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L A N D F O R M E V O L U T I O N I N M O U N T A I N A R E A S

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THE LITTLE ICE AGE IN THE ALPS: ITS RECORD IN GLACIAL DEPOSITS AND ROCK GLACIER FORMATION

Abstract. Climatic fluctuations of glaciers of the Little Ice Age (LIA)-rank used to show comparable magnitudes during the Holocene. That is why the extent of maximal LIA moraines in the Alps attains similar position as those of the previous LIA-type Holocene advances. Consequently, the LIA moraines represent a massive and frequently multi-ridge depositional system. Certain differences in reaction to climatic changes and in moraine development result from the glacier size, percentage of debris-covered ice surfaces, and topography of alimentary areas. Such differences are particularly spectacular regarding "pure ice" and debris-covered glaciers. Another system represent rock glaciers which, however, frequently accompany regular glaciers. Some rock glaciers originated due to one-time transformation of a "glacial" glacier during its recession, while others formed owing to long-term accumulation of rock debris within glacial cirques during successive oscillations of small cirque glaciers in the Holocene, given suitable topographic and topoclimatic conditions. The relationship between the ELA and altitude of cirque bottom and/or debris supply from surrounding walls is of key importance in this case.

Key words: Little Ice Age, moraines, rock glaciers, ELA, climatic change, Alps

INTRODUCTION

The Little Ice Age (LIA) was a global climatic deterioration at a scale of hundreds of years, which has strongly influenced on natural environment and human civilization. In polar and high mountain areas glaciers advanced and achieved their maximum postglacial positions during this period. The strongest impact of cooling was noticeable in the North Atlantic region: in northern latitudes of Europe, Asia and North America (Hunt 2006).

The term "Little Ice Age" was first used by F. Matthes (1939; after Luckman 2004) to describe the late Holocene (last 4,000 years) glacier re-advances in the Sierra Nevada Mountains, following the Hipsithermal warm climate of the early Holocene. At present, this term is reserved for the last cool episode in the Holocene, which took place in the last millennium. The older meaning of Little Ice Age is termed today as Neoglacial (Luckman 2004). It is commonly ac-

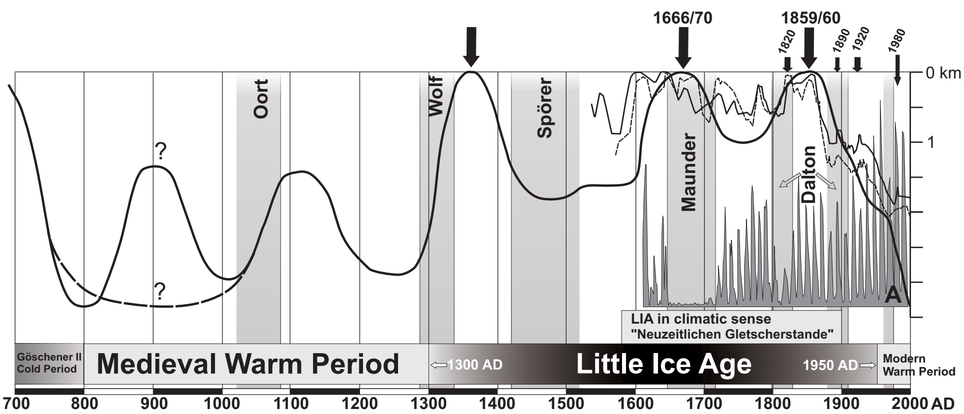

Fig. 1. Glaciers fluctuations in the Alps in the last millennium. Thick line — Grosser Aletsch glacier fluctuations, thin line — Lower Grindelwald Glacier (Holzhauser and Zumbühl 1996), dashed line — Mer de Glace (Nussbaumer et al., in press). Note a longer response time in case of Grosser Aletsch Glacier; minor climatic changes are not noticeable in front oscillation. The LIA time period according to J. A. Matthews and K. R. Briffa (2005). A — sunspot cycle derived from the number of sunspot group (Beer et al. 2000). Period of reduced solar activity (Dalton, Maunder, Spörer; Büntegen et al. 2006) are shaded. Thick arrows indicate main maxima of the Grosser Aletsch glacier front, tiny arrows indicate minor oscillations of glaciers during the last 200 years

cepted that the LIA occurred between the 16th through 19th centuries (Mann 2002; Grove 2004). Frequently, the termination of LIA is assigned to the years1850–1860, i.e. the time of maximum extent of glaciers and the beginning of their retreat. This view, however, does not seem to be justified, because both climatic conditions and glacier sizes at that time were far from typical of climatic amelioration. According to J. A. Matthews and K. R. Briffa (2005), it seems necessary to consider two different time spans: 1300–1950 AD in glaciological sense, and 1570–1900 AD in climatic sense (Fig. 1). This chronology is based mainly on research on Alpine glacier fluctuations and European palaeoclimate, although the 14th century initiation of LIA around the polar North Atlantic region is also recorded (see Grove 2001). In glaciological sense, the duration of LIA is comparable to a period when glaciers were larger than before or after that time. For instance, the Great Aletsch Glacier had a greater extent in the years 1280–1935/40 AD than before and after this period (Holzhauser 1983).

