between 929 and 1374 cal yr BP, a similar age to that reported by a Salix sp. leaf trapped within the ice of the Foscagno active rock glacier dated at 919-961 cal yr BP (Stenni et al., 2007). Recently, Krainer et al. (2015) found that the age of ice within an active rock glacier in the central Italian Alps ranged between 8960 and 2240 cal yr BP at 23.5 and 2.8 m depth, respectively, showing evidence that the ice preserved within the rock glaciers can be useful paleoclimate archives. The investigation in fact shows that frozen material, which had existed during most of the Holocene, is now thawing; this is strong evidence that temperatures during the past about 9 ka have never been as high as they are today.

LIA and present-day

Widespread wetter and cooler condition over the Alps promoted a well-documented glacier advance during the LIA (~1400-1860 AD; Ivy-Ochs et al., 2009). After the LIA, the warming trend in the Alps has been about twice the global trend over the time period 1906-2005 (1.4°C; Brunetti et al., 2009) and in the SE Alps temperature at 2200 m increased by about 1.7°C from 1851 to 2012 (Colucci and Guglielmin, 2015) leading to a 96% reduction in volume of glacial bodies (Colucci and Žebre, 2016). In the French western Maritime Alps, geophysical investigations provided resistivity values coherent with permafrost existence at about 2550 m in the Mercantour sector (Evin and Fabre, 1990). In the Ubaye-Queyras sector, rock glaciers below 2400 m lack internal ice, while above 2500 m permafrost starts to be common (e.g. Marinet rock glacier; Evin and Fabre, 1990; Evin et al., 1990; Assier et al., 1996; Smiraglia et al., 1996; Ribolini et al., 2010). This indicates that 2500 m represents here the modern limit for the existence of active rock glaciers. In the Lautaret and Mt. Viso areas, active rock glaciers were found at 2400-2450 m (Evin, 1991; Francou and Reynaud, 1992; Bodin et al., 2009). According to Bornet et al. (2014) more than 50% of the active and inactive rock glaciers (following Barsch, 1996) in the SW French Alps, from a total number of 780 inventoried landforms, are located above 2600 m and 51% of them are lying on N and NW facing slopes. They cover 2% of the total surface (50 km²) and mostly develop between 2100 and 3000 m. Most of the active rock glaciers in the southern French Alps are smaller than 0.1 km², whereas the relict ones are generally larger (up to a few km²). More than 80% of the active and inactive rock glaciers are probably located below the regional mean annual 0°C-isotherm, raising the question of their stability under climate change (Bodin et al., 2015). This probable imbalance with the present climatic conditions may be responsible for the fast flow observed on some very active rock glaciers (Delaloye et al., 2008), and may also have led to the collapse of the Bérard rock glacier during summer 2006 (Bodin et al., 2017).

In the western Italian Alps active rock glaciers reach with their fronts a mean elevation of 2647 m although a few active rock glaciers appear sporadically between 2000 m and 2300 m. In the Central Italian Alps the mean elevation reached by the active rock glaciers is lower (2526 m) but ice was found within rock glaciers locally also at very low elevation in ventilated talus (1000 m; Guglielmin, unpublished data). As well known the lower altitudinal limit of permafrost based on the elevation of the active rock glaciers, although investigated through geophysical methods, can be not representative of the real regional permafrost; in fact, the lower limit can decrease up to 200 m due to the local density-driven air flows (Balch, 1900). Several rock glaciers have developed since the LIA in the Adamello-Presanella area (eastern Italian Alps) where the current mean elevation of the fronts of active rock glaciers is 2527 m, around 400 m above the limit of the relict ones and where patches of sporadic permafrost can reach an elevation of 2200 m (Baroni et al., 2004). Only a few active rock glaciers occur in the western Alps within LIA deglaciated areas; this is the case of an active rock glacier in the Dolomites (SE Alps) at about 2400 m forming in an area previously occupied by the tongue of a LIA glacier and now dominated by periglacial processes. This study provides an example of how periglacial processes are replacing glacial processes under present climate change during the paraglacial stage (Seppi et al., 2014). Recently, possible sporadic permafrost in an area occupied by two protalus rampart have been deduced based on BTS measurements in the Carnic Prealps at 2258 m (Colucci et al., 2016a). Nine active protalus and pronival ramparts, located in front of permanent snow/firn bodies and ice patches, have been also reported in the Julian Alps and in the Kamnik-Savinja Alps at an average altitude of 2158 m (Colucci, 2016).

Although in the Alps permafrost is limited to the highest and driest areas, ground ice is also reported from several caves (Figure 6). Ice caves are generally widespread in high elevated karstic environments of the Alps, mainly in the central and eastern parts. The real distribution and size of ground ice in Alpine caves is still very uncertain, although recently in the southeastern Italian Alps 1100 ice caves have been reported. Here, they are generally located between 1500 and 2200 m and their distribution is well-correlated with MAAT and altitude, being significantly more frequent for MAAT lower than 2 °C and 5°C; therefore, they are placed within the periglacial domain characterized by high precipitations (Colucci et al., 2016b). In Slovenia, 551 ice caves have been reported by Mihevc (2008). According to the Slovenian Cave Registry (2014), approximately 520 cryo caves, of which ca. 100 ice caves, can be found in the Slovenian Alps. Most of them have the entrance between 1000 and 2400 m (Mihevc, 2008). The radiocarbon age of an insect found in an ice core extracted in 2003 from the Abisso sul Margine dell'alto Bregai (central Alps) reported an age of 185 ± 30 cal yr BP (Citterio et al., 2005). In the cave *Ledena jama pri Planini Viševnik* (eastern Alps), an ice core was recently extracted and larch trapped in the ice yielded a radiocarbon age of 300 ± 45 cal yr BP (Staut et al., 2016). In the light of accelerated climate change, other important paleoclimate archives exploitable from ice caves are represented by *in-situ* coarse cryogenic cave carbonates, recently discovered for the first time in a vertical ice outcrop of a cave in the southeastern Alps (Colucci et al., 2017).

Figure 6

4.3 Italian Peninsula

The Italian Peninsula lies between the Tyrrhenian Sea on the west, the Ionian Sea on the south, and the Adriatic Sea on the east. It includes large plains of northern Italy (Po Plain and Venetian Plain) and extends over more than 1200 km crossing 8° of latitudes between 38 and 46 ºN. The Apennine Mountains run along the entire peninsula and culminate in the Gran Sasso d'Italia (2912) and La Majella (2793 m). Geologically, the Apennines are quite homogeneous with a clear dominance of Mesozoic carbonates although sandstones, shales and flysch are also found, especially in the northern Apennines.

In the Italian Peninsula, from the Late Pliocene/Early Pleistocene transition the climate was characterized by rapid succession between xeric cool (glacials) and humid warm (interglacials) phases (Bertini, 2010). These alternating climatic conditions favoured the existence of "warm" steppes or coniferous forests alternated with subtropical to warm-temperate deciduous forests. During these periods, the paleogeography of the Italian coasts changed significantly due to sea level variations; indeed, the coastline of Adriatic Sea moved hundreds of km southward during the LGM and did not reach the current position until ca. 7 ka (Lambeck et al., 2011; Anzidei et al., 2014).

Last Glaciation

In the Apennines valley glaciers were only present on the highest mountains, with the maximum glacial expansion dated at 27.2 ± 0.9 ka cal BP in the Campo Imperatore valley and a rapid glacial retreat at around 21 ka cal BP (Giraudi and Frezzotti, 1997). Subsequently, lake level oscillations of Fucino Lake showed an aridity period between 20 and 17 ka cal BP and between 15 ka cal BP and the Early Holocene (Giraudi, 1998). More recently, at Campo Felice (central Apeninnes) Giraudi (2012) suggested that the maximum glacial advance was dated earlier than in other areas (33-27 ka cal BP) driven by a relatively milder and wetter period. Kuhlemann et al. (2008) modelled the air circulation and the ELA of the glaciers in the Mediterranean area indicating possible conditions for glaciers occurrence also on Mt. Pollino in Calabria and on Mt. Etna in Sicily.

Several landforms were related to periglacial or permafrost conditions during the LGM or during the deglaciation process, particularly during the OD in Italy (Figure 7). In the northern Apennines only a couple of relict rock glaciers described by Chelli and Tellini (2002) were associated to

LGM period on the northern slopes of the highest mountains of this belt between Mt. Cusna (2121 m) and Mt. Cimone (2165 m). Rock glaciers from this period are more frequent in the central Apennines, where 38 relict rock glaciers were identified by Giraudi (2002) almost all located along the eastern side of the central Apennines. They are mainly present on the Gran Sasso Massif (2912 m), the Maiella Massif (2793 m) and Mt. Velino (2486 m). In the southern Apennines only one relict rock glacier has been detected on the northern slope of Mt. Pollino (2267 m).

Figure 7

Block streams from this stage are located around Mt. Beigua (1287 m), few km far from the Ligurian Sea. These landforms are located between 900 and 1100 m on all the slopes and are characterized by an open-work texture in the upper 1.5 m thickness with angular or subangular blocks and frequent vertical dipping, or imbrication of the boulders (Firpo et al., 2006). Remarkably, none has any rock cliff or free face at their head and they show generally the smallest blocks in their frontal part. These landforms are all shaped on the same bedrock (serpentine and metabasalts) and they could have formed at a maximum age of 33 ka (Firpo et al., 2006). Moreover, in the upper parts of the Mt. Beigua thors and blockfields are also present, suggesting intense periglacial conditions (Firpo et al., 2006).

In the Apennines stratified scree deposits (gréeze litée) are the most widespread and studied periglacial sedimentary archives (Table 4). They were found in different localities of the Northern Apennines, such as Corniglio and Bratica Valley (Chelli and Tellini, 2002), Monte Prampa (Bernini et al., 1978), Alpi Apuane (Federici, 1981) and, above all, in the central Apennines (Coltorti and Dramis, 1988; Coltorti et al., 1983) and in different valleys in the Marche region. In the central Apennines stratified screes were described in the late 1970's mainly along the Adriatic side of Umbria-Marche Apennines (Coltorti et al., 1979, Castiglioni et al., 1979; Boenzi, 1980), where these deposits are particularly widespread and connected with outcrops of limestone and marly limestone bedrock (Coltorti et al., 1983). These deposits are documented at different elevations from the coast to more than 2000 m (Mt. Conero) generally at the foot of rectified slopes or down-slope ancient nivation hollows. Features identified as stratified scree were also found southward in Abruzzo (Maielama valley; Frezzotti and Giraudi, 1992; Sulmona valley, Miccadei et al., 1999), Molise (Coltorti, 1983; Scarciglia, 2000) and Puglia regions (Boenzi et al., 1977). In some cases, they were also found close to the coast of the western side of the Apennines on the Palmaria Island, close to La Spezia (Chelli and Pappalardo, 2006) and on the Sorrento Peninsula (Brancaccio, 1968). The southernmost stratified screes were found in Calabria (Mt. Pollino and Mt. Sila down to 500 m; Boenzi and Palmentola, 1975; Praia a Mare and Mormanno down to 700 m; Robustelli and Scarciglia, 2006). In most of the cases, they were probably associated to seasonal frost conditions, with no permafrost regime.

Table 4

Finally, cryoturbations have been also documented on soils of Mt. Beigua at 650 m (Rellini et al., 2014) and at even lower elevations in the Cilento hills (at 30 m asl; Scarciglia et al., 2003) where severe winter frost was hypothesized during the Last Glaciation. Only in one case, in the Sila Mountain at 1350 m, a fossil sand wedge, an ice wedge cast and some other cryogenic features within the soils were described by Dimase (2006).

Holocene

According to Giraudi et al. (2011), the onset of the Holocene in the Italian Peninsula and in Sicily was characterized by dramatic environmental and climatic changes. Between ca. 9.5 and ca. 6-5.5 ka cal BP the climate was generally warmer and wetter than present, although in some records the wettest period seems to have occurred later at ca. 6-5 ka cal BP. Subsequently, a progressive cooling with higher precipitation variability started at 4.5 ka cal BP, with wet periods recorded at 2.8-2.6, 1.4-1.2 and 0.5-0.3 ka cal BP and an aridity peak between 3.7 and 3.2 ka cal BP (Calderoni et al., 1998; Giraudi, 1998). No dated evidence of glacial advances has been documented in the Italian Peninsula during the Holocene except for Gran Sasso area, where the Calderone glacier extends now over a few ha (Hughes et al., 2006a). Here, the glacier is thought to have persisted in this cirque throughout the Early Holocene, although the cirque was probably ice-free between ca. 4.3 and 3.9 ka cal BP (Giraudi, 2004; Hughes, 2006a). Subsequently this glacier experienced several phases of expansion, namely at ca.: 2.9- 2.7, 1.4-1.3 and 0.64-0.58 ka cal BP (Giraudi, 2003, 2004; Hughes, 2006a).

The ages for periglacial landforms developed during the Holocene are rare and limited to few rock glaciers in the central-eastern Apennines. According to these ages, it is possible that permafrost existed during the Early Holocene at elevations higher than 1900-2000 m and started to degrade at ca. 7 ka cal BP (Dramis et al., 2003). Nevertheless, a possible new phase of permafrost aggradation may have occurred at high elevations during the aridity phase that took place between 3.7 and 3.2 ka cal BP (Dramis et al., 2003).

LIA and present-day

It is reasonable to hypothesize that colder conditions prevailing during the LIA favoured the development of some sporadic permafrost at elevations above 2200 m in the central Apennine (Mt. Gran Sasso, Maiella, Mt. Velino). A still active rock glacier is located on the north face of Mt. Amaro, with its front reaching 2522 m (Figure 8) (La Maiella massif, Dramis and Kotarba, 1992). As documented by BTS measurements, at elevations higher than 2400 m in the same massif (Bisci et al., 2003) and locally lower in Mt. Velino permafrost occurrence was possible, until at least 15 years ago (Guglielmin, unpublished data).

Figure 8

4.4 Balkan Peninsula

The Balkan Peninsula (also referred to as Balkans) is situated in south-eastern Europe and is bounded by the Adriatic, Ionian, Aegean and Black Seas. Situated between the Eurasian and African lithospheric plates, the Balkan Peninsula constitutes a mosaic of several mountain systems: Dinaric, Balkan, Macedonia–Rhodopes and Pindus, separated by internal tectonic depressions and major valleys. This internal topographic variability, coupled with general orientation of the mountain ranges, NNW to SSE Dinaric and Pindus and W to E for Bulgarian Mountains, generates a wide diversity of local geographical conditions, reflected in the presence and evolution of permafrost through time. The highest peak of the Balkans is Musala (2925 m), found in the Rila Mountain.

The geology of the Balkan Peninsula is very diverse. For instance, the Dinaric Alps are dominated by up to 8000 m thick Mesozoic carbonates (Jurasic limestone, Triasic dolomite and Werfen sandstones and schists, and Wengen sandstones). The Šar Mountains and Mount Korab on the eastern Albania border consist of polymetamorphic metasedimentary sequence with Hercynian granitoid intrusions in its central part. The major Bulgarian mountains are composed largely of Precambrian and Palaeozoic crystalline rocks (crystalline schist, gneiss, amphibolites, metamorphosed sedimentary rocks etc.) intruded by the Rila-West Rodophian batholith (Sinnyovsky, 2015), in some areas also by crystalline limestone and marble. In Greece the Pindus Mountains and Olympos Massif are dominated by Mezozoic (Lower Jurassic - Upper Cretaceous) sedimentary rock, especially limestone, dolomite and breccia, with ophiolithic rocks also present in some areas, such as at Mount Smolikas and Mount Vasilitsa.

The Balkan Peninsula is one of the Mediterranean regions where information on present and past permafrost conditions is very general and has a speculative character (King and Akermann, 1993; Dobiński, 2005). There is evidence that "isolated patches of permafrost" (Brown et al., 2001) are still present in the high mountain environment of the Balkan Peninsula. However, apart from ice caves (Bočić et al., 2014; Buzjak et al., 2014; Kern et al., 2006a) no other active permafrost features have been recognised in the study area. Relict permafrost features in the Balkan Peninsula have been recognised, although only recorded in detail in the Pindus, Prokletije, Rila and Pirin Mountains, where rock glacier inventories have been compiled (Palmentola et al., 1995; Hughes et al., 2003; Palmentola and Stamatopoulos, 2004; Dimitrov and Gikov, 2012; Gikov and Dimitrov, 2011; Magori et al., 2017). Sand wedges and cryoturbations present in the Pannonian Basin indicate a deep seasonal frost and perhaps isolated patches of sporadic permafrost (Ruszkiczay-Rüdiger and Kern, 2015) in the SW lowlands of the Pannonian Basin and in Getic Basin (Romanian Plain) (Frenzel et al., 1992) during the Pleistocene. However, there is no evidence of continuous permafrost as suggested by some previous studies (e.g., Poser, 1948; Velichko, 1982). At the LGM only the highest mountains of the Central Balkans were located in the discontinuous permafrost zone (Van Vliet-Lanoë and Hallegouët, 2001). Hence, it follows that rock glaciers are the main source of paleo-environmental information to infer past permafrost conditions in low-latitude areas of the Balkan Peninsula, while the ice caves are the best indicators for evaluating the present permafrost distribution in this area. Other elements of periglacial morphogenesis associated with permafrost conditions, such as large and medium size patterned ground and block streams, are very little-studied in the Balkan Peninsula. As is other parts of the Mediterranean region, thick Pleistocene stratified scree deposits are present across the Balkans - especially in the uplands - and attest to active frost action under cold climate conditions. These sediments can be well preserved in limestone caves and rockshelters (e.g. Bailey and Woodward, 1997; Morley and Woodward, 2011), but establishing their paleoclimatic significance is not always straighforward (Woodward and Goldberg, 2001).

Last Glaciation and Deglaciation

The Last Glaciation in the Balkan Peninsula was characterised by the presence of ice fields and valley glaciers over some of the highest mountain massifs (e.g., Milivojević et al., 2008; Hughes et al., 2010, 2011; Kuhlemann et al., 2009, 2013; Žebre and Stepišnik, 2014; Sinnyovsky, 2015; Žebre et al., 2016). Late Pleistocene glaciers reached their maximum extents before the LGM in Pindus Mountains and this is recorded indirectly in the fluvial record downstream where glaciofluvial sediments have been dated at 25-30 ka (Lewin et al., 1991; Woodward et al. 1995; 2008; Hughes et al 2006b). In the Peloponnese, pre-LGM moraines have recently been dated to 30-40 ka using ^{36}Cl exposure dating (Pope et al., 2017). In the Rila Mountains, Bulgaria, it is most likely that two glacial advances occurred, one at the beginning (25-23 ka) and one at the end (16-18 ka) of the global LGM (Kuhlemann et al., 2013). However, the exact timing of the local LGM over the Balkan Peninsula is still a matter of debate with some of the best evidence coming from the fluvial record in NW Greece (Woodward et al., 2008). In the Dinaric Alps, relatively high moisture (Hughes et al., 2010; Žebre and Stepišnik, 2014) was more favourable for glacier development in the highest massifs, meanwhile the lowlands were likely characterised by periglacial environment (Table 5). Almost no research on permafrost features from the Pleistocene period has been presented until now from this area. According to the interpretation by van Vliet-Lanoe et al. (2004) and Ruszkiczay-Rüdiger and Kern (2015), the Dinaric Alps area (at least its NW part) probably hosted discontinuous permafrost, bounded by a seasonal frost or even isolated patches of sporadic permafrost in the Pannonian basin to the NE. In fact, ice wedges and cryoturbations have been found in till in the Lovćen Mountain (Liedtke, 1962), supporting the idea of the permafrost presence in the Dinaric Alps. The presence of tors on the highest peaks indicates intensive periglacial processes in the ice-free areas above the local Equilibrium Line Altitude (ELA).

Table 5

In contrast, in the continental, more arid part of the Balkan Peninsula relict rock glaciers, and also block streams and patterned ground are widespread (Figure 9). Early periglacial studies from the Bulgarian mountains revealed a broad presence of scree deposits, block streams and blockfields (Stoyanov and Gachev, 2012). In addition, 27 and 55 rock glaciers were identified in Rila (Gikov and Dimitrov, 2011) and Pirin Mountains (Dimitrov and Gikov, 2012), respectively, though Magori et al. (2017) recently found new field evidence of the existence of 122 units in both massifs. Rock glaciers are distributed at elevations between 2080 and 2600 m, with a mean altitude of 2340 m in Rila Mountains and 2390 m in Pirin Mountains (Magori et al., 2017). They are considered to be from the end of Pleistocene and early Holocene (Dimitrov and Gikov, 2012; Gikov and Dimitrov, 2011; Kuhlemann et al., 2013), although the spatial relation between different altitudinal groups of landforms suggest other timing interpretation. In the Šar Mountains several rock glaciers, most of them in direct spatial connection with latero-frontal moraines, were recognized above 1700 m by Kuhlemann et al. (2009). Several generations of rock glaciers appear on the northern side of the ridge Bistra (2651 m) – Jezerski (2604 m) between 1750 and 2550 m. A similar situation is found on the north and north-western slope of Mount Korab, where relict rock glaciers are present at altitudes between 1480 and 2600 m. Relict rock glaciers are widespread also in the Pindus and other mountains in Greece. These features have been identified on Mount Parnassus (Pechoux, 1970) in central Greece, and Mount Tymphi (Hughes et al., 2003), Mount Smolikas (Hughes et al., 2006c) and Mount Lakmos-Peristeri (Palmentola and Stamatopoulos, 2004) in northwest Greece. The relict rock glaciers of the Pindus Mountains have wide-ranging elevations and occupy cirques between 1330 and 2300 m (Hughes et al., 2003; Palmentola and Stamatopoulos, 2004; Hughes et al., 2006a) and are likely to represent different generations of rock glacier. Hughes et al. (2003) argued that the rock glaciers of Mount Tymphi must belong to the LGM and formed as climate became colder and drier forcing cirque glaciers to retreat and become overwhelmed by debris. Further research is necessary to test the ages of the rock glaciers in the Balkans.

Figure 9

Besides rock glaciers other permafrost features can also be identified in the inland part of the Balkans (Figure 10). On the Vitosha Mountain, relict stone circles, tors and large blockfields were identified at the elevation range of 1850-1900 m. Some of the largest block streams in Europe, called Zlatnite Mostove (Golden Bridges) and Golyamata Gramada (Big Pile) are also present in this area. The first is 2.2 km long and 150 m wide, descending from 1800 to 1410 m, and the latter 1 km long and 300 m wide at the altitude between 1900 and 1550 m. Block streams are present in the Stara Planina (Balkan) Mountains, descending even down to 930 m in the northern slopes and 1130 m in the southern slopes. They can be found also in the granite part of Rila and Pirin Mountains where they descend down to 1500 m, and in Osogovo Mountain, where they reach 1700 m. Patterned ground and frost-shattered features were noted on Mount Tymphi by Hagedorn

(1969) and Hughes (2004). Separating active or recent periglacial activity from Pleistocene features is difficult for many of the landforms on the highest mountains, although the lower periglacial forms must be Pleistocene in age.

It is certainly true that some rock glaciers form during deglaciation when glacier retreat causes glaciers to become overloaded by debris and morphing into ice-cored rock glaciers (Morris and Olyphant, 1990). It is also true that rock glaciers are most likely to form when climate is cold and dry (Haeberli, 1985). The period characterised by "deglaciation" depends on the timing of the local LGM. In many parts of the world this predates the global LGM, with cold and dry conditions at the global LGM causing glacier retreat (Hughes et al., 2013).

Figure 10

In the Pindus Mountains, on Mount Smolikas (2632 m), a rock glacier occurs at much higher altitude (>2100 m) than on neighbouring Mount Tymphi (1800-2200 m) (Table 6). This led Hughes et al. (2006a) to argue for a YD age for the Smolikas rock glacier. Based on climate modelling, Hughes et al. (2006a) argued that this feature and the associated last ever glaciers on Mount Smolikas could not be Holocene in age. On Mount Tymphi Late-glacial rock glaciers are absent because the level of permafrost was above the cirque floors and consequently, rock glaciers only formed on higher mountains such as Mount Smolikas. On Mount Lakmos-Peristeri, just 25 km southeast of Mount Tymphi, Palmentola and Stamatopoulos (2004) argued for an OD age for the numerous relict rock glaciers in the cirques of these mountains. This was based on earlier suggestions made for similar relict rock glaciers in the Prokletije Mountains, where the study of 16 inactive rock glaciers located above 1700 m was presented by Palmentola et al. (1995). Formation of several relict rock glaciers in the inland Dinaric Alps of Bosnia and Herzegovina and Montenegro (never studied before, but clearly recognized from different satellite images), indicating past permafrost conditions at elevations above 1650 m, may also be related to one of the dry and cold periods of the late Pleistocene. This is in accordance with the recently compiled rock glacier inventory in the nearby SE Alps (Colucci et al., 2016a) where the formation of rock glaciers was related to the YD cold phase. However, in Greece Hughes et al. (2003) argued that similar elevation rock glaciers were older, forming during the LGM. The OD ELA in Šar Mountains was calculated to 2200-2350 m (Kuhlemann et al., 2009). Thus, it is likely that rock glaciers present in the altitudinal belt 1950-2060 m on the northern side of the ridge Bistra (2651 m) – Jezerski (2604 m) belong to this period. During YD, the ELA was placed at 2300-2400, so the rock glacier at 2260-2550 m could belong to this period. Relict rock glaciers on Mount Korab located above 2400 m appear to be fresher (without vegetation) and could belong also to one of these two periods. In the Rila Mountains, rock glaciers above 2200 m and below local ELA are likely to be formed in the deglaciation period. During the deglaciation period, the stadial moraines at 1700 m, 1850 m, and 2000 m point to gradual retreat of the glaciers in the Osogovo Mountain (Milevski, 2008). Embryonic rock glaciers at 1850 and 2030 m in the cirque below the Ruen Peak (2251 m) were likely active during this period. However, none of the rock glaciers in the Balkans have been dated and therefore the competing geochronological hypotheses proposed by Palmentola et al. (1995), Hughes et al. (2003) and Palmentola and Stamatopoulos (2004) remain to be tested.

Table 6

Holocene

Holocene cooling periods in 8.2 ka cal BP (Tonkov et al., 2016), 7.5 ka cal BP (Stefanova et al., 2006) and between 6.7 and 5 ka cal BP (Bozilova and Tonkov, 2000) based on pollen records were reported for the Rila and Pirin Mountains. It is likely that these cooling events triggered the formation of some newly rock glaciers, while the highest lying rock glaciers continued their existence. Rock glaciers in the Rila and Pirin Mountains situated above 2500 m were probably active also during Holocene. One example is the rock glacier on the northern slope of Polezhan Peak (2851 m) with the front at 2490 m where at least three rock glacier generations can be recognised. Also in the Šar Mountains, on the northern slopes of the Bistra (2651 m) – Jezerski (2604 m) ridge, rock glaciers above ~2300 m likely persisted in their active mode during the Holocene, as suggested by Kuhlemann et al. (2009).

Some of the cirque and valley glaciers that were present during the YD likely persisted to Holocene in some areas of the Balkans. Calcite cements in cirque moraines, which provide minimum ages for the moraines, yield early Holocene ages in central Montenegro (Hughes et al., 2011). On Mount Olympus, in the north-facing cirque of Megali Kazania, large moraines were considered to be Holocene Neo-glacial moraines by Smith et al. (1997). This possibility has recently been advocated by Styllas et al. (2015). This would seem plausible given the presence of small glaciers today further north in the Balkans. It is possible that this was also the case in other inland mountains in the Balkan Peninsula, although there is very little known about the Holocene glacial history. Periglacial forms related to these small glaciers, such as nivation hollows, are also poorly understood. However, given the warmer conditions at the Holocene optimum it is arguable that smaller glaciers and snow patches nowadays present in Pirin, Prokletije and Durmitor Mountains (Gachev et al., 2016) would have been less than today.

LIA

Owing to cooling conditions during the LIA there were numerous small cirque glaciers (Hughes, 2010) in the highest massifs of the Balkan Peninsula. There would have also been many more perennial snow fields. Nival processes would have therefore been a significant geomorphological agent across the mountains in the Balkan Peninsula at this time. Rock glaciers above 2350 m in the Rila and Pirin Mountains probably reactivated during the LIA period.

Present-day

Because of relatively low altitude and low latitude, the mountain permafrost in the Balkan Peninsula is nearly inexistent in recent climate. Permafrost may be present in some niche localities on the highest peaks, and this possibility was highlighted in Dobinski (2005) who suggested that permafrost conditions may exist above ~2700 m on Mount Olympus (2912 m), and above ~2350 m on Rila Mountain (2925 m). This idea is partially supported by Brown et al. (2001) modelling results, suggesting that sporadic permafrost is present only in the highest peaks of the southern Dinaric Alps, Šar Mountains, and Rila and Pirin Mountains. In fact, Milivojević et al. (2008) mapped two active rock glaciers in the central Prokletije Mountains. According to Hughes (2009) these features may not be actively moving because they lack the clear lobate form. However, they may be described as patches of sporadic permafrost. Perennial snow fields and modern nival forms are widespread across the highest mountains of the Balkan Peninsula. For example, Styllas et al. (2016) reported perennial snow and ice masses in the north-facing cirque of Megali Kazania (Olympus Mountain). Apart from this 16 glacierets and small cirque glaciers exist in the Durmitor, Prokletije and Pirin Mountains (Hughes, 2008; Djurović, 2013; Milivojević et al., 2008; Hughes, 2009; Gachev and Stoyanov, 2012; Gachev et al., 2016). Some of these ice masses are debris-covered and exhibit similar geomorphology to the buried ice patches of the Corral

Veleta site in the Sierra Nevada, Spain (Gómez-Ortiz et al., 2001). Therefore, the possible presence of patches of sporadic permafrost below these perennial snow patches can not be excluded.

Instead ice caves are permafrost phenomena in carbonate massifs of the Balkan Peninsula which often exist at elevations where outside MAAT is well above 0 °C (Colucci et al., 2016b) owing to karstic topography which functions as cold air trap and thus promote the survival of permanent snow and ice. Majority of caves containing permanent frozen materials, the so-called cryo caves (sensu Colucci et al., 2016b), occur at elevations above 800-1000 m in the Dinaric Alps (Bočić et al., 2014; Kern et al., 2006b; Zupan Hajna, 2016) and mountains in Macedonia (Temovski, 2016), although some cryo caves in Slovenia were reported from lower altitudes (Mihevc, 2008). In the Dinaric karst of Slovenia about 100 cryo caves (Slovenian Cave Registry, 2014) and in the northern Velebit Mountain in Croatia 150 cryo caves (Buzjak et al., 2016) were explored by speleologists. Moreover, active patterned ground likely related to seasonal frost heaving was reported from some Slovenian caves by Zupan Hajna (2007), Mihevc (2009) and Obu et al. (2018).

Nevertheless, periglacial activity is a significant geomorphological agent in the highest areas of the Balkan Peninsula. In the Rila and Pirin Mountains, periglacial processes are active above 1900-2000 m (Stoyanov and Gachev, 2012).

4.5 Anatolian Peninsula

Located between one of the world's most seismically active areas, the Anatolian Peninsula constitutes a relatively small orogenic plateau that is bounded by the Mediterranean, Aegean, Marmara and Black Seas. Despite its modest average elevations just ~1200 m and low overall exhumation, the Anatolian Peninsula is a first-order morphotectonic feature that has fundamentally impacted the geologic, geomorphic, and climatic evolution of the Eastern Mediterranean (Çiner et al., 2013). The tectonically active boundaries of the plateau delineate the Anatolian plate, which has been extruding toward the west with respect to Eurasia since the Miocene as the result of extension in the Aegean and the Arabia-Eurasia collision (Şengör and Yılmaz, 1981; Bozkurt, 2001). The Eastern Black Sea Mountains (also known as the Pontides) and the Taurus Mountain Range bound the northern and southern flanks of the plateau respectively and attain elevations more than 3000 m in places. Several paleoglacial valleys with moraines, occasional small glaciers and periglacial features are preserved in these mountains (Palgrave, 1872; Luis, 1944; Kurter, 1991; Çiner, 2003; Sarıkaya et al., 2011; Çalışkan et al., 2012; Sarıkaya and Tekeli, 2012; Bayrakdar et al., 2015; Yavaşlı et al., 2015). In the central parts, high volcanoes such as Ağrı (also known as Ararat, 5137 m), Süphan (4058 m) and Erciyes (3917 m) also contain active glaciers and periglacial features (Penther, 1905; Erinç, 1951; Messerli, 1964; Kurter, 1991; Kesici, 2005; Sarıkaya et al., 2009; Sarıkaya, 2012).

The geology of the Anatolia Peninsula is rather complicated. However, the Taurus Mountains along the Mediterranean coast of Turkey are mainly made up of platform carbonate rocks with occasional presence of ophiolitic rocks emplaced as trust sheets (Monod, 1977). On the other hand, the Eastern Black Sea Mountains along the NE Black Sea coast of Turkey contain mostly quartz bearing lithologies composed of plutonic and volcanic rocks. The volcanoes scattered in the central parts of Anatolian Peninsula are generally andesitic to rhyolitic in composition.

Late Pleistocene glacial deposits in Turkey are now well described and constitute one of the best-dated records of its kind in the world (Hughes and Woodward, 2017) (see a review of glacial landforms and TCN dating chronologies by Sarıkaya and Çiner, 2015, 2017). On the other hand, periglacial landforms are only scarcely described, often in local journals and in Turkish and lack quantitative age results. Here, we present an overview of the distribution of mountains with periglacial landforms based mainly on literature but also on field observations on periglacial processes and landforms formed since the Last Glaciation (Table 7).

Table 7

Last Glaciation
There are no age data on the periglacial landforms related to the Last Glaciation in Turkey. However, moraines in the Taurus Range, particularly in Mt. Sandıras (2295 m) (Sarıkaya et al., 2008), Mt. Akdağ (3016 m) (Sarıkaya et al., 2014), Mt. Dedegöl (2992 m) (Zahno et al., 2009; Çılğın, 2015; Köse et al., 2017), Mt. Aladağlar (3756 m) (Zreda et al., 2011), Mt. Bolkar (3524 m) (Çiner and Sarıkaya, 2017) and Mt. Geyikdağ (2877 m) (Çiner et al., 2015, 2017; Sarıkaya et al., 2017) yield TCN ages that vary between ~50 to 5 ka. To the NW of Turkey, on Başyayla Valley of Mt. Kaçkar, (3937 m) along the Eastern Black Sea Mountain Range, the MIE is reported to be around 56 ka (Reber et al., 2014). It is therefore reasonable to assume that the extent of glacier advances and retreats also controlled the intensity of periglacial activity as well as the development of periglacial landforms in these areas.

The synchronicity of the LGM in the eastern Mediterranean region is under current debate (e.g., Clark et al., 2009; Hughes and Woodward, 2008; Hughes et al., 2013). However, there seems to be a consensus on the rather synchronous LGM extent and timing (centred to around 21 ka) on the Taurus Mountains (e.g., Sarıkaya and Çiner, 2017), Eastern Black Sea Mountains (Akçar et al., 2007, 2008), Lesser Caucasus Mountains (Dede et al., 2017) and the individual mountains in the central Anatolia such as Mt. Uludağ (2543 m) (Akçar et al., 2014, 2017; Zahno et al., 2010) and Mt. Erciyes (Sarıkaya et al., 2009). It is therefore very likely that during the Last Glaciation and at LGM, periglacial processes below LGM snowline, approximately between 2400-2600 m in the Taurus Mountains and 2300-2500 m in the Eastern Black Sea Mountains (Erinç, 1952; Messerli, 1967; Çiner, 2004), were active and permafrost mainly controlled the development of some relict rock glaciers as well as patterned ground features (Figure 11).

Figure 11

Deglaciation
Following the LGM, at around 21 ka a rapid deglaciation is documented throughout the Anatolia Peninsula (e.g., Sarıkaya and Çiner, 2015). As in other periglacial alpine environments (Barsch, 1992; Knight and Harrison, 2008), mountains in Anatolia also experienced increasing rockfall and mass movements, leading to rock glacier instability, formation of outwash plains and sediment release to the rivers. This deglaciation was interrupted by two cold intervals; OD, which was rather long compared to the relatively shorter YD, where glaciers regained momentum and descended down valleys (Çiner et al., 2015; Sarıkaya and Çiner, 2015). These two cold stages were followed by rock glacier development, especially in the Eastern Black Sea Mountains (e.g., Akçar et al., 2007) and in Mt. Mercan (also known as Mt. Munzur; 3463 m) in central Anatolia (Bilgin, 1972). A relict rock glacier terminus that lies around 2100 m was recently TCN dated to

15.7 ± 1.3 ka in the Mt. Karçal (3932 m) of the Lesser Caucasus Mountains (Dede et al., 2017) (Figure 12).

Figure 12

Holocene

The Holocene record of periglacial conditions is not well known in Turkey. Although the existence of glaciers in southeastern Taurus Mountains (e.g., Mt Cilo, 4135 m) (İzbıkak, 1951; Erinç, 1953) and high volcanoes such as Mt. Ararat are known (Sarıkaya, 2012), only few glacial deposits were TCN dated. For instance, on the Erciyes Volcano in central Anatolia, reported Early Holocene (9.3 ± 0.5 ka) and Late Holocene (3.8 ± 0.4 ka) glacial advances suggesting also periglacial conditions surrounding this mountain (Sarıkaya et al., 2009). On the other hand, rock glaciers probably were extensively developed during the Holocene especially in the northern parts of the Anatolia Peninsula. However, the development of rock glaciers is rather restricted in the Taurus Range probably because of the well-developed karst that restrained surface water flow. Nevertheless, rock glaciers were previously described in Mt. Geyikdağ in the central Taurus Mountains (Arpat and Özgül, 1972; Çiner et al., 1999) and recently Çiner et al. (2017) TCN dated a rock glacier to < 6 ka (Fig. 2c). On the other hand, a study carried out on the southern slopes of Mt. Ararat (Avcı, 2007) claimed that during the glacial/periglacial climates of the past, mass movements gave rise to debris on the slopes and blocky colluvium in the valley floor.

LIA

With the exception of Mt. Ararat ice cap that covers around 5.6 km^2 (Sarıkaya, 2012) and Mt. Cilo Uludoruk glacier (<3 km long), only few small glaciers are present today in Turkey, where two third are located in the southeastern Taurus Mountains (Kurter, 1991; Çiner, 2004; Akçar and Schlüchter, 2005; Sarıkaya and Tekeli, 2014). However, glaciers were more numerous and much larger during the LIA in Turkey and in the Mediterranean mountains in general (Hughes, 2014, 2017).

For instance, the highest moraines in Mt. Uludağ were attributed to LIA (Erinç, 1952). More recently Zahno et al. (2010) TCN dated a boulder from the innermost moraine (sample TRU-12 in Zahno et al., 2010) with negligible cosmogenic nuclide concentration supporting a probable LIA advance in Mt Uludağ. Birman (1968) also tentatively proposed a recent age for the formation of this moraine.

In Mt. Aladağlar in the central Taurus Mountains a small glacier covered by rock debris (Altın, 2006; Gürgen et al., 2010) is probably a remnant of LIA glacier development. Even though at least 12 glaciers exist in the Eastern Black Sea Mountains, moraines associated with a LIA advance are reported to be absent in Kavron and Verçenik valleys of Mt. Kaçkar probably because of dry and cold climatic conditions (Doğu et al., 1993; Akçar et al., 2007, 2008; Bayrakdar and Özdemir, 2010; Reber et al., 2014). We can therefore presume that periglacial conditions surrounding these glacial environments were more extensive during LIA compared to today's conditions in these areas with probable development of rock glaciers and protalus lobes.

Present-day

Although no active glaciers exist in Mt. Uludağ in western Turkey, it is probably one of the best-described mountains concerning periglacial landforms. Studies carried out by Erinç (1949, 1957) identified two distinct periglacial levels on Mt. Uludağ. The first one lies between 1900-2300 m and is characterised by garland soils that developed in slopes ranging between 2 and 40°. The second level lies above 2300 m and is mainly represented by stone accumulations, 40-50 cm in diameter, stone stripes and stone circles preserved over quasi-flat (0-10°) surfaces (Figure 12) (Öztürk, 2012). Additionally, Türkeş and Öztürk (2008, 2011) indicated that while several alpine plant species are effective in the formation of garlands, only two types of *Festuca* sp. are effective in the formation of circles. Although much less developed compared to Mt. Uludağ, Bilgin (1960) also reported garland soils at the summit and NE facing slopes of Mt. Kazdağ (~1700 m), in the westernmost part of the Anatolia Peninsula, near the Aegean Sea. This mountain composed of granite and schist was too low to be glaciated during the Last Glaciation but relict block streams were encountered at 1350 m facing northeast suggesting past permafrost conditions (Bilgin, 1960).

Mt. Ilgaz in central Anatolia also shows some periglacial features (Erinç et al., 1961). While large features such as block flows and cyroplanation surfaces are classified as inactive, smaller stone circles (50 cm in diameter) and garlands are thought to be active. Garlands and polygonal soils above 2500 m near Yedigöller Lake on the north-facing slopes of Mt. Esence (also known as Mt. Keşiş; 3549 m) are present (Akkan and Tuncel, 1993). Stone rings at 2650 m on Mt. Mescid (3239 m) are also reported (Atalay, 1983).

On the Eastern Black Sea Range at Mt. Karagöl (3107 m), stone circles ranging from 25 cm to 1 m in diameter and ovoid depressions 1 to 1.5 m in diameter and few decimetres in depth are reported at ca. 1800 and 1900 m, respectively (de Planhol and Bilgin, 1964). Turoğlu (2009) also reported recent solifluction, frost creep and mass movements such as rock falls, talus, talus creeps, rock avalanches and rock flows on the same mountain. In the nearby Mt. Karadağ (3331 m) periglacial features are also known to exist (Bilgin, 1969; Gürgen, 2001). On Elevit and Hacıvanak glacial valleys of Mt. Göller (3328 m) active garlands, stone circles and solifluction terraces related to seasonal frost conditions are best observed on the south facing slopes (Çiçek et al., 2006).

Active rock glaciers are mostly concentrated on the northeastern Turkey and central Anatolia. In a recent study, Gorbunov (2012) detected several active, inactive and fossil rock glaciers in Turkey using satellite imagery. Among around 600 rock glaciers located mainly between 2800-3400 m, of which about 200 are reported to be active, with some reaching 1200-1300 m in length. In some places, fossil rock glaciers descend downvalleys to 2200-2300 m. For instance, in the upper parts of the northern tributary of the Başyayla Valley in the Eastern Black Sea Mountains and in the Mt. Karçal in the Lesser Caucasus Mountains several active rock glaciers were reported (Çalışkan, 2016; Gürgen and Yeşilyurt, 2012; Reber et al., 2014; Dede et al., 2015). In Kavron Valley intensive rock glacier activity is responsible for the destruction of LIA moraines (Akçar and Schlüchter, 2005). In the so-called ski valley of Erciyes Volcano a rock glacier is also described as being active (Sarıkaya et al., 2003; Ünal and Sarıkaya, 2013) (Figure 12b). On the other hand, according to Yeşilyurt and Doğan (2010) several debris-covered glaciers were mistakenly interpreted as rock glaciers on Mt. Munzur.

Mt. Dedegöl is the only site in western Taurus Mountains where rock glaciers are developed (Dellanoy and Maire, 1983). Although some are not active, others situated between 2500 and 2800 m are active with calcareous blocks of 0.5 to 1.5 m in diameter. In Mt. Aladağlar in the central Taurus Mountains, speleologists discovered a karstic cave at ~3400 m, containing 120 m thick ice along a shaft that was probably inherited from LIA (Bayarı et al., 2003; Klimchouk et al., 2006) (Figure 12). This is the only known ice occurrence within a cave system in Turkey. On the Yedigöller plateau of Mt. Aladağlar where the paleo-ice cap is known to have melted at the onset of the Holocene (Zreda et al., 2011), stone stripes are also reported at 3200 m (Figure 12). Rock glaciers also developed in front of the rapidly melting glaciers of Mt. İhtiyar Şahap (also known as Mt. Kavuşşahap; 3650 m) on the southeastern Taurus Mountains (Doğu, 2009; Yeşilyurt et al., 2018).

4.6 Northern Africa

Northern Africa mountainous regions are dominated by the Atlas, a series of ranges and plateaus, running eastwards from the Atlantic at ca. 31º N in coastal Morocco to 35 ºN in the Mediterranean in northern Tunisia, across almost 2000 km. In Morocco, the Atlas is divided in the Middle Atlas (Djebel Bou Naceur, 3340 m), the High Atlas (Djebel Toubkal, 4167 m) and the Anti-Atlas (Djebel Sirwa, 3304 m), not analysed in this paper. The three major massifs in the High Atlas are, from west to east, the Djebel Toubkal massif (4167 m), the Irhil M'Goun massif (4071 m) and the Djebel Ayachi (3751 m). Towards the east, the High Atlas extends to the Saharian Atlas, culminating in the Aurès Massif in Algeria at 35ºN (2324 m). North of the Atlas and close to the Mediterranean coast, lie the Rif (3445 m) and the Algerian Tell (2308 m) divided from the former by the Algerian high plateaus (1729 m). Further south, already in the interior Sahara Desert, lie two other large massifs, with a drier climate and with evidence of relict periglacial phenomena: the Hoggar (2981 m) in southern Algeria at 23ºN, and the Tibesti (3445 m) in northern Chad at 21ºN.

The major geological structures of the mountains in North Africa are fold-thrust belts formed by the collision between the African and the European plates (Gómez et al., 1977; Dewey et. al., 1989) along the plate boundary. The materials in the Rif-Tell Atlas are allochthonous, with the presence of flysch sequences and limestones covering African shield metamorphics and granitoids. The High Atlas – Saharan Atlas is an intracontinental fold-thrust belt in the foreland of the Rif (Arboleya et al., 2004), with autochthonous rocks. Mesozoic limestones prevail in the Middle Atlas, also with Cenozoic basaltic lava flows infilling pre-existing valleys (Arboleya et al., 2004). The High Atlas has three major morphostructures, with the western part showing limestone plateaus, the central zone showing Paleozoic magmatics from the African shield (granites, rhyolites, andesites and trachytes) and the eastern zone, a folded structure of carbonates (Joly, 1962; Hughes et al., 2004). The Saharan Atlas is composed by folded structures affecting mostly Cenozoic limestones.

The Hoggar massif is located in a cratonic setting, showing in its western part, middle-Proterozoic thick meta-sedimentary units and alkaline-peralkaline intrusives, magmatic complexes of basic to ultrabasic rocks and volcanoclastic deposits, andesites to dacites and calc-alkaline batholiths. The central Hoggar is mainly composed of gneisses and schist belts, while the eastern Hoggar shows mainly gneisses and granites (Bertrand and Kaby, 1978). The Tibesti massif is composed of Precambrian crystalline rocks, being formed by a core of intrusive and metamorphic rocks, which is bounded by Paleozoic and sedimentary sequences. These units are partially capped by Tertiary volcanics (Ghuma and Rogers, 1978, Permentier and Oppenheimer, 2007).

The mountains of North Africa mark strong climatic gradients between the Atlantic in the west, the Mediterranean in the north, and the Sahara influence in the south. This impacts essentially the precipitation amounts, varying from over 2,000 mm in the Rif mountains to145 mm in the Hoggar and 11 mm in the Tibesti, and are characterized by regimes, from the Mediterranean type, in the north, to semi-arid and arid regimes as one moves towards the Sahara. Air temperatures show both the effects of continentality and latitude, with striking impacts on the estimated altitudes of the 0 ºC MAAT, which range from slightly above 3000 m in the Rif to about 4900 m in the Hoggar, much higher than the maximum elevations in these ranges (Table 2).

Periglacial features in North African mountains have been described by several authors, but most through a classical naturalistic approach, lacking quantification, sedimentological studies and absolute dating of relict features (Table 8). Most features relate to evidence of frost shattering, talus slopes, stratified slope deposits and more sporadically, to rock glaciers, although lacking detailed observations. At least one case of the latter, in the High Atlas, has been recently reinterpreted as a catastrophic rock slope failure deposit, rather than as a rock glacier (Hughes et al., 2014), which calls for a need for encompassing detailed studies and reinterpretation. Present-day periglacial activity has been vaguely described by the presence of solifluction landforms and ground frost features at high altitude localities, but again the observations lack modern objective assessments and monitoring data are lacking almost everywhere. A recent study by Vieira et al. (2017) based on geomorphological evidence and ground surface temperature data indicates the possible presence of permafrost in the upper reaches of the High Atlas. Both relict and present-day periglacial features reflect the climatic gradient, with the evidence showing an increasing moisture content from SE to NW and a decrease of temperatures.

Table 8

Last Glaciation

No absolute age records exist from periglacial deposits or landforms in North Africa and the only chronological framework for the cold event derives from glacial evidence, which have been especially well-studied in the High Atlas (De Martonne, 1924; Dresch, 1949; Wiche, 1953; Messerli, 1967; Messerli and Winiger, 1992; Chardon and Riser, 1981; Hughes et al., 2004, 2011, 2018; Hannah et al., 2017). In the Djebel Toubkal region, Hughes et al. (2018) have identified three phases of glacier advance or sustained stabilisation, with ages of ca. 50 ka, 22 ka and 12 ka, pointing to a pre-LGM maximum and to the presence of glaciers during both the LGM and the YD. These relate to several valley glaciers that descended in some areas down to 1900 m, with typical glacier lengths of 2 to 9 km. Other ranges showing glacier evidence are the Rif (Mensching, 1960) and the Middle Atlas (Dresch and Raynal, 1953; Raynal et al., 1956; Awad, 1963) in Morocco, and the Tell Atlas (Barbier and Cailleux, 1950) and the Aurès mountains (Ballais, 1981) in Algeria, but with no known absolute age records.

Relict periglacial deposits in northern Africa mountains have been attributed by the different authors mostly to the Last Glaciation (Figure 13). The most frequent reference to periglacial features are probably slope deposits, which vary from head-type deposits to stratified slope deposits and grèzes in some limestone areas. They were identified in the Algerian Tell above 1400 m and in the Aurés Massif above 1800 m (Tihay, 1973), and are widespread in the High Atlas (Chardon and Riser, 1981). We have observed a good outcrop at the Middle Atlas southeast of the Djebel Bou Iblane at 2350 m. Other references to slope processes relate essentially to a diffuse concept of solifluction, which has to be accounted for with great care due to the difficulty relating to the identification of processes, but also to the variability of the interpretation of the concept. As such, relict solifluction features have been identified in the High and Middle Atlas (Awad, 1963) and Messerli (1973) associates frost action in the Hoggar and Tibesti to the presence of slope deposits and filling of hollows, which are present even below 2000 m. Nivation forms (presumably hollows), as a set of geomorphic processes, have been identified by Messerli (1973) in the Hoggar (above 2400 m), in the Tibesti (above 3000 m), by Tihay (1973) in the Aurés above 1800 m and in the Algerian Tell above 1500 m, by Marre and Quinif (1981) in the Algerian High Plateaux above 1500 m, and in the Rif by Mensching (1960). Rock glaciers, which are the single landform identified which is an indicator of the presence of permafrost, have been reported for the Algerian Tell (possible) above 1100 m by Tihay (1973) (an altitude that seems excessively low), for the Middle Atlas between 2100 and 2500 m (Awad, 1963) and for the Rif by Mensching (1960). Nivation forms would have been widespread marginal to glaciated areas of the High Atlas during the Last Glaciation. Patterned ground can be observed on plateau surfaces in the High Atlas, especially between 3000 and 3600 m over the extensive high plateaus of Iouzagner (3502 m) and Tazaghart (3980 m). Hannah et al. (2017) described large polygonal and linear patterned ground (10-20 m wide and up to 1000 m long) as well as extensive covering of thick regolith, blockfields, blockstreams and smaller-scale stone stripes on these high plateaus. The plateaus were covered in ice during the most extensive glacial phase of the Last Glaciation (before ~50 ka based on dating in Hughes et al. 2018) and the periglacial features formed afterwards and are probably still active today given the high altitude of the plateau.

Figure 13

Deglaciation

Few observations exist for the deglaciation phase. However, the rock glaciers reported for the M'Goun Massif by Wiche (1953) which are located inside the valleys well-within the glacier limit can be included in this phase. These features are probably derived from frontal moraines in the cirques, as suggested by Hughes et al. (2006). It is also possible that the rock glaciers described by Awad (1963) for the Middle Atlas, close to the Djebel Bou Naceur, as well as some of the High Atlas rock glaciers described by Dresch (1941) and Wiche (1953) also correspond to this phase.

Present-day

Observations on present-day periglacial processes and landforms are scarce and lack quantitative data. The lack of a common methodology by the different authors limits a regional analysis. Messerli (1973) shows that the Tibesti and Hoggar lack present-day frost activity, a fact that they attribute more to the dryness than to the lack of sub-zero temperatures. At the Aurés Massif, Tihay (1973) indicates that frost action occurs above 2300 m, while Ballais (1981) mentions solifluction between 1300 and 1800 m. For the Algerian Tell, Tihay (1973) lowers these limits, putting frost action above 1500 m. In the Middle Atlas, Dresch and Raynal (1952) and Raynal et al. (1956) indicate active frost shattering. Above 2700 m a marginal periglacial zone starts to be evident in the eastern part of the Bou Naceur massif, with vegetation crescents and very shallow solifluction.

The most consistent reports on present-day periglacial activity in North Africa are from the Djebel Toubkal massif in the High Atlas, the highest peak in north Africa. Periglacial features are widespread (Hughes et al., 2011) and were described for the Central High Atlas by Couvreur (1966), who reported active solifluction above 2200 m (Figure 14). For the western High Atlas, Chardon and Riser (1981) indicate the limit of frost activity to be around 2500 m and considered that frost action dominates the morphogenesis above 3000 m. Chardon and Riser (1981), seemingly were the first to interpret a lobate feature in the Irhzer Ikhibi south at 3800 m as an active rock glacier. Vieira et al. (2017) installed a series of ground surface temperature and air temperature dataloggers during one full year, from 3210 to 4160 m altitude. The authors analysed the ground temperature regimes and snow cover and identified a hot season from late-May to late-September and a long cold season from mid-October to mid-April. Freeze-thaw regimes were analysed and the most important finding was the possible presence of permafrost at a location close to the rock glacier identified by Chardon and Riser (1981). This interpretation was based on the very low temperatures (c. -5.8 ºC) measured at the ground surface beneath a stable snow pack that lasted from mid-December until late-March. Other lobate rock debris features are found at similar altitudes in other cirques nearby and it is possible that these are also associated with sporadic patches of permafrost (Hughes, 2018). Permanent snowfields were present in the High Atlas in the mid-20th century and were probably widespread in the LIA (Hughes, 2018).

Figure 14

4.7 Mediterranean islands

The Mediterranean Sea encompasses several archipelagos and some major islands, some of which exceed 10,000 km^2 (Sicily, Sardinia, Corsica, Crete). Some of the islands are very mountainous, with elevations exceeding 2000 m in Sicily (Etna, 3350 m), Corsica (Monte Cinto, 2706 m), Crete (Lefka Ori, 2452 m). Lithologies are highly variable, with abundant limestones and granites and also volcanic rocks in the case of the highest elevation within the limits of the Mediterranean Sea, the Etna. While past glacial activity in Mediterranean islands has received some attention (e.g. Hughes et al., 2006a, Hughes and Woodward, 2016), periglacial processes have been less examined.

Last Glaciation

Glaciation has affected the mountains of Corsica, and the active volcano of Mt. Etna in Sicily, and the mountains of Crete. In the highest mountains in Corsica, Kuhlemann et al. (2008) dated *roches moutonnées* at 18 ka BP. In Crete, Bathrellos et al. (2014) found 17 glacial cirques between 1870 and 2360 m distributed around the two main mountains groups: Mt. Lefka Ori (2453 m) and Mt. Idi (2456 m). However, the ages of the Cretan glaciations remain untested and could potentially be older than the Late Pleistocene as in parts of mainland Greece (cf. Hughes et al., 2007).

Past periglacial activity has been documented in several Mediterranean islands associated in most of the cases to seasonal frost regime. In Corsica, despite the quite extensive glaciation during the LGM periglacial landforms are not common if we exclude granite weathering features like tafoni, weathering pits or grooves that are extensively widespread in the island and that occur largely also in modern cryotic conditions in Antarctica (i.e Guglielmin et al., 2005; Strini et al., 2008). Kuhlemann et al. (2009) recorded blockfields and tors above 2200 m and describe them as possible periglacial landforms, even reporting less common similar features at lower altitudes.

Landforms associated to intense periglacial conditions are also observed in Sardinia as indicated by Ginesu (1990) who described several blockfields and block streams composed by basalts in the centre of the island (Pranu Mannu plateau) at an altitude of about 600 m. Ginesu and Sias (2006) suggested a periglacial origin also for other block accumulations on different lithologies in the higher reliefs of Sardinia (Gennargentu, Limbara, Perdasdefogu). More recently, Ginesu et al. (2014) interpreted some block deposits found on Asinara island as block streams. In this case, block streams are composed by granites subrounded blocks and located close to the coast and even below the present sea level (4 m depth) in Cala Arena. Block streams and block slopes close to Tyrrenian coast have been described in the Pisani Mountains between 400 and 800 m on quarzites by Casarosa and Pappalardo (2006).

Poser (1957) found patterned ground and solifluction lobes above 1800 m in Crete, suggesting that the lower limit of periglacial activity in this island was represented by boulder pavements extending down to elevations of 800 m. Cemented stratified scree of unknown age are present on the southern slope of Mt. Idi (Guglielmin, unpublished), probably related to the coldest phases of the Last Glaciation.

In the Balearic Islands there is evidence of periglacial activity during the Last Glaciation, with scree deposits located at elevations above 1200 m in the limestone Tramuntana massif (Rosselló, 1977) though no evidence of permafrost conditions was detected at this altitude.

Deglaciation

The temperature increase recorded following the LGM conditioned the deglaciation of Mediterranean mountains, with only the possible existence of minor glaciers during the OD and YD in the highest mountains in Corsica. No evidence of permafrost conditions has been reported for this stage, though periglacial processes driven by seasonal frost may have existed during the coldest stages in the highest mountains.

Holocene

It is very unlikely that climate conditions prevailing during the Holocene allowed the existence of permafrost even in the high lands of the major islands.

LIA and present-day

LIA cold-climate conditions probably reactivated periglacial processes in the highest mountains in Sardinia and Sicily. In fact, evidence of permanent ice deposits in lava tubes has been found at 2043 m on the north flank of Mt Etna volcano (*Grotta del Gelo,* Cave of Frost), in Sicily. It is probably the southernmost European ice cave (Marino, 1992; Hughes and Woodward, 2009; Scoto et al., 2016). The lava tube formed during the historic long-lasting eruption of 1614-24 and despite its geological setting and latitude, after about twenty years from the last phases of the eruption (coinciding with the Minimum Maunder, 1645-1715), subterranean freezing inside the cave started to take place. Beside the development of seasonal ice formations (seasonal lake ice, ice stalactites, stalagmites and columns generally located close to the entrance), perennial ground ice is present in the deepest zone. The ice extension reaches about 240 m^2, and the volume is estimated at about 220-260 m^3 (Scoto et al., 2016). On Mt. Etna, at elevations above 2900 m, nivo-aeolian deposits were found with frozen ground beneath that can persist for more than 2 years, and therefore can be considered as current permafrost (Guglielmin, unpublished).

5. Discussion

Climate variability during the Quaternary has conditioned the spatial distribution of glacial and periglacial processes in the Mediterranean region, and therefore the area under permafrost conditions (Figure 15). Glacial stages favoured a substantial expansion of glaciers in the currently still glaciated mountain ranges and the formation of glaciers in presently deglaciated mountain environments (Woodward, 2009). The glaciation in the Mediterranean mountains has been long debated and several reviews have focused on the calendar and geography of the maximum glacial expansion in the different mountain regions (i.e. Allen et al., 1999; Hughes et al., 2006a, 2008, 2013; Hughes and Woodward, 2016). The temperature increase recorded during interglacial phases - such as the Holocene - conditioned the complete disappearance or substantial retreat of glaciers and the migration of permafrost and periglacial processes to higher elevations (i.e. Oliva et al., 2016b).

Figure 15

5.1 Last Glaciation

The calendar of the maximum glacial expansion of the Last Glaciation shows a diachronous pattern among the massifs, which must be also framed with the dating method used in each study (Hughes et al., 2013). Whereas radiocarbon, U-series and OSL indicate an early glacial advance that occurred several thousand to tens of thousands of years earlier than the global LGM in the Sierra Nevada, Cantabrian Mountains, Central Pyrenees, Italian Apennines and Pindus Mountains, cosmogenic exposure ages suggest a local MIE in the Iberian Central Range, Maritime Alps and Anatolia mountains (almost) synchronous to the LGM (Hughes et al., 2008).

The FIS extended over northern Europe until latitude 50ºN in its southern fringe (Hughes et al., 2015; Stroeven et al., 2016). Therefore, the Mediterranean region was not extensively glaciated by comparison, with glaciers concentrated in mountain environments in the form of small ice caps and mountain glaciers filling the valleys. The ELA decrease in Mediterranean mountains during the LGM compared to present-day was about 800 and 1200 m, which is translated in a minimum temperature difference of ca. 6-10 ºC (Allen et al., 2008; Kuhlemann et al., 2008). The glaciated environments in the western and central Mediterranean region was associated with ELAs above 1500-2000 m in most of mountain ranges (Hughes et al., 2006a, 2008), slightly increasing towards the easternmost fringe in mountains of Turkish and Lebanon where the ELA was located around 2500-3000 m (Messerli, 1967).

The southern ice-free land surrounding the large FIS was underlain by permafrost, which extended across the lowlands in much of central Europe as well as the northern fringe of the Mediterranean region, in areas such as central France and the northern Balkan region (Brown, 2001; Vandenberghe et al., 2014). Dry and cold climate conditions prevailing in these permafrost areas during the LGM promoted intense aeolian activity, with the deposition of thick loess deposits in non-glaciated environments from central-northern Europe (Antoine et al., 2009) and northern Italy (Cremaschi et al., 2015). In the Mediterranean region, the presence of ice wedges and cryoturbations features in some basins, such as in the Pannonian basin (van Vliet- Lanoe et al., 2004; Ruszkiczay-Rüdiger and Kern, 2015) or in some basins in the Spanish Meseta (Badorrey et al., 1970; Asensio-Amor and González-Martín, 1974; Serrano et al., 2010a) has been related to deep seasonal frost conditions or even isolated patches of sporadic permafrost. Despite few data in areas surrounding the valley glaciers in the southern Alps, environments at the foot of the northern Apennines and even at the Po plain should have been affected by permafrost conditions as suggested by loess deposits and the existence of some cryoturbation features (Cremaschi et al., 2015). In the Mediterranean mountains, immediately below the glaciated environments was the periglacial belt affected by permafrost regime at high elevations and seasonal frost at the foot of the mountains and high-altitude plateaus (e.g. 600-1200 m).

Permafrost was also widespread in non-glaciated environments above the snowline as revealed by the present-day distribution of inactive permafrost-derived features. Rock glaciers formed during the Last Glaciation have been described in several Mediterranean mountains at a wide range of altitudes (e.g. Figure 3 and Table 5). Block streams, some km long and hundreds of meters wide, also developed during this stage at elevations between 700 and 1800 m. In relatively flat summit areas where wind action did not favour snow and ice accumulation, meter-sized stone circles developed in several massifs except for the Italian Peninsula where there are no evidences of large stone circles in the LGM ice-free areas. Similar patterned ground features are observed today in present-day polar environments where permafrost conditions are widespread with mean annual temperatures below -6 ºC (French, 2007). It is therefore feasible to relate the formation of these currently inactive features to permafrost occurrence. Concurrently, very intense periglacial conditions favoured the development of large blockfields and tors, which are remnants of original surfaces in the highest lands. The thermal regime of the ground in the nunataks standing out the glaciated slopes must have been also characterized by permanent frozen conditions well below 0 ºC.

5.2 Deglaciation

During the most important phases of the last deglaciation, the evolution of the glaciers of the different Mediterranean mountains shows a much more homogenous pattern than during the LGM. The latest studies indicate interruptions during the deglaciation process and even some important glacial advances, in full synchrony within the Mediterranean region and with the mountainous systems of central Europe, as well as with the FIS. The onset of a massive retreat of the FIS is dated at 19 ka and accelerated at 18 ka (Marks, 2015; Stroeven et al., 2015; Toucanne et al., 2015; A.L. Hughes et al., 2016; Petrini et al., 2018), in parallel to most European mountains. For example, some of the largest glaciers in the Alps had lost ca. 80% of their mass at 18 ka (Ivy-Ochs et al., 2004, Ivy-Ochs, 2015) and glaciers could almost have disappeared completely in the Pyrenees at that time, as probably occurred in many other Mediterranean mountains (Palacios et al., 2017a).

The first glacial advance within the deglaciation took place during the OD, when advances have been detected in numerous sectors of the FIS (Marks, 2015; Stroeven et al., 2015; Toucanne et al., 2015; A.L. Hughes et al., 2016). In the Alps, glaciers expanded along the valley bottoms with their fronts reaching elevations only 400 m above the LGM moraines (Kerschner and Ivy-Ochs, 2008; Ivy-Ochs et al., 2009; Ivy-Ochs, 2015). Similar glacial advances occurred during this period in the Tatra Mountains (Makos, 2015) as well as Mediterranean glaciers that also experienced important advances, as in the Iberian Peninsula (Palacios et al., 2017a), Southern Alps (Federici et al., 2012), Apennines (Giraudi, 2015), Balkan Peninsula (Kuhlemann et al., 2013) and Anatolian Peninsula (Sarıkaya et al., 2008, Akçar et al., 2014). A significant drop of the North Atlantic temperature induced a significant reduction of the meridional overturning circulation (Bard et al., 2000; McManus et al., 2004), with extreme seasonality of cold winters and mild summers (Denton et al., 2005; Williams et al., 2012) favouring glacial expansion during this period. Similar conditions have also been detected in southern Europe and in the Mediterranean region (Fletcher et al., 2010). For example, glaciological models indicate that during the OD the MAAT in the Alps was 10 ºC lower than present-day, and precipitation was one-third less (Ivy-Ochs, 2015). Undoubtedly, these conditions would have favoured a substantial expansion of permafrost conditions in mountain environments. Many fossil periglacial landforms existing in Mediterranean regions, such as patterned ground, protalus lobes or rock glaciers, could have originated at this time. But these climatic conditions changed drastically during the BO, when the environmental conditions shifted abruptly and become very similar to present in the Mediterranean region (Fletcher et al., 2010). Consequently, there was a massive glacial retreat in Europe, including the FIS and British-Irish Ice Sheet (Marks, 2015; Stroeven et al., 2015; Toucanne et al., 2015; Hughes et al., 2016), central European mountains (Ivy-Ochs, 2015; Makos, 2015) and Mediterranean mountains (Sarıkaya et al., 2008; Federici et al., 2012; Kuhlemann et al., 2013; Akçar et al., 2014; Giraudi, 2015; Palacios et al., 2017a).

The recent application of absolute dating methods, fundamentally cosmogenic, to fossil rock glaciers is allowing a better comprehension of the timing of their formation and stabilization, remaining therefore exposed to cosmogenic radiation. In many European mountains, this method is showing that numerous rock glaciers developed on polished glacial surfaces that were deglaciated just at the end of OD. In many cases, the fronts of these rock glaciers were stabilized soon after formation, although their roots remained active during thousands of years (Oliva et al., 2016b, Palacios et al., 2017a,b). This fact shows evidence that most fossil rock glaciers in different Mediterranean mountain ranges developed under paraglacial conditions (Ballantyne, 2002; Mercier, 2008; Oliva et al., 2016b). They occupied the formerly glaciated cirques, extending over very active geomorphological areas at the end of the OD. Subsequently, as temperatures increased during the OD permafrost conditions migrate to higher elevations and rock glaciers became relict. This process has been mostly described in many Iberian mountain environments (Rodríguez-Rodríguez, 2016; Fernández-Fernández, 2017; Palacios et al., 2017a,b), and should be tested in the future in other Mediterranean ranges. Noteworthy, a similar pattern has been also reported in other mountains, such as in the British Isles (Ballantyne et al., 2009) and the Alps (Hippolyte et al., 2009).

During the YD, in full agreement with the last major advance of the FIS (Greenwood et al., 2015; A.L. Hughes et al., 2016; Stroeven et al., 2015), Mediterranean glaciers advanced again for the last time during the Pleistocene, but only in the highest massifs. Glaciers rarely exceeded the limits of the cirques in the Pyrenees (García-Ruiz et al., 2016a,b) and were moderately longer in the southern Alps (Ivy-Ochs et al., 2008, 2009). This glacial advance was related to a cold and arid period during the YD in the Mediterranean region that ended abruptly at 11.8 ka cal BP (Fletcher et al., 2010). In the Alps, where glaciers had virtually disappeared during the BO, they formed again during the YD and progressed considerably (Ivy-Ochs, 2015) due to 3.5-5 ºC lower MAAT than present and precipitation was up to 30% less (Kerschner and Ivy-Ochs, 2008). A similar pattern was also detected in the Tatra Mountains (Makos, 2015).

As occurred at the end of the OD, the end of the YD favoured the rapid retreat of glaciers and, in some cases, the reactivation of paraglacial processes in the walls of the cirques and the formation of rock glaciers and protalus lobes. Again, their fronts became inactive soon after formation, while the roots could have remained active even until the warmest Holocene periods (García-Ruiz et al., 2016a,b; Oliva et al 2016b). Many generations of Mediterranean rock glaciers have been considered to form during or at the end of the YD, as reported in many Iberian mountains (Palacios et al., 2015, 2016; Andrés et al., 2018) and southern Alps (Colucci et al., 2016a), though in other high mountain Mediterranean regions this fact still needs to be confirmed.

Therefore, the cold periods interrupting the long-term deglaciation process showed both glacial and periglacial evidence in the Mediterranean mountains, mainly with the formation of permafrost-related features associated with paraglacial dynamics. The existence of these inactive landforms suggests the minimum altitude for the presence of permafrost during these stages.

5.3 Holocene

The northern Atlantic region has been subjected to significant climate shifts during the Holocene, as revealed by marine sediment records (Witak et al., 2015) and Greenland ice cores (Masson-Delmotte et al., 2005). Climate variability has been also significant across the Mediterranean region, where Holocene temperature oscillations of the order of ca. ± 2 ºC have caused significant disruptions in early civilizations (Mayewski et al., 2004). During the Holocene, temperature and moisture shifts have also affected the type and intensity of cold-climate geomorphological processes prevailing in Mediterranean mountains.

The onset of the Holocene saw an accelerated shrinking of the glaciers, which disappeared until nowadays in many massifs during the Early Holocene (Gómez-Ortiz et al., 2012a; García-Ruiz et al., 2016a,b; Palacios et al., 2016). The formerly glaciated environments became occupied by periglacial dynamics, which expanded gradually upvalleys (Oliva et al., 2016b). In many mid-altitude mountain ranges – with highest peaks around 2000 m in the central and western Mediterranean and 2500-3000 m in the eastern part of the region – the periglacial belt disappeared during the Holocene, and significantly shrunk in the highest ranges, particularly during the warmest stages, such as the Holocene Warm Period (HWP) and the Medieval Climate Anomaly (MCA). In the southern Alps, even during warmer Holocene stages some currently active rock glaciers were still active (Calderoni et al., 1998; Stenni et al., 2007; Scapozza et al., 2010); indeed, the major part of the dated rock glaciers reported an age between 2720 and 2850 cal yr BP with some older exceptions (all younger than 5900 cal yr BP; Calderoni et al., 1998; Guglielmin et al., 2001; Dramis et al, 2003). During the warmest phases, only the environments above 2500-3000 m included a periglacial belt mostly related to seasonal frost conditions, with permafrost regime limited to the highest ranges in the Maritime Alps and the highest mountains in Anatolia.

Most of the active periglacial features during the YD became gradually inactive, and only reactivated during the coldest stages of the Holocene. Permafrost-related features showed activity until the Early Holocene in many of the highest Mediterranean massifs (Kuhlemann et al., 2009; Gómez-Ortiz et al., 2012a; Palacios et al., 2016, 2017a,b). Some authors proposed the initial formation of currently active rock glaciers synchronously to the HWP, such as in the central Taurus Mountains (Çiner et al., 2017) and the Pyrenees at ca. 6 ka (Serrano et al., 2010c). Periglacial landforms related to seasonal frost conditions, such as solifluction landforms located today in the present-day periglacial belt of Sierra Nevada at elevations between 2500 and 3000 m, were inactive during the HWP until 5 ka cal BP but reactivated later during cold and wet phases, namely at 5-4, 3.6-3.4, 3-2.8, 2.5-2.3, 1.8-1.6, 0.85-0.7, 0.4-0.15 ka cal BP (Oliva et al., 2011). Therefore, 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 permafrost conditions.

5.4 Little Ice Age

The LIA has been defined as the coldest most recent period where many glaciers reached their largest volume over the last 10,000 years (Bradley and Jones, 1992), and, thus, it has been widely employed in mid-latitude mountain environments as the reference for Holocene cold stages (Grove, 2004; Oliva et al., 2018). In the Mediterranean region, colder and (generally) wetter than present-day climate conditions prevailing during the LIA also favoured the presence of larger and more numerous glaciers (Hughes, 2014). Consequently, the spatial domain of periglacial dynamics expanded down-valleys and cryogenic processes reappeared in some areas where seasonal frost activity was limited during the MCA and is no longer active at present.

During the LIA permafrost features such as rock glaciers and protalus lobes formed and/or reactivated in the highest slopes of the main massifs, above 2350 m in some areas of the Balkan region and 2560 m in the Pyrenees (Tables 6 and 2). In the southern Alps, especially in the eastern part several rock glaciers developed since the LIA around 2400-2500 m (Baroni et al., 2004; Seppi et al., 2014) while only a few active rock glaciers formed in the western Alps and central Alps.

Ice caves, namely natural caves formed in bedrock containing perennial accumulations of ice, are considered as sporadic permafrost phenomena (Holmlund et al., 2005; Luetscher et al., 2005; Hausmann and Behm, 2011; Luetscher et al., 2013). As part of the cryosphere, ice caves occurrence is closely linked to cold climates, even if they do also exist in different kind of environments, often at an altitude with an outside mean annual air temperature well-above 0°C (Holmlund et al., 2005; Stoffel et al., 2009; Obleitner and Spötl, 2012; Colucci et al., 2016b). Ice caves existing in several high karstic mountains have been examined for paleoenvironmental purposes. In most cases these ice caves are several hundreds of meters below the glaciated environments during the LIA, reaching elevations even below 1000 m in the Dinaric Alps (Kern et al., 2006b; Bočić et al., 2014). The dating of the organic remnants preserved in the ice suggests that it accumulated during the coldest centuries of the LIA (Bayarı et al., 2003; Kern et al., 2006b; Klimchouk et al., 2006; Bočić et al., 2014; Gómez-Lende, 2015; Bartolomé et al., 2015; Temovski, 2016; Zupan-Hajna, 2016; Sancho et al., 2016), and even during the MCA and HWP (Sancho et al., 2016). Interestingly, the frozen ground in contact with the ice inside the caves has been also described as permafrost (Gómez-Lende, 2015; Colucci et al., 2016b), which in many cases still persist in environments with MAAT well above 0 °C due to specific microtopographic conditions (very low solar radiation, existence of stable temperature inversion or density driven air flows and chimney effects).

In all Mediterranean massifs, long-lasting and perennial snow-patches were more extensive than today, which also enhanced nival processes that left some small well-preserved landforms such as protalus ramparts at the foot of slopes.

5.5 Present-day

In the western Mediterranean region, the temperature increase since the last cold stages of the LIA has been estimated at ca. 1 ºC (González-Trueba, 2006; González-Trueba et al., 2008; Oliva and Gómez-Ortiz, 2012) and ca. 2 ºC since the Minimum Maunder (Oliva et al., 2018). A similar increase has been inferred for other parts of the Mediterranean, such as in the Balkans (Repapis and Philandras, 1988 Xoplaki et al., 2001; Hughes, 2010). This temperature increase has led to glacial retreat/disappearance (Zumbühl et al., 2008), enhanced paraglacial activity (Cossart and Fort, 2008), shift of periglacial processes to higher elevations (Oliva et al., 2011), geoecological changes (Cannone et al., 2007, 2008; Pauli et al., 2012; García-Ruiz et al., 2015; Camarero et al., 2016), as well as degradation of alpine permafrost in mid-latitude high mountain environments, e.g. the Alps (Harris et al., 2003; Gruber et al., 2004; Lugon et al., 2004; Zenklusen-Mutter et al., 2010).

A wide range of approaches have been used to infer the current altitudinal limit of permafrost conditions, such as geomorphological techniques, geophysical surveying and monitoring activities (BTS measurements, boreholes, terrain deformation). Today, permafrost conditions generally increase in elevation towards the eastern part of the region and from north to south. In the western and central Mediterranean, permanently frozen ground is rarely found below 2500 m, discontinuous permafrost is generally detected between 2500 and 2800/3000 m and continuous permafrost is distributed in ice-free environments above this level. An alpine permafrost belt is detected above 2630 m in northern aspects and 2800 m in southern ones in the Pyrenees (Serrano et al., 1999, 2001, 2002, 2006, 2009, 2011a; González-García et al., 2014), above 2400 m in the Southern Alps (Bodin et al., 2009), above ~2350 m on Rila Mountain and ~2700 m on Mount Olympus (Dobinski, 2005) and above 2800-3400 m in NE Turkey and central Anatolia (Gorbunov, 2012). No permafrost belt is found in the highest mountains in southern Europe (i.e. Sierra Nevada) and northern Africa (i.e. Atlas) where permanent frozen conditions are only found in form of isolated patches at the highest elevations at 3000-3100 m (Oliva et al., 2016a) and 3800 m (Vieira et al., 2017), respectively. Certain climate conditions (i.e. reduced snow cover) can favour the presence of permafrost patches at relatively low elevations in the Central Apennines on La Majella and M. Velino massifs (ca. 2400 m) or by lithological conditions (i.e. volcanic sediments), as detected in the highest active European volcano (Mt. Etna) at elevations above 2900 m.

Scientific papers and international reports coincide in anticipating that the Mediterranean basin will be one of the areas on Earth where annual and seasonal climate variations will be more pronounced, with significant warming and reduced rainfall (e.g. IPCC, 2013). These climate conditions would lead to a migration of permafrost conditions to upper elevations, even disappearing in those areas where isolated patches persist but undergo a rapid degradation process (e.g. Atlas, Sierra Nevada Cantabrian Mountains), possibly triggering large rock avalanches or other mass wasting processes with important socio-economic impacts.

6. Conclusions

The present and past distribution of cold-climate geomorphological processes in the Mediterranean region is conditioned by the topography and wide spectrum of microclimatic conditions prevailing in the region. Both glacial and periglacial processes (including permafrost conditions) since the Last Glaciation have been almost exclusively restricted to mountain environments.

Since the Last Glaciation there has been a long-term decrease in the area occupied by permafrost in the Mediterranean mountains. The existence of abundant inactive permafrost-derived landforms (rock glaciers, block streams, patterned ground features, ice wedges) formed during the last Pleistocene glacial cycle suggests that ice-free slopes in mid-altitude mountain environments and high summit plateaus together with the lowlands of the northernmost fringe of the Mediterranean region could have been underlain by permafrost conditions to a greater or lesser extent during that stage. The thermal increase following the LGM meant that, after that time, periglacial conditions were limited to mountain areas. The gradual upward shift of the snowline in all Mediterranean massifs conditioned a massive deglaciation that was only interrupted by brief periods of glacial readvance. Paraglacial dynamics during the deglaciation process favoured the development of most of the rock glaciers distributed in the majority of the massifs, almost all of which are inactive under present-day climate conditions. Therefore, their formation is primarily associated with adjustment of the cirque walls to a new morphodynamic setting and not strictly related to a climatic origin. Warmer temperatures during the onset of the Holocene saw a shrinking of periglacial activity to the highest elevations. Many rock glaciers and protalus lobes became inactive during the Early Holocene as temperatures rose. Permafrost disappeared from most of the massifs and only reappeared in some mountains during the coldest stages of the Holocene, such as the LIA. Since then, post-LIA warming led to the spatial confinement of continuous/discontinuous permafrost conditions in the highest mountain areas, such as some areas in the Pyrenees, Southern Alps, Apennines and Anatolian mountains, or as isolated patches in north-facing cirques that were glaciated during the LIA in the Sierra Nevada, Atlas Mountains and the Balkan region.

In contrast to neighbouring mountain environments where there has been a richer analysis of permafrost history (i.e. Alps), permafrost research in the Mediterranean basin still has some gaps:

- The improvement of the glacial chronology that has taken place over the last decade in many Mediterranean massifs has not been paralleled by a strengthening of the chronology of periglacial activity, and therefore of permafrost evolution in the Mediterranean region.
- Data about periglacial dynamics and spatial domains of permafrost are substantial for certain periods (i.e. the last deglaciation), though current knowledge is still poor for others (i.e. Holocene).

- Knowledge of permafrost evolution in some mountain environments is still limited, particularly in North Africa ranges, Apennines and Mediterranean islands.
- Future studies should focus on precise descriptions of landforms and quantitative age proxy data, especially on rock glaciers as indicators of present and past permafrost conditions.

Finally, permafrost scientists in the Mediterranean region should focus on these gaps in order to better understand the spatio-temporal evolution of permafrost conditions in the region. A better characterization of the evolution of permafrost - a key component of the cryosphere in landscape dynamics in mid-latitude high mountain ranges – may be also helpful to anticipate the future geoecological response towards the changing climate scenarios forecasted in these highly sensitive mountain ecosystems.

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Figure captions

Figure 1. Location of the different study areas examined in this paper within the Mediterranean region (top), together with mean annual temperatures and altitudes where the 0 ºC isotherm is located (middle), and annual precipitations with values for each of the mountain ranges considered in this study (bottom). Annual precipitation and mean annual temperatures for the period 1960-1990 was obtained from Hijmans et al. (2005), spatial resolution 30 arc-seconds, modified after WorldClim 1.4 (www.worldclim.org).

Figure 2. Distribution of permafrost-related features since the Last Glaciation in the Iberian Peninsula. The symbols of the legend are the same for all the other graphs.

Figure 3. Examples of permafrost-related landforms generated during different phases in different massifs of the Iberian Peninsula.

Figure 4. Distribution of permafrost-related features since the Last Glaciation in the southern Alps.

Figure 5. Minimum mean altitude of the relict rock glacier fronts in the southern side of the Alps, including data from (1) Bornet et al. (2014); 2) Seppi et al. (2012); (3) Colucci et al. (2016a).

Figure 6. Examples of periglacial and permafrost-related landforms generated during different phases in the southern Alps.

Figure 7. Distribution of permafrost-related features since the Last Glaciation in the Italian Peninsula.

Figure 8. Examples of permafrost-related landforms generated during different phases in the Apennines and Italian Peninsula.

Figure 9. Distribution of permafrost-related features since the Last Glaciation in the Balkan Peninsula.

Figure 10. Examples of permafrost-related landforms generated during different phases in different massifs of the Balkan Peninsula.

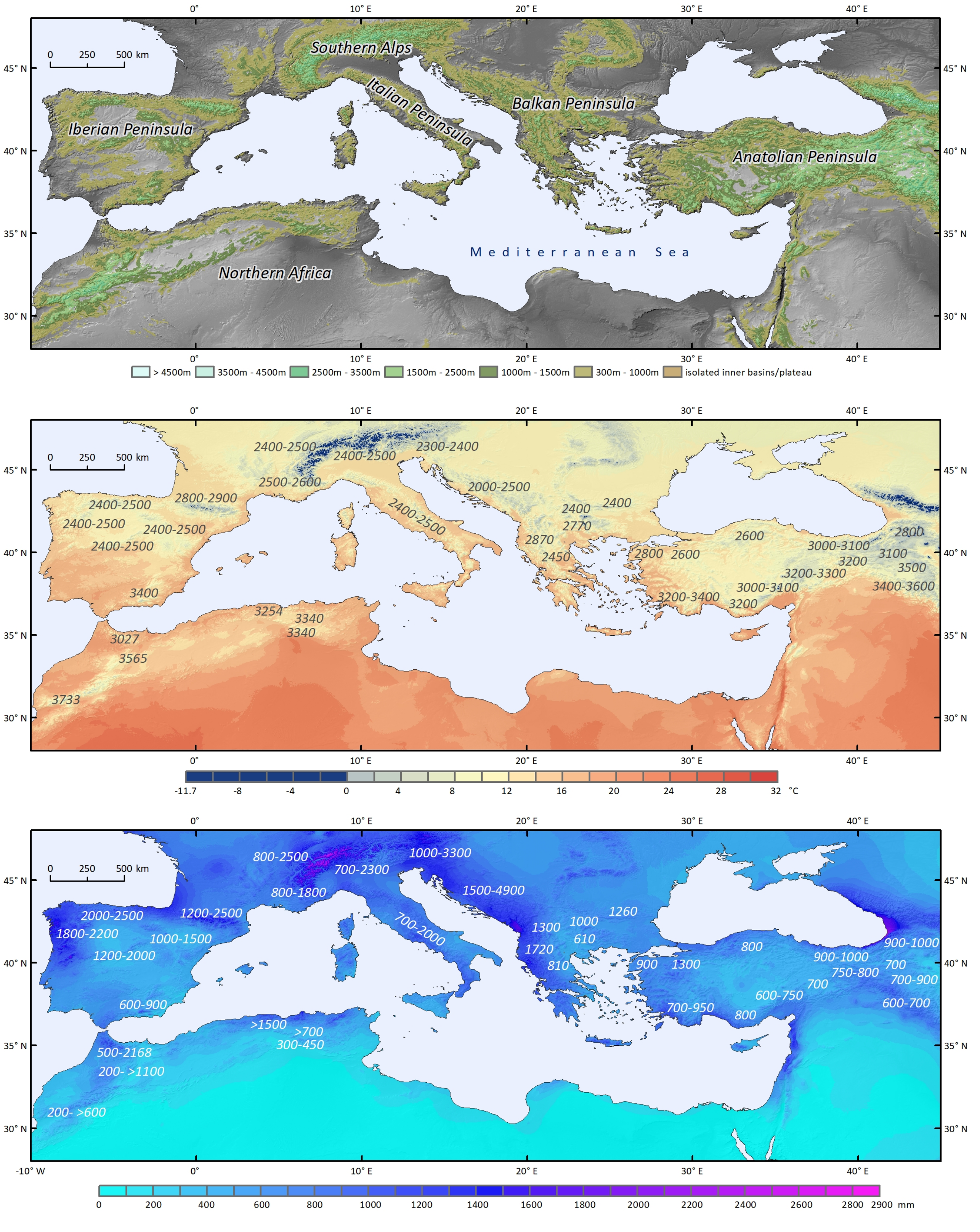
Figure 11. Distribution of permafrost-related features since the Last Glaciation in the Anatolia Peninsula.

Figure 12. Examples of permafrost-related landforms generated during different phases in different massifs of the Anatolia Peninsula.

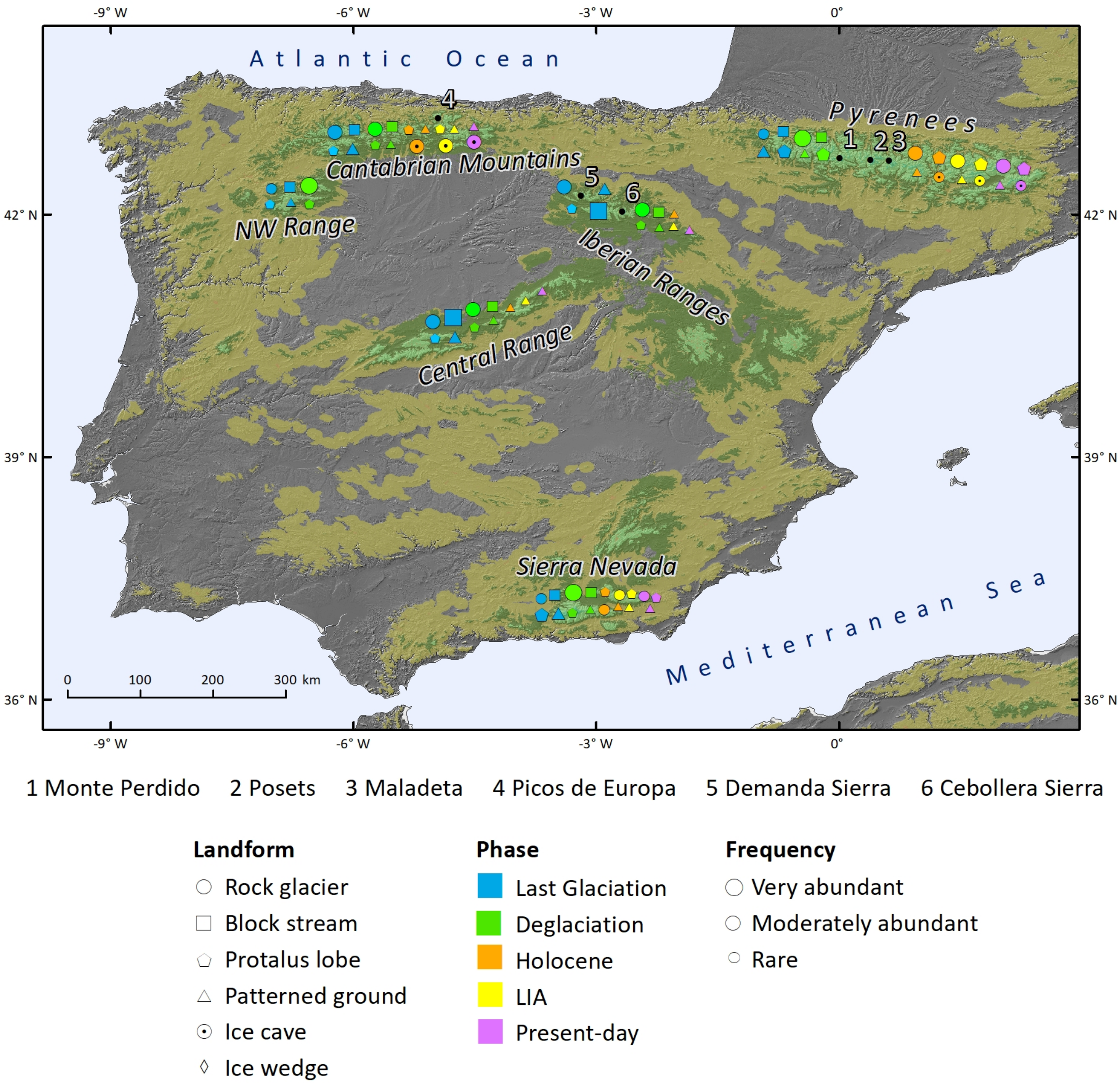
Figure 13. Distribution of permafrost-related features since the Last Glaciation in northern Africa.

Figure 14. Examples of permafrost-related landforms generated during different phases in northern Africa.

Figure 15. Geomorphological sketch of the formation of different generations of permafrost-related features in Mediterranean mountains since the Last Glaciation.

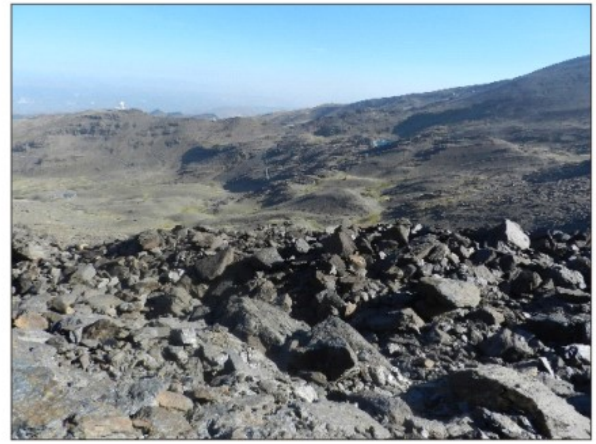


0°
10° E
20° E
30° E
40° E
45° N
40° N
35° N
30° N
0 250 500 km
Southern Alps
Italian Peninsula
Balkan Peninsula
Iberian Peninsula
Anatolian Peninsula
Northern Africa
M e d i t e r r a n e a n S e a
> 4500m
3500m - 4500m
2500m - 3500m
1500m - 2500m
1000m - 1500m
300m - 1000m
isolated inner basins/plateau
2400-2500
2400-2500
2300-2400
2500-2600
2000-2500
2400-2500
2800-2900
2400-2500
2400-2500
2400-2500
2400-2500
2400
2400
2770
2870
2450
2600
2800
2600
3000-3100
3200
2800
3100
3500
3200-3300
3000-3100
3200-3400
3200
3400-3600
3400
3254
3340
3340
3027
3565
3733
-11.7
-8
-4
0
4
8
12
16
20
24
28
32
°C
800-2500
1000-3300
700-2300
1500-4900
800-1800
2000-2500
1200-2500
1800-2200
1000-1500
1200-2000
700-2000
1300
1000
1260
610
1720
810
900
1300
800
900-1000
900-1000
750-800
700
700-900
700
600-750
600-700
700-950
800
600-900
>1500
>700
300-450
500-2168
200- >1100
200- >600
-10° W
0
200
400
600
800
1000
1200
1400
1600
1800
2000
2200
2400
2600
2800
2900
mm

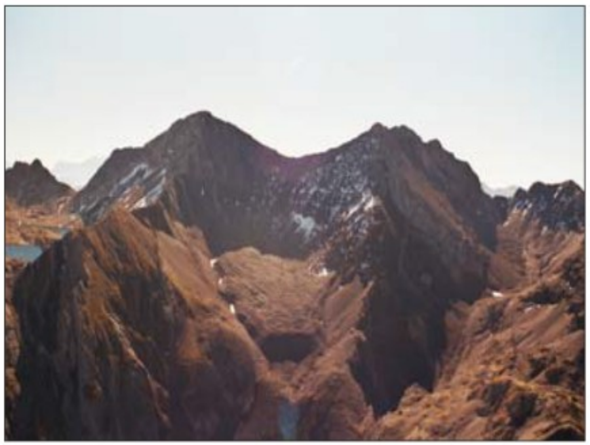

Atlantic Ocean
Pyrenees
Cantabrian Mountains
NW Range
Iberian Ranges
Central Range
Sierra Nevada
Mediterranean Sea
1 2 3 4 5 6
-9° W
-6° W
-3° W
0°
42° N
39° N
36° N
0 100 200 300 km
1 Monte Perdido
2 Posets
3 Maladeta
4 Picos de Europa
5 Demanda Sierra
6 Cebollera Sierra
Landform
Rock glacier
Block stream
Protalus lobe
Patterned ground
Ice cave
Ice wedge
Phase
Last Glaciation
Deglaciation
Holocene
LIA
Present-day
Frequency
Very abundant
Moderately abundant
Rare

Last Glaciation (block streams; Iberian Range)

Deglaciation (protalus lobes; Sierra Nevada)

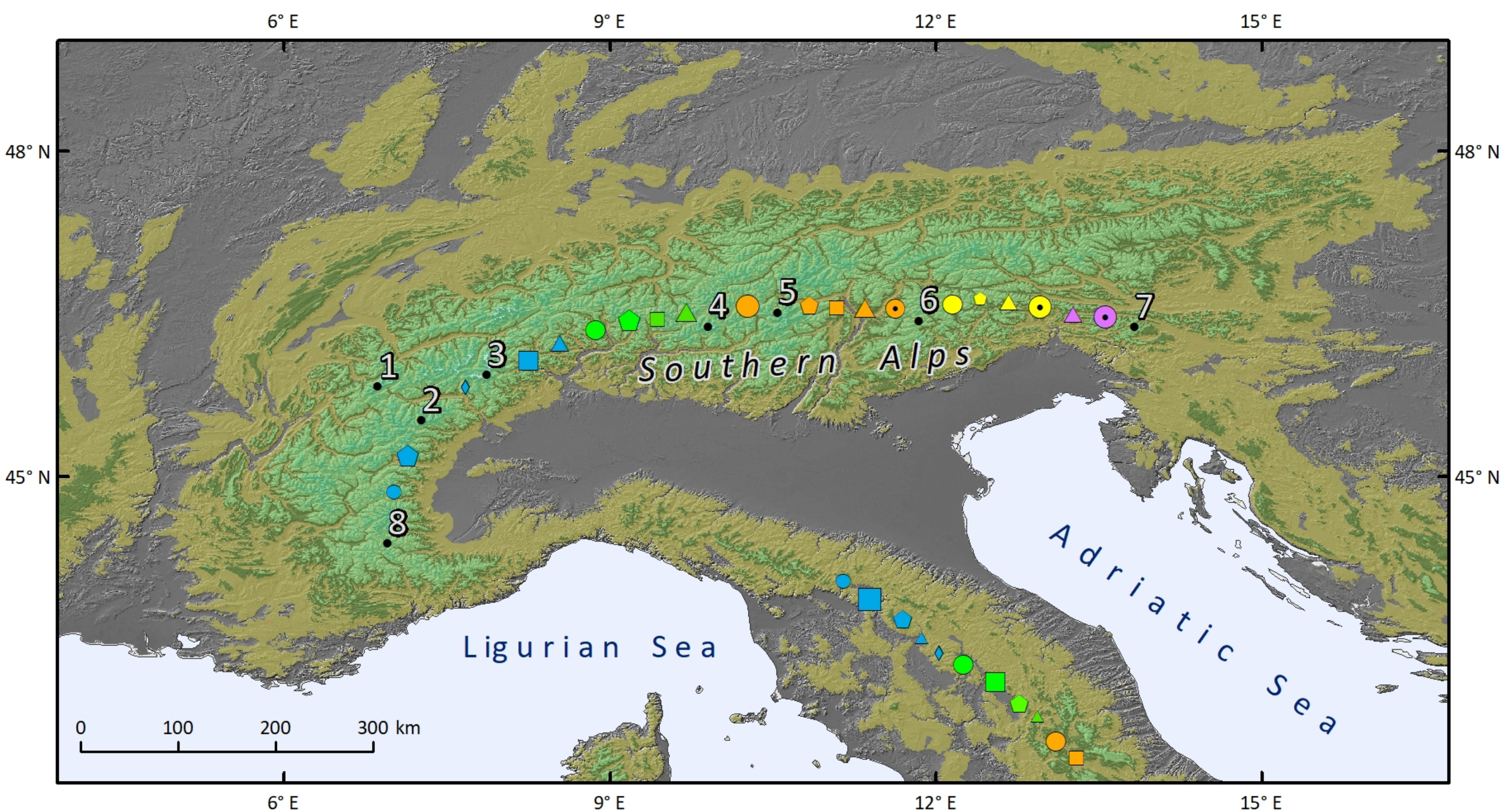


Holocene (rock glacier; Sierra Nevada)



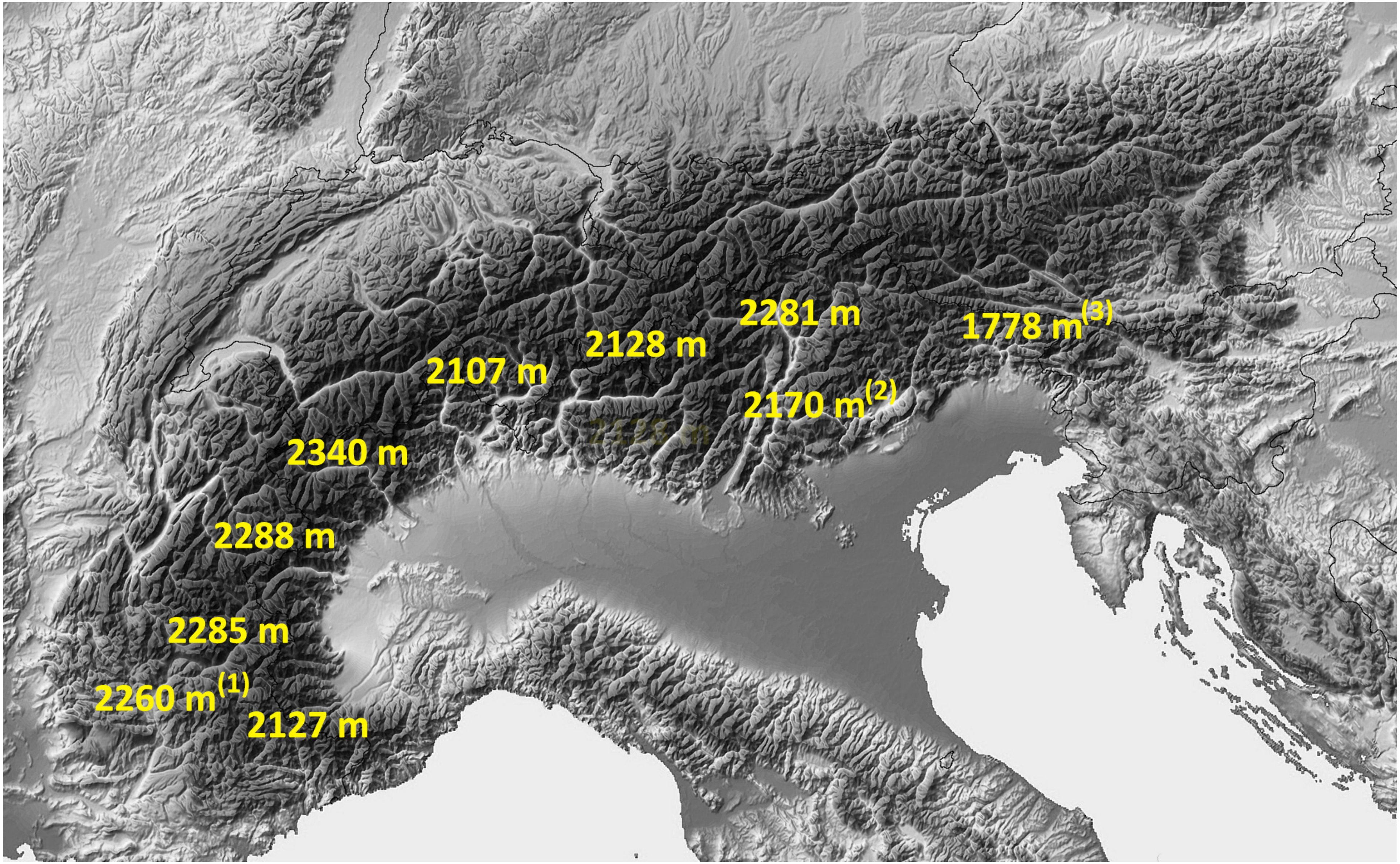
LIA (rock glacier; Pyrenees)

Present-day (rock glacier; Sierra Nevada)



1 Mt. Bianco 2 Gran Paradiso 3 Mt. Rosa 4 Piz Bernina 5 Mt. Ortles 6 Marmolada

7 Mt. Triglav 8 Argentera


2281 m
1778 m(3)
2128 m
2107 m
2170 m(2)
2340 m
2288 m
2285 m
2260 m(1)
2127 m

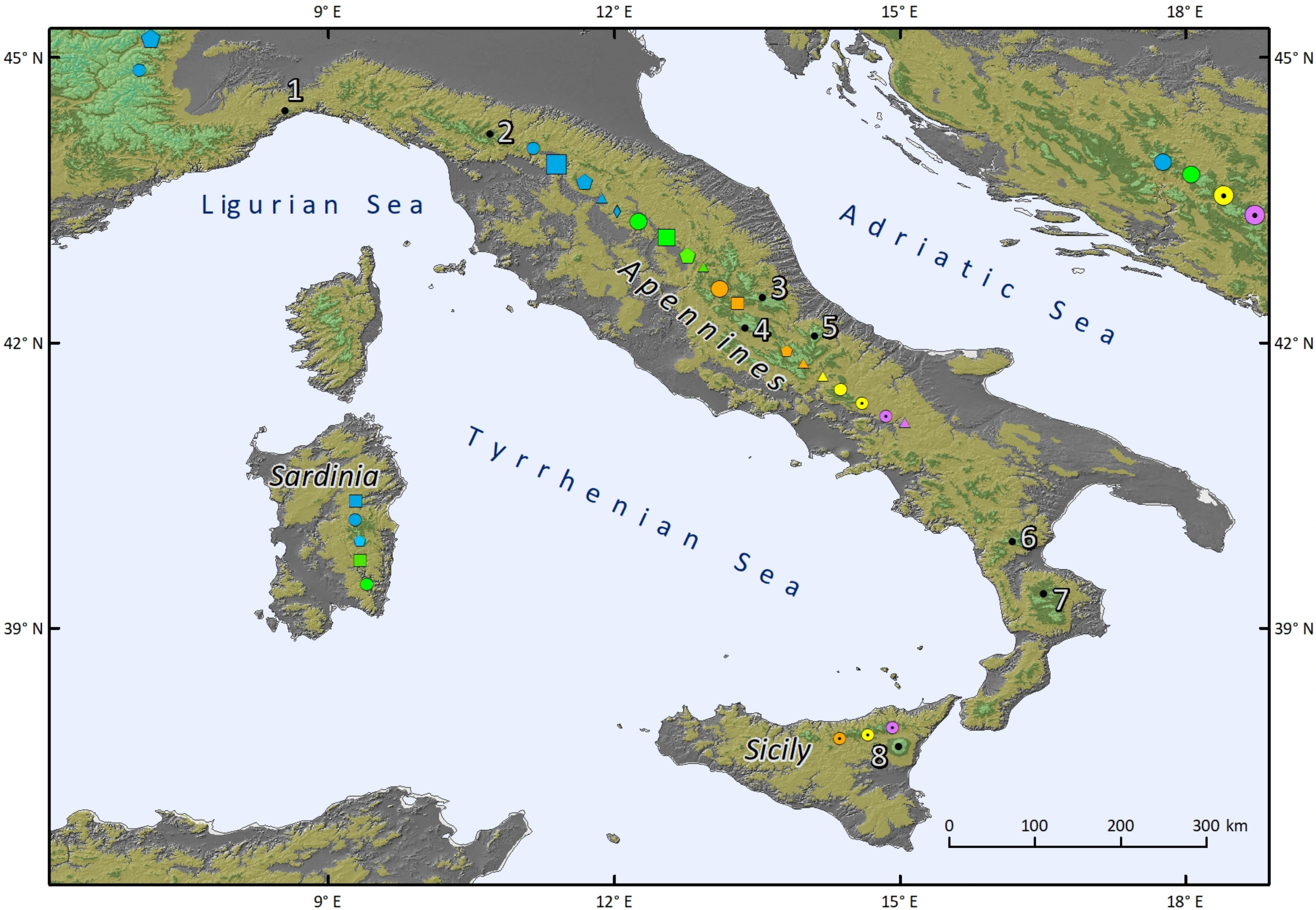
LGM (relict rock glacier, central Alps)
Photo: M.Guglielmin

Deglaciation (patterned ground; central Alps)
Photo: M. Guglielmin

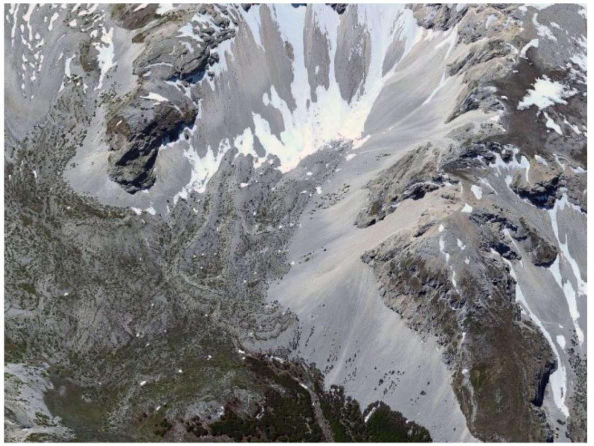
Holocene (active rock glacier; central Alps)
Photo: R.R.Colucci

LIA (active rock glacier; Dolomites)
aerial view from free BingMaps

Present-day (ground ice, ice cave; eastern Alps
Photo: R.R.Colucci)



1 Mt. Beigua 2 Mt. Cimonel 3 Gran Sasso 4 Mt. Velino 5 Mt. Maiella 6 Mt. Pollino 7 Mt. Sila 8 Mt. Etna



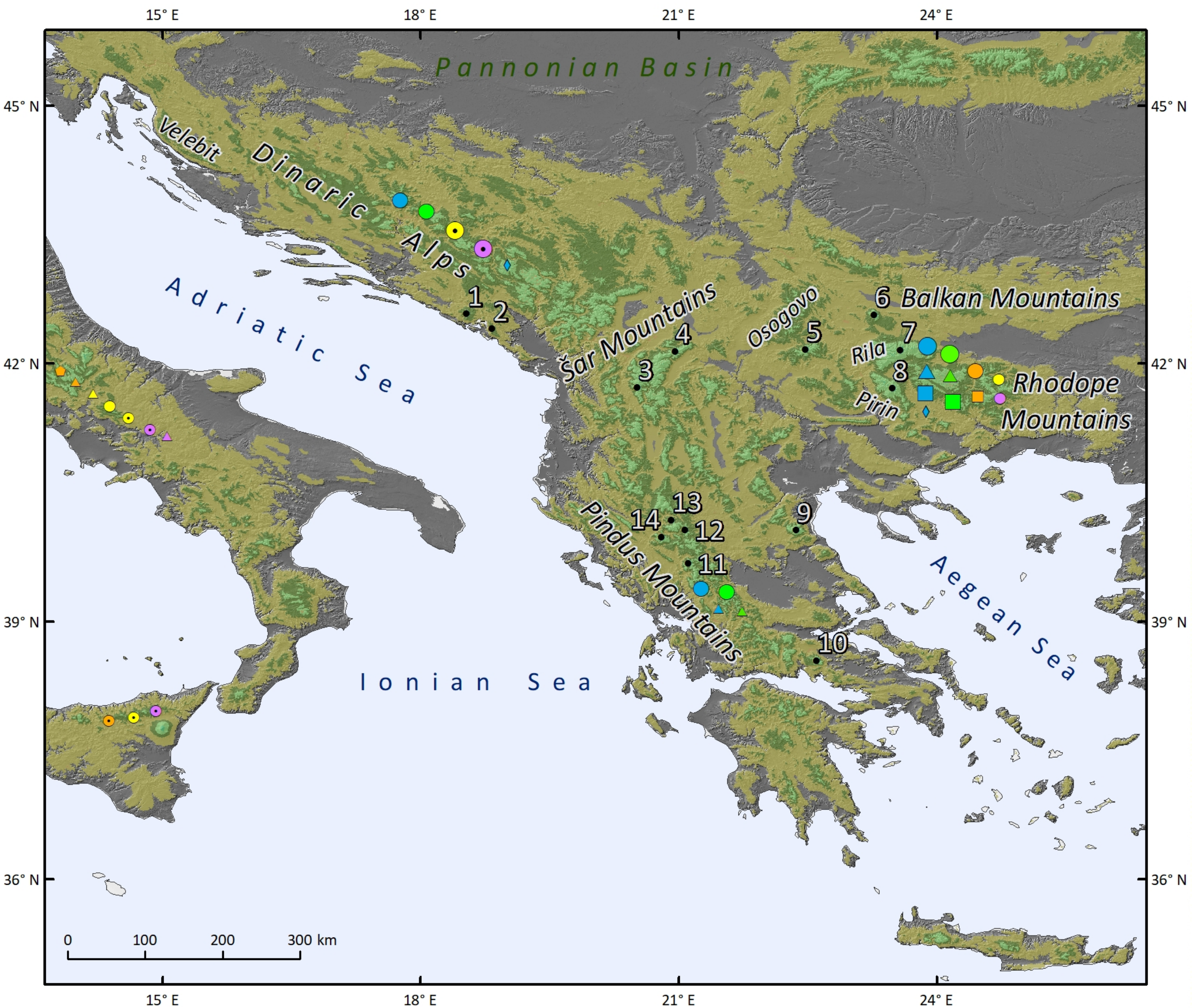
LGM (block stream; Apennines
Photo: M.Guglielmin

Deglaciation (relict rock glacier; Apennines)
aerial view from free BingMaps

Holocene (protalus lobe; Apennines)
Photo: R.R.Colucci

LIA (active rock glacier; Apennines)
Photo: M.Leoni

Present-day (ground ice, ice cave; Sicily)
Photo: M.Restivo



1 Mt. Orjen 2 Mt. Lovćen 3 Mt. Korab 4 Mt. Bistra 5 Mt. Ruen 6 Mt. Vitosha
7 Mt. Musala 8 Mt. Polezhan 9 Mt. Olympus 10 Mt. Parnassus 11 Mt. Lakmos-Peristeri
12 Mt. Vasilitsa 13 Mt. Smolikas 14 Mt. Tymphi

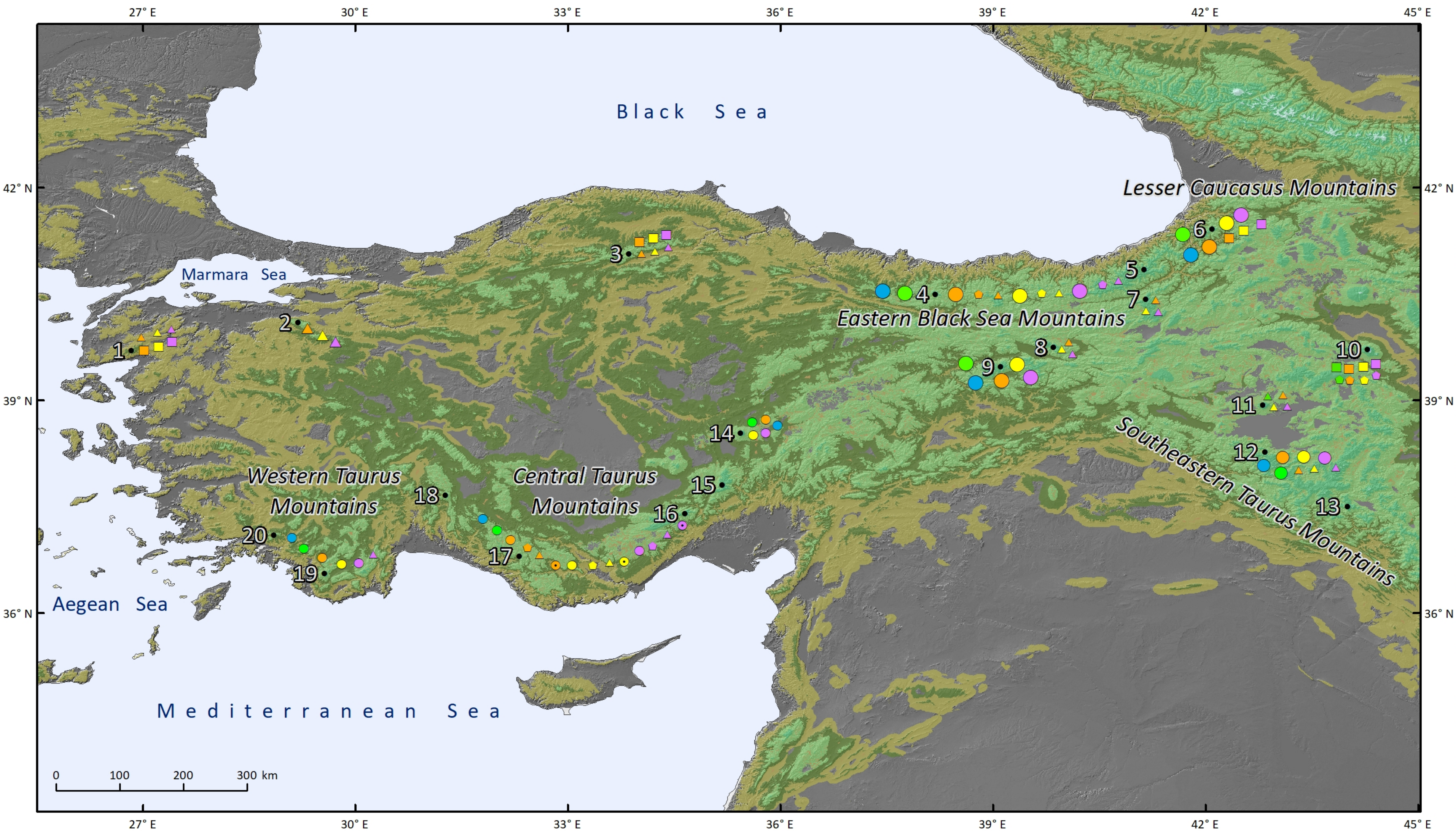
Last Glaciation (rock glacier; Pindus Mountain)
Photo: P.D. Hughes

Deglaciation (rock glacier; Rila Mountain)
Photo: B. Magori

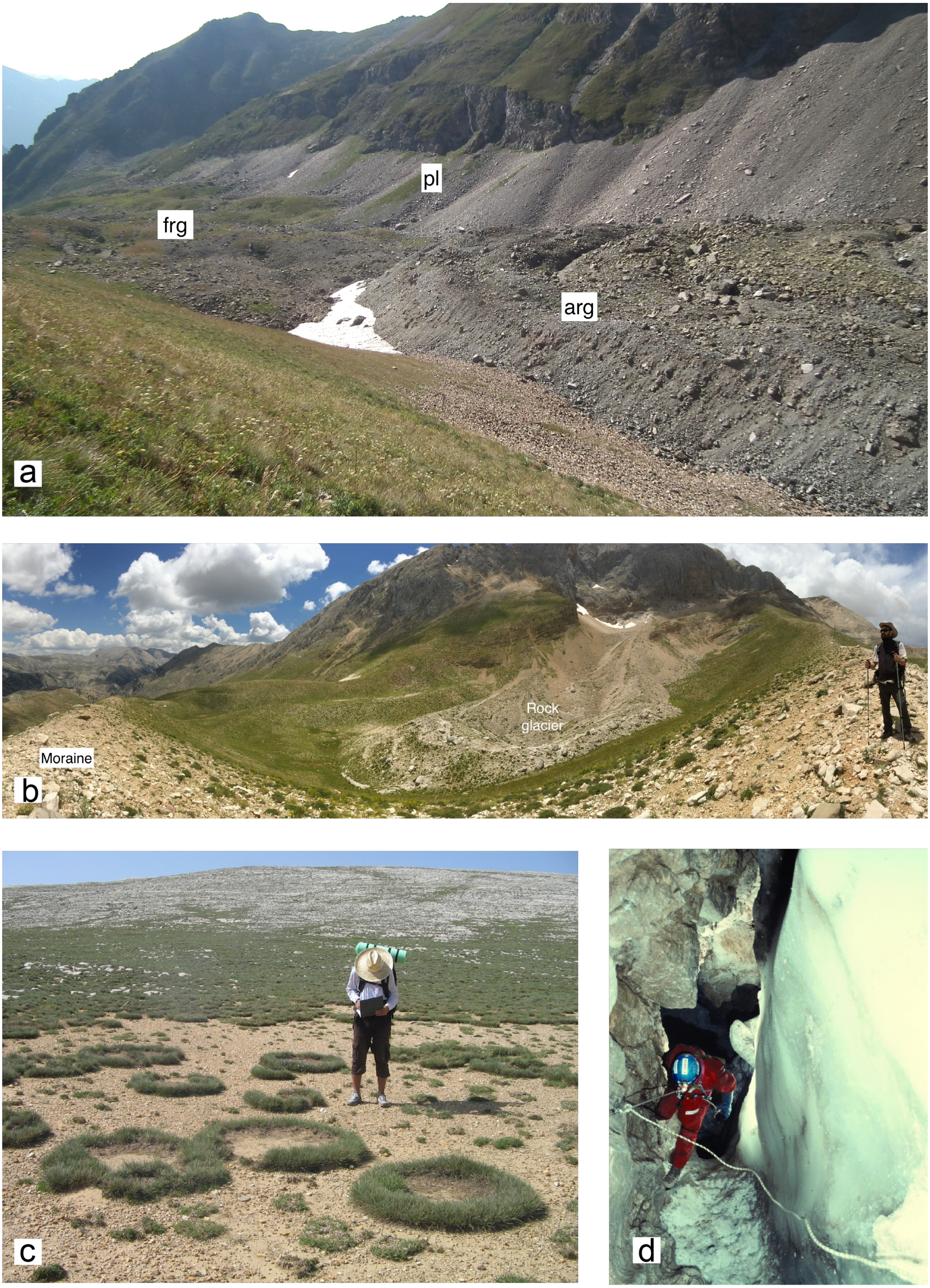
Holocene (rock glacier; Pirin Mountain)
Photo: B. Magori

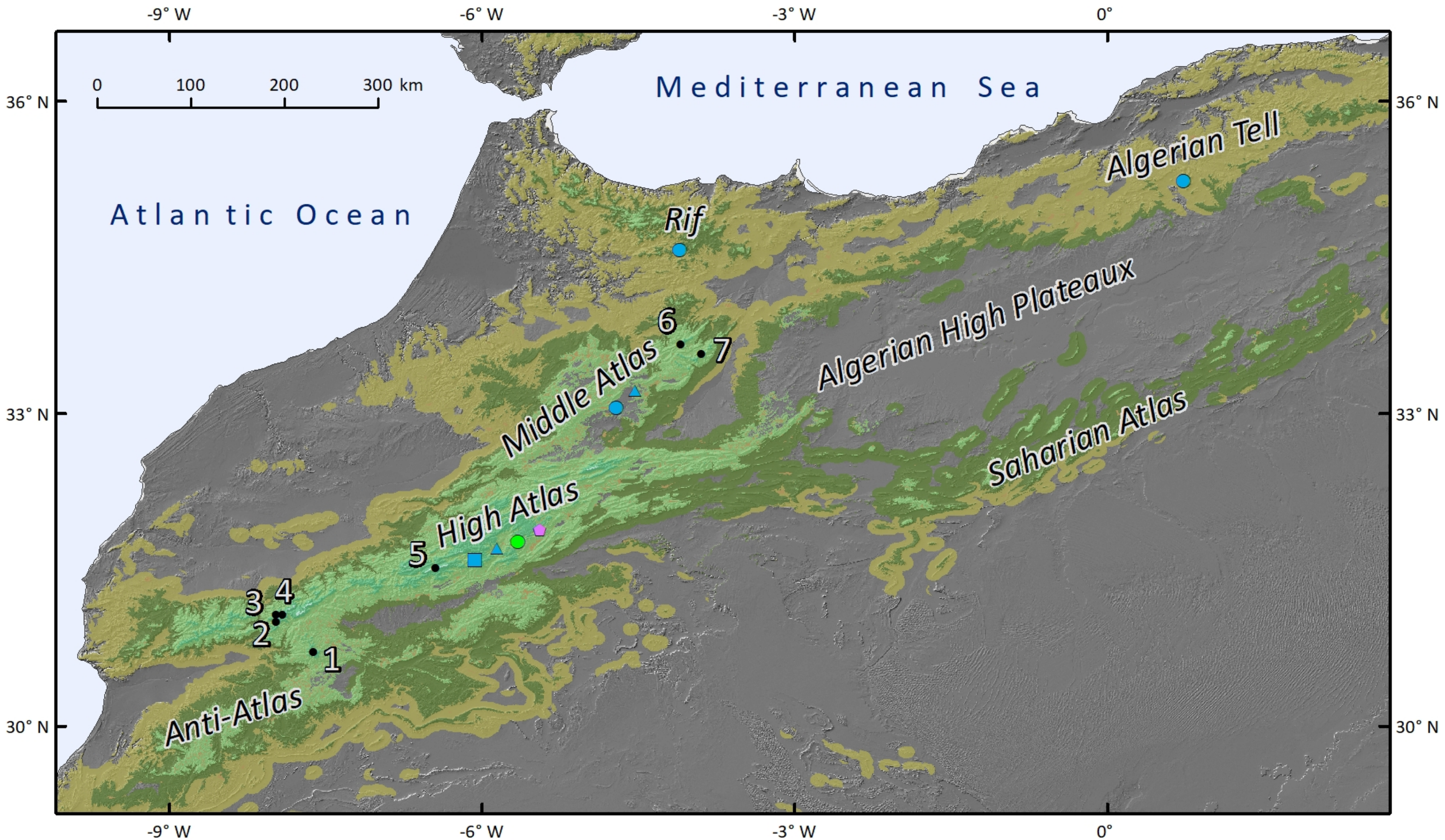
LIA (glacieret; Pirin Mountain)
Photo: B. Magori

Present-day (ice cave; Trnovski gozd, Dinaric Alps)
Photo: R.R. Colucci



1 Mt. Kazdağ 2 Mt. Uludağ 3 Mt. Ilgaz 4 Mt. Karagöl 5 Mt. Kaçkar 6 Mt. Karçal 7 Mt. Mescid 8 Mt. Esence (Keşiş)
9 Mt. Mercan (Munzur) 10 Mt. Ağri (Ararat) 11 Mt. Süphan 12 Mt. İhtiyar Şahap (Kavuşşahap) 13 Mt. Cilo 14 Mt. Erciyes
15 Mt. Aladağlar 16 Mt. Bolkar 17 Mt. Geyikdağ 18 Mt. Dedegöl 19 Mt. Akdağ 20 Mt. Sandıras


pl
frg
arg
a
Moraine
Rock glacier
b
c
d



1 Djebel Sirwa 2 Iouzagner 3 Tazaghart 4 Djebel Toubkal 5 M'Goun 6 Djebel Bou Iblane 7 Djebel Bou Naceur

Table 1. Study areas of this research together with the main massifs and altitude distribution of present-day climate conditions and elevation of the lower periglacial belt.

Region	Mountain range (highest peaks, m asl)	Annual precipitation (mm)	*Altitude of 0 °C isotherm (m asl)	Lower altitude periglacial belt (m)
Iberian Peninsula	Pyrenees (Aneto, 3404; Posets, 3371)	1200-2500	2800-2900	2100 (N), 2300 (S)
	Cantabrian Mountains (Torre Ceredo, 2650; Torre del Llambrión, 2642)	2000-2500	2400-2500	1900 (N), 2100 (S)
	NW ranges (Peña Trevinca, 2127)	1800-2200	2400-2500	1900 (N), 2100 (S)
	Central Range (Pico Almanzor, 2591)	1200-2000	2400-2500	2100 (N), 2300 (S)
	Iberian Range (Moncayo, 2313; San Lorenzo)	1000-1500	2400-2500	2100 (N), 2300 (S)
	Betic Range (Mulhacén, 3478; Veleta, 3398)	600-900	3400	2500 (N), 2650 (S)
Southern Alps	Western Alps (Monte Bianco, 4810; Monte Rosa, 4634)	800-2500	2400-2500	2200 (N);2500 (S)
	Central Alps (Piz Bernina, 4059; Cima de Piazzi, 3439)	700-2300	2300-2400	2200 (N); 2400 (S)
	Eastern Alps (Ortles, 3905; Triglav, 2869)	1000-3300	2400-2500	2100 (N); 2400 (S)
	Maritime Alps (Argentera, 3297)	800-1800	2500-2600	2200 (N); 2500 (S)
Italian Peninsula	Northern Apennine (Cimone, 2165)	1100-2300	-	-
	Central Apennine(Gran Sasso, 2912; La Majella, 2793)	700-1300	2400-2500	2200 (N)
	Southern Apennine (Pollino, 2233; Sila, 1930)	700-1600	-	-
Balkan Peninsula	Dinaric Alps (Maja Jezercë, 2694)	1500-4900	2000-2500	
	Pirin Mountains (Vihren, 2914)	610	2770	
	Stara Planina (Botev, 2376)	1260	2400	
	Rila Mountains (Musala, 2925)	> 1000	2400	
	Pindus Mountains (Smolikas, 2637)	1720	2870	
	Mount Olympus (Mount Olympus , 2917)	810	2450	
	Šar Mountains (Titov Vrv, 2747)	1300	-	
Anatolia	Western Taurus Mountains (Akdağ, 3016; Dedegöl, 2992; Geyikdağ, 2877)	700-950	3200-3400	2500
	Central Taurus Mountains (Aladağlar, 3756; Bolkar, 3524)	800	3200	3200
	Southeastern Taurus Mountains (Cilo, 4135)	600-700	3400-3600	?
	Eastern Black Sea Mountains (Kaçkar, 3937)	900-1000	3000-3100	1800-2200 (N)
	Karçal (Lesser Caucasus Mts.) (Karçal, 3932)	900-1000	2800	2100
	Kazdağ (Kazdağ, 1774)	900	2800	1350 (N)
	Uludağ (Uludağ, 2543)	1300	2600	1900
	Ilgaz (Ilgaz, 2546)	800	2600	?
	Esence (Keşiş, 3549)	750-800	3200	2500 (N)
	Mercan (Munzur, 3463)	700	3200-3300	?
	Mescid (3239)	700	3100	2650
	Erciyes (Erciyes, 3917)	600-750	3000-3100	2950
	Ağrı (Ararat, 5137)	700-900	3500	3000
Northern Africa	Hoggar (Mount Tahat, 2981)	145	4904	NA
	Tibesti (Emi Koussi, 3445)	11	4516	NA
	Rif Mountains (Tidirhine, 2456)	500 - 2168	3027	NA
	Middle Atlas (Djebel Bou Naceur, 3340)	200 - >1100	3565	2500
	High Atlas (Djebel Toubkal, 4167)	200 - >600	3733	2500-3000
	Algerian Tell (Djurdjura, 2308)	>1500	3254	1500-1800 (?)
	Algerian High Plateaux (Djebel Guerioun, 1729)	>700	3340	NA
	Aurès Massif (Aurès, 2324)	300-450	3340	2300

*Extrapolated used on a lapse rate of 0.6 C per 100 m altitude.

Table 2. Periglacial processes and landforms in the Iberian Peninsula since the Last Glaciation until today.

Phase	Areas	Environment: landforms and processes	Chronology (ka BP)	References
	Pyrenees	Non-glaciated summit plateaus: patterned ground with meter-sized sorted circles Mid-low slopes: rock glaciers, block streams Valleys: fragipan and permafrost evidences >1130 m Stratified debris above 800-1000 m	> 32 18-20 20-22	Gómez-Ortiz and Serrat (1979), Gómez-Ortiz (1980), Chueca et al. (1994), Peña et al. (1998, 2000), Serrano (1998), García-Ruiz et al. (2001, 2005), Chueca and Julián (2008), Boixadera et al. (2008)
	Cantabrian Mountains	Summit plateaus: patterned ground with sorted circles Mid slopes > 1300 m: rock glaciers, block streams, blockfields Stratified debris above 400 m Ice wedge cast above 700 m		García de Celis (1991, 2002), Ugarte (1992), Pérez-Alberti et al. (1994), Pérez-Alberti and Rodríguez-Gutián (1993), Castañón and Frochoso (1994, 1998), Valcárcel (1998), Pérez-Alberti and Valcárcel (2000), Válcárcel and Pérez-Alberti (2002), González-Gutiérrez (2002), González-Trueba (2007a,b), Pellitero (2009, 2012), Rodríguez-Pérez (2009), Santos-González (2010), Pellitero et al. (2011), Serrano et al. (2013), Ruiz-Fernández (2013)
Last Glaciation	NW ranges	Slopes > 700 m: rock glaciers, block streams; > 1500 m: block slopes	> 44 to 16	Pérez-Alberti (1979), Pérez-Alberti and Rodríguez-Guitián (1993), Pérez-Alberti et al. (1994, 1998, 2011), Valcárcel (1998), Valcárcel (1998), Valcárcel et al. (1998), Martínez-Cortizas et al. (1999), Valcárcel and Pérez-Alberti (2002), Hall-Riaza et al. (2016)
	Central Range	Non-glaciated summit plateaus: blockfields, tors Mid slopes > 1500 m: scarce rock glaciers, block slopes, solifluction Low slopes 700-1200 m: periglacial slope deposits (stratified and head), with possible permafrost conditions	<30	Ontañón and Asensio-Amor (1973), Daveau (1973, 1978) Daveau et al. (1997), Ontañón (1985), Sanz-Herráiz (1988), Ferreira et al. (2000), Palacios et al. (2003), Vieira (2004), Carrasco et al. (2012)
	Iberian Range	High slopes > 1800 m: block streams, rock glaciers Slopes > 1100 m: stratified debris Non-glaciated summit plateaus: patterned ground, cryoturbation, blockfields		Thornes (1968), Asensio-Amor (1974), Gutiérrez-Elorza and Peña (1977), García-Ruiz (1979), Pellicer (1980), Ortigosa (1986), Arnáez (1987), Sanz and Pellicer (1994), Peña and Lozano (1998), Arnáez and García-Ruiz (2000)
	Sierra Nevada	Non-glaciated summit plateaus > 3000 m: development of patterned ground with meter-sized sorted circles Non-glaciated high slopes (2500 m): formation of three rock glacier	20-30	Messerli (1965), Gómez-Ortiz (1987, 2002), Simón et al. (2000), Gómez-Ortiz and Salvador (1992), Gómez-Ortiz et al. (1994), Oliva (2011), Palma et al. (2017)
	Lowlands	Ice-wedges in Pleistocene fluvial terraces (200-1000 m)		Badorrey et al. (1970), Asensio-Amor and González-Martín (1974), Serrano et al. (2010a)
	Pyrenees	Glacial cirques and summits: rock glacier formation >2250 m, with widespread permafrost >2490 m, patterned ground formation Glacial cirques and summits: rock glaciers becoming gradually inactive <2300 m and new ones forming >2350 m, permafrost widespread >2525 m Stratified debris > 700 m	15-17 11.5-12.9	Gómez-Ortiz (1980), Soutadé (1980), Chueca et al. (1994), Serrano (1998), García-Ruiz et al (2000, 2001, 2005, 2015, 2016), Hirsch and Raab (2014), Palacios et al. (2015)
Deglaciation	Cantabrian Mountains	Summit plateaus: patterned ground formation Glacial cirques: rock glaciers, block streams		Alonso (1989), Redondo et al. (2002, 2004, 2010), González-Trueba (2007a,b), Alonso and Trombotto (2009), Rodríguez-Pérez (1995, 2009), Santos-González (2010), Pellitero et al. (2011), Pellitero (2012), Serrano et al. (2013), Ruiz-Fernández (2013)
	NW ranges	High slopes: rock glaciers		Valcárcel (1998), Valcárcel and Pérez-Alberti (2002)

	Central Range	Glacial cirques: formation of rock glaciers Slopes: solifluction lobes	15-17	Palacios et al. (2011a,b, 2012), Carrasco et al. (2012, 2015)
	Iberian Range	Glacial cirques: formation of rock glaciers and block streams		García-Ruiz (1979), Ortigosa (1986), Pellicer (1980), Sanz and Pellicer (1994), García-Ruiz et al. (1998)
	Sierra Nevada	Glacial cirques: formation of rock glaciers and protalus lobes	12.8-7.4	Palade et al. (2011), Gómez-Ortiz et al. (2012a, 2013), Oliva (2009), Oliva et al. (2014), Palacios et al. (2016)
Holocene	Pyrenees	Glacial cirques and high slopes: degradation of permafrost in the highest areas and probable complete deglaciation, elevation rise of snow patches, formation of rock glaciers, development of ice caves	Mid Holocene	Grove and Gellatly (1995), Serrano (1998), García-Ruiz et al. (2005), Serrano et al. (2010b, 2011), Sancho et al. (2016)
	Cantabrian Mountains	Glacial cirques: active rock glaciers becoming gradually inactive	Cold stages	Alonso and Trombotto (2009), Pellitero et al. (2011), Serrano et al. (2013)
	Sierra Nevada	Glacial cirques: rock glaciers and protalus lobes becoming gradually inactive	Until 7.4	Gómez-Ortiz et al. (2012a, 2013), Oliva and Gómez-Ortiz (2011)
LIA	Pyrenees	Glacial cirques and high slopes: glacial advance, greater abundance of snow patches, formation and reactivation of rock glaciers and protalus lobes, permafrost >2560 m, patterned ground formation, development of ice caves	XIV-XIX centuries	Martínez de Pisón and Arenillas (1988), García-Ruiz et al. (1988), Copons and Bordonau (1994), Grove and Gellatly (1995), Julián and Chueca (1998), Lugon et al. (2004), Serrano (1998), Serrano et al. (2001, 2002), García-Ruiz and Martí-Bono (2001), Chueca et al. (2005), González-Trueba et al. (2008); Fernandes et al. (2017), Leunda et al. (2015), Bartolomé et al. (2015), Sancho et al. (2016)
	Cantabrian Mountains	Highest cirques > 2200 m: development of small glaciers, greater abundance of snow patches, frost mounds, probable permafrost > 2400 m in northern cirques with seasonal ground ice above 1600 m, development of ice caves	XIV-XIX centuries	González-Trueba (2007a,b), Serrano et al. (2013), Pellitero (2014)
	Sierra Nevada	Highest cirques > 3000 m: presence of glaciers in the highest northern cirques, more extensive snow fields >2500 m, possible permafrost conditions near glaciers	XVII to XIX centuries	Gómez-Ortiz and Plana-Castellví (2006), Gómez-Ortiz et al. (2009, 2012b), Oliva and Gómez-Ortiz (2012), Oliva et al. (2011)
Present-day	Pyrenees	Cirques and high slopes ≥ 2500 m: probable permafrost above ~2630 m (N), ~2800 m (S), active rock glaciers above 2510 m, protalus lobes, ice patches, active ice cave processes, frost mounds, patterned ground formation	since the late XIX century	Soutadé (1980), Gómez-Ortiz (1980), Höllermann (1985), Chueca (1992), Serrano and Agudo (1998, 2004), Serrano et al. (1999, 2000, 2001, 2002, 2006, 2009, 2010b, 2010c, 2011a), Chueca et al. (2000), Julián and Chueca (2007), Feuillet (2010), Feuillet and Mercier (2012), Bartolomé et al. (2015), González-García (2014), García-Ruiz et al. (2015), González-García et al. (2017)
	Cantabrian Mountains	Highest cirques > 2200 m: degradation of buried ice and sporadic permafrost (e.g. Jou Negro, Forcadona), ice patches, active ice cave processes, frost mounds	since the late XIX century	Castañón and Frochoso (1998), González-Trueba (2007a,b), Serrano et al. (2011b), Ruiz-Fernández (2013), Gómez-Lende et al. (2014), Ruiz-Fernández et al. (2014), Pisabarro et al. (2016)
	Sierra Nevada	Highest cirques > 3000 m: formation of rock glaciers, degradation of buried ice and permafrost, with subsidence and collapses of the rock glacier of the Veleta cirque (accelerating during the last decade)	since the late XIX century	Gómez-Ortiz et al. (2001, 2004, 2014), Salvador-Franch et al. (2010, 2011), Tanarro et al. (2010), Oliva et al. (2016b)

Table 3. Periglacial processes and landforms in the Southern Alps since the Last Glaciation until today.

Phase	Areas	Environment: landforms and processes	Chronology (ka BP)	References
Last Glaciation	Po plain	Cryoturbations, stratified slope deposits, loess cover	35-LGM	Cremaschi et al. (2005), Guglielmin (unpublished data)
	Piedmont	Block streams, blockfields	LGM	Fioraso and Spagnolo, (2009), Paro (2011)
	Julian Alps	Rock glacier at 1076 m on average	LGM	Colucci et al. (2016a)
Deglaciation	Southern Alps	Rock glaciers at 2260 m on average	YD	Bornet et al. (2014)
	Western Alps	Rock glaciers at 2340-2260 m on average	YD	Guglielmin and Smiraglia (1997), Bornet et al. (2014)
	Central Alps Julian Alps	Rock glaciers at 2280-2160 m on average Rock glaciers at 1778 m on average	YD YD	Guglielmin and Smiraglia (1997), Seppi et al. (2012), Scotti et al. (2013), Colucci et al. (2016a)
Holocene	Western Alps	Rock glaciers at different altitudes		Dramis et al. (2003)
	Central Alps	Rock glaciers at different altitudes	8.9, 2.2 2.7-2.9 0.9-1.4	Krainer et al. (2015) Calderoni et al. (1998), Dramis et al. (2003) Calderoni et al. (1998), Stenni et al. (2007), Scapozza et al. (2010)
LIA	Southern Alps	Active rock glaciers at 2500 m on average	LIA-present	Ribolini et al. (2010)
	Western Alps	Active rock glaciers at 2647 m on average	LIA-present	Guglielmin and Smiraglia (1997)
	Central Alps Julian Alps	Active rock glaciers at 2526 m on average	LIA-present	Guglielmin and Smiraglia (1997), Seppi et al. (2012)
Present-day	Southern Alps	Active rock glaciers at 2500 m on average		Ribolini et al. (2010)
	Western Alps	Active rock glaciers at 2647 m on average		Guglielmin and Smiraglia (1997)
	Central Alps Julian Alps	Active rock glaciers at 2526 m on average BTS measurements suggest permafrost at 2258 m		Guglielmin and Smiraglia (1997), Seppi et al. (2012), Colucci et al. (2016a)

Table 4. Periglacial processes and landforms in the Italian Peninsula since the Last Glaciation until today.

Phase	Areas	Environment: landforms and processes	Chronology (ka BP)	References
Last Glaciation	Northern Appenines	Rock glaciers around 2000 m. Block streams at Mt. Beigua, Stratified scree slopes	LGM-YD	Chelli and Tellini (2002), Firpo et al. (2006), Federici (1981), Rellini et al. (2014)
	Central Appenines	Several rock glaciers around the highest mountain. Stratified scree slope from the coast upward; cryoturbation in the Mt. Beigua area	LGM-YD	Coltorti et al. (1979), Castiglioni et al. (1979), Boenzi (1980), Giraudi (2002)
	Southern Appenines	One rock glacier at Mt. Pollino and several stratified scree slopes and one sand wedge at 1350 m at Sila Mountain. Cryoturbation at Cilento Mountains at 30 m	LGM-YD	Dimase (2006), Scarciglia et al. (2003)
Deglaciation				
Holocene	Central Appenines	A few rock glaciers developed at 1900-2000 m around the highest mountains	> 7	Dramis et al. (2003)
LIA	Central Appenines	One rock glacier is still active at 2520 m at La Majella Mountain. Permafrost can be present around the highest mountains at more than 2200 m		Dramis and Kotarba (1992), Dramis et al. (2003)
Present-day	Central Appenines	Permafrost is documented at more than 2300 m around La Majella and Mt. Velino by BTS measurements		Bisci et al. (2003), Guglielmin (unpublished)

Table 5. Periglacial processes and landforms in the Balkan Peninsula since the Last Glaciation until today.

Phase	Areas	Environment: landforms and processes	Chronology (ka BP)	References
Last Glaciation	Dinaric Alps*	Probable discontinuous permafrost, ice wedges, cryoturbations		Liedtke (1962), van Vliet-Lanoe et al. (2004), Ruszkiczay-Rüdiger and Kern (2015)
	Bulgarian Mountains	Discontinuous and sporadic permafrost above 1200 m, rock glaciers, block streams, patterned ground		King and Akerman (1993), Van Vliet-Lanoë and Hallegouët (2001), Dobinski (2005), Gikov and Dimitrov (2011), Dimitrov and Gikov (2012), Kuhlemann et al. (2013)
	Mountains in Greece	Probable discontinuous permafrost, ice wedges, cryoturbations, patterned ground Glacial cirques: possible rock glacier formation above ~1800 m		Hughes et al. (2003)
Deglaciation	Dinaric Alps*	Glacial cirques: possible rock glacier formation above ~1700 m		Palmentola et al. (1995)
	Bulgarian Mountains	Discontinuous and sporadic permafrost above 1950-2000 m, rock glaciers, block streams, patterned ground	OD, YD	King and Akerman (1993), Dobinski (2005), Gikov and Dimitrov (2011), Dimitrov and Gikov (2012), Kuhlemann et al. (2013)
	Mountains in Greece	Glacial cirques: possible rock glacier formation above ~2100 m (age uncertain, probably multiple generations of rock glacier in Greece with altitudinal ranges of 1330-2300 m)	age uncertain	Palmentola and Stamatopoulos (2004), Hughes et al. (2006a)
Holocene	Bulgarian Mountains	Discontinuous and sporadic permafrost, rock glaciers, block streams (?)		Gikov and Dimitrov (2011), Dimitrov and Gikov (2012), Kuhlemann et al. (2013)
	Mountains in Greece	Perennial snowfields, possible sporadic permafrost		Styllas et al. (2015)
LIA	Dinaric Alps*	Highest cirques: presence of small cirque glaciers above ~1850 m, greater abundance of snow patches		Hughes (2010, 2014)
	Bulgarian Mountains	Discontinuous and sporadic permafrost, rock glaciers, glacierets and small cirque glaciers, perennial snow patches		Grünewald et al. (2008), Gachev et al. (2016)
	Mountains in Greece	Perennial snowfields, greater abundance of snow patches. Sporadic permafrost. Possibly small cirque glaciers on Mt Olympus		Styllas et al. (2015)
Present-day	Dinaric Alps*	Karst depressions: cryo and ice caves above ~800 m Highest cirques: permanent firn/ice features above 1910 m Probable sporadic permafrost above 2300 m		Brown et al. (2001), Dobinski (2005), Kern et al. (2006), Mihevc (2008), Milivojević et al. (2008), Hughes (2009), Bočić et al. (2014), Košutnik et al. (2014), Gachev et al. (2016), Buzjak et al. (2016), Zupan Hajna (2016)
	Bulgarian Mountains	Sporadic permafrost, glacierets and small cirque glaciers, perennial snow patches		Grünewald et al. (2008), Grünewald and Scheithauer (2010), Nojarov (2012a,b), Gachev et al. (2016)
	Mountains in Greece	Perennial snowfields, probable sporadic permafrost above 2700 m		Styllas et al. (2015)

*Slovenia, Croatia, Bosnia and Herzegovina, Montenegro, Albania

Table 6. Altitudinal range of rock glaciers in the Balkan Peninsula.

Location	**Min altitude (m)**	**Max altitude (m)**	**Data source**
Pindus Mountains	1330	2300	Hughes et al. (2003), Palmentola and Stamatopoulos (2006)
Pirin Mountains	2090	2700	Dimitrov and Gikov (2012)
Rila Mountains	2140	2690	Gikov and Dimitrov (2011)
Prokletije Mountains	1700	2120	Palmentola et al. (1995)
Mount Korab	1480	2600	this paper
Šar Mountains	1750	2250	Kuhlemann et al. (2009)
Dinaric Alps (without Prokletije)	1650	2150	this paper

Table 7. Periglacial processes and landforms in the Anatolia Peninsula since the Last Glaciation until today.

Phase	Areas	Environment: landforms and processes	Chronology (ka BP)	References
Last Glaciation	Western Taurus Mts.	Glaciated summit plateaus above 2400 m: strong glacial and karst control on the intensity of periglacial activities Periglacial processes below LGM snowline (2000-2500 m)		Messerli (1967), Çiner (2004), Sarıkaya et al. (2008, 2014), Zahno et al. (2009), Çılğın (2015), Çiner et al. (2015), Köse et al. (2017), Sarıkaya and Çiner (2017), Sarıkaya et al. (2017)
	Central Taurus Mts.	Glaciated summit plateaus above 3000 m: strong glacial and karst control on the intensity of periglacial activities Periglacial processes below LGM snowline (2650-2700 m)		Zreda et al. (2011), Çiner and Sarıkaya (2017), Sarıkaya and Çiner (2017)
	Southeastern Taurus Mts.	Periglacial processes below LGM snowline (2100-2800 m)		Sarıkaya and Çiner (2015), Yeşilyurt et al. (2016)
	Eastern Black Sea Mts.	Periglacial processes below LGM snowline (2300-2500 m), relict rock glaciers		Erinç, (1952), Messerli (1967), Çiner, (2004), Akçar et al. (2007, 2008), Reber et al. (2014)
	Karçal (Lesser Caucasus Mts.)	Relict rock glaciers (2540-2430 m)		Dede et al. (2015, 2017)
Deglaciation	Western Taurus Mts.	Strong glacial and karst control on the intensity of periglacial activities		Zahno et al. (2009), Çiner et al. (2015), Sarıkaya et al. (2008, 2014), Sarıkaya and Çiner (2017)
	Central Taurus Mts.	Glaciated summit plateaus Strong glacial and karst control on the intensity of periglacial activities (rock falls, mass movements, rock glaciers, patterned grounds)		Çılğın (2015), Sarıkaya et al. (2008, 2014, 2017), Çiner et al. (2015), Zreda et al. (2011), Çiner and Sarıkaya (2017), Sarıkaya and Çiner (2017)
	Southeastern Taurus Mts.	Relict rock glaciers		Çiner (2003), Sarıkaya and Çiner (2015)
	Eastern Black Sea Mts.	Relict rock glaciers		Akçar et al. (2007, 2008), Reber et al. (2014)
	Karçal (Lesser Caucasus Mts.)	Relict rock glaciers (2540-2400 m), protalus lobes	15	Dede et al. (2015, 2017)
	Mercan	Relict rock glaciers		Bilgin (1972)
Holocene	Western Taurus Mts.	Non-glaciated summit plateaus: strong karst control on the intensity of periglacial activities		Arpat and Özgül (1972), Çiner et al. (1999, 2015), Çılğın (2015), Sarıkaya et al. (2008, 2014), Zahno et al. (2009), Sarıkaya and Çiner (2017)
	Central Taurus Mts.	Non-glaciated summit plateaus: strong karst control on the intensity of periglacial activities, rock glaciers (1800 m)	> 6	Arpat and Özgül (1972), Çiner et al. (1999, 2015, 2017), Zreda et al. (2011) Çiner and Sarıkaya (2017), Sarıkaya et al. (2017)
	Southeastern Taurus Mts.	Strong glacial control on the intensity of periglacial activities		İzbıkak (1951), Erinç (1953)
	Eastern Black Sea Mts.	Non-glaciated summit plateaus, rock glaciers		Akçar et al. (2007, 2008), Reber et al. (2014)
	Karçal (Lesser Caucasus Mts.)	Rock glaciers (2935-2730 m)		Dede et al. (2015, 2017)
	Ilgaz	Block flows, cyroplanation surfaces		Erinç et al. (1961)
	Erciyes	Early to late Holocene glacial advances implying periglacial conditions surrounding the glaciers		Sarıkaya et al. (2009)
	Ağrı (Ararat)	Debris on the slopes and blocky colluvium in the valley floor		Avcı (2007), Sarıkaya (2012), Azzoni et al. (2017)

LIA	Western Taurus Mts.	Rock glaciers		Çiner et al. (2015), Sarıkaya et al. (2014, 2017) Sarıkaya and Çiner (2017)
	Central Taurus Mts.	Rock debris on glaciers, rock glaciers (1800 m)	>6	Altın (2006), Gürgen et al. (2010), Çiner and Sarıkaya (2017), Sarıkaya et al. (2017)
	Eastern Black Sea Mts.	Rock glaciers, protalus lobes		Doğu et al. (1993), Akçar et al. (2007, 2008), Bayrakdar and Özdemir (2010), Reber et al. (2014)
	Karçal (Lesser Caucasus Mts.)	Rock glaciers (2935-2730 m), protalus lobes		Dede et al. (2015, 2017)
	Uludağ	Moraines in cirque areas with periglacial landforms		Erinç (1952), Birman (1968), Zahno et al. (2010), Akçar et al. (2014, 2015)
Present-day	Western Taurus Mts.	Non-glaciated summit plateaus; active karst influence on periglacial landforms, rock glaciers (2500-2800 m)		Dellanoy and Maire (1983), Sarıkaya et al. (2014), Sarıkaya and Çiner (2015)
	Central Taurus Mts.	Debris covered glaciers Cave ice at 3000 m Stone stripes at 3200 m		Bayarı et al. (2003), Klimchouk et al. (2006), Çalışkan et al. (2012)
	Southeastern Taurus Mts.	Rock glaciers		Doğu (2009)
	Eastern Black Sea Mts.	Stone circles (1800 m) and ovoid depressions (1900 m) Solifluction terraces, frost creep and mass movements (rock falls, talus, talus creeps, rock avalanches and rock flows) Garland soils, rock glaciers		de Planhol and Bilgin (1964), Bilgin (1969), Gürgen (2001), Akçar and Schlüchter (2005), Çiçek et al. (2006), Turoğlu (2009), Gorbunov (2012), Reber et al. (2014)
	Karçal (Lesser Caucasus Mts.)	Rock glaciers between 2935-2730 m Protalus lobes at ca. 2000 m		Dede et al. (2015, 2017), Çalışkan (2016)
	Kazdağ	Relict block streams (1350 m) Garland soils (1700 m)		Bilgin (1960)
	Uludağ	The best-described mountain concerning periglacial landforms Garland soils (1900-2300 m) Stone accumulations >2300 m		Erinç (1949, 1957), Zahno et al. (2010), Türkeş and Öztürk (2008, 2011), Öztürk (2012), Akçar et al. (2014, 2015)
	Ilgaz	Small stone circles and garlands		Erinç et al. (1961)
	Esence	Garlands and polygonal soils >2500 m		Akkan and Tuncel (1993)
	Mercan	Debris covered glaciers mistakenly interpreted as rock glaciers		Yeşilyurt and Doğan (2010)
	Mescid	Stone rings >2650 m		Atalay (1983)
	Erciyes	Rock glacier (2950 m)		Sarıkaya et al. (2003, 2009), Ünal and Sarıkaya (2013)
	Ağrı (Ararat)	Ice cap of which 1.82 km^2 is debris-covered >4000 m: a large recently deglaciated area 3000 to 4000 m; gravitational processes (mass wasting)		Sarıkaya (2012), Yavaşlı et al. (2015), Azzoni et al. 2017)

Table 8. Periglacial processes and landforms in northern Africa since the Last Glaciation until today.

Phase	Areas	Environment: landforms and processes	Chronology (ka BP)	References
Last Glaciation	Hoggar	Nivation forms (> 2400 m), frost action (< 2000 m)		Messerli (1973)
	Tibesti	Nivation forms (> 3000 m), frost action (< 2000 m)		Messerli (1973)
	Aurès Massif	Nivation forms (> 1800 m), stratified slope deposits (> 1800 m)		Tihay (1973)
	Algerian High Plateaux	Nivation (> 1500 m)		Marre and Quinif (1981)
	Algerian Tell	Nivation forms (> 1500 m), stratified slope deposits (> 1400 m), possible rock glaciers (> 1100 m)		Tihay (1973)
	Rif	Rock glaciers, perennial snowpatches		Mensching (1960)
	Middle Atlas	Possible rockglaciers (2100-2500 m), stone circles (?), solifluction features (?), stratified slope deposits		Dresch and Raynal (1953), Raynal et al. (1956), Awad (1963), this paper
	High Atlas	Stratified slope deposits		Chardon and Riser (1981)
Deglaciation	High Atlas	Possible rock glaciers (M'Goun Massif) (but possibly also in the Middle Atlas)		Wiche (1953)
Present-day	Hoggar	No active periglacial processes		Messerli (1973)
	Tibesti	No active periglacial processes		Messerli (1973)
	Aurès Massif	Frost action (2300 m), solifluction (1300-1800 m)		Tihay (1973), Ballais (1981)
	Algerian Tell	Nivation (1800 m), frost action (1500 m)		Tihay (1973)
	High Atlas	Possibly active talus rock glaciers/lobe (> 3700 m), possible permafrost (3800 m), frost shattering (> 2500 m), solifluction lobes (3900 m)		Chardon and Riser (1981), Hughes et al. (2011), Hughes (2017), Vieira et al. (2017)
	Middle Atlas	Frost shattering, terracettes/vegetation crescents, shallow solifluction (2700 m at Djebel Bou Naceur)		Dresch and Raynal (1952), Raynal et al. (1956), this paper