The LIA-related glacial landforms and deposits together with associated periglacial landforms in the Alps represent perfectly preserved evidence of a cold geological epoch. The cooling took place in historic times; therefore, it was observed and registered by man; an unique opportunity in Earth history. Based on the record of glacier fluctuations and climatic changes throughout the LIA, as well as related landforms and sediments, one can conclude about older cooling phases, the geomorphic and sedimentary evidence of which are not so well preserved.

In Southern Europe outside the Alps, there are only few areas where the LIA left its traces in glacial sediments. Well developed LIA moraines are to be found in those areas which are presently glaciated or bear remnant glaciers, namely: in the Pyrenées (Cia et al. 2005), Sierra Nevada in Spain, or Central Apennines with Ghiacciaio del Calderone (D'Orefice et al. 2000). In other mountain massifs the LIA glacial landforms are either poorly preserved or small, i.e. in Corsica (Kuhlemann et al. 2005), Kaçkar Mountain range of northeastern Anatolia (Akçar et al. 2007), and Mount Olympus in Greece (Smith et al. 1997). Still other mountains, like the Carpathians, lack such landforms at all.

LIA AND "LIA TYPE EVENTS" IN THE HOLOCENE

Throughout the Holocene, Alpine glaciers fluctuated in a relatively narrow range; the snow line and timberline have never exceeded a certain value of amplitude, i.e. ca 200 m (Patzelt and Bortenschlager 1973). During the LIA, glacier advances were most extensive in the entire Holocene and definitely most extensive in the Neoglacial period. That is why geomorphic records of the Postglacial oscillations, older than LIA, were mostly destroyed by LIA advances or overprinted by new moraine deposits. However, the glacier front positions during the LIA maxima did not surpass significantly the previous Holocene maxima. In a few areas, for instance, where palaeoclimatic and topographic situation was favourable enough, moraines left by the early Holocene advances are preserved. These moraines are situated farther down the valley or quite close and parallel to the LIA moraines, and represent the following stadials: Schalten, Kartell (Preboreal — Patzelt and Bortenschlager 1973; Ivy-Ochs et al. 2006), Venediger (Boreal — Patzelt and Bortenschlager 1973), and Kromer (older Atlanticum — Kerschner et al. 2006).

Despite the lack of geomorphic evidence, such as moraines, knowledge of glacial history of the entire Holocene is increasing owing to dendrochronological, sedimentological, pedological and palynological investigations and radiocarbon dating performed on glacier forefields (cf. Patzelt and Bortenschlager 1973; Nicolussi and Patzelt 2000; Hormes et al. 2001; Joerin et al. 2006). The results of such studies are not always compatible, particularly regarding the number, duration and intensity of cold Holocene episodes. Some results are even contradictory; probably due to the use of different methodologies, but also due to genuine, regional palaeoclimatic and topoclimatic differences (cf. Nicolussi and Patzelt 2000). It should also be underlined that the above results usually determine the time and environmental conditions of warm intervals, based on which cold episodes can be detected.

During the Holocene climatic optimum there was no prolonged glacier-free period in the Alps; even in the Atlantic phase glacier advances of the LIA rank did occur. The late Holocene is typified by increasing frequency and duration of such episodes. According to U. E. Joerin et al. (2006), during the middle Holocene Alpine glaciers advances were associated with short, ca 200 years long, cold intervals which were separated by 500 years long warmer intervals. These proportions became reversed in the late Holocene. The beginning of intensive glaciation of the Alps in the Neoglacial is usually dated to 3.5 ka (Gamper and Suter 1982). Deline and Orombelli (2005) place this event in the interval of 4.6–5.6 ka, based on evidence of first stages of glacial accumulation in the Miage Glacier ablation complex in the Mont Blanc massif. In the late Holocene, after 3.5 ka, the pattern of glacier fluctuations in the Alps was more equivocal. Detailed data pertaining to fluctuations of individual glaciers are available for this period, obtained mainly due to dendrochronological methods (e.g., Grosser Aletsch Glacier, Gorner Glacier, Lower Grindelwald Glacier — Holzhauser and Zumbühl 1996; Holzhauser et al. 2005). These data point to at least four glacier advances of the LIA rank in that period: Löbben (1300–1900 BC), Göschener I (400–1000 BC), Göschener II (350–800 AD) and, finally, LIA (1300–1950 AD; Patzelt and Bortenschlager 1973; Gamper and Suter 1982; Holzhauser and Zumbühl 1996; Deline and Orombelli 2005).

GLACIER FLUCTUATIONS IN THE ALPS IN THE LAST MILLENNIUM

The Little Ice Age was preceded by the Medieval Warm Period, lasting from 800 to 1300 AD. In this warm interval, glaciers fluctuated as well, although being considerably smaller than during the LIA (High Middle Ages advances; Holzhauser and Zumbühl 1996; Holzhauser et al. 2005). Alpine glaciers advanced in the LIA in three principal stages (Fig. 1). The first advance took place around 1350 AD and terminated the Medieval Warm Period. The second LIA maximum occurred in the middle of the 17th century (ca 1650 AD), whereas the last one marked its appearance in the middle of the 19th century (ca 1850 AD). The first maximum was separated from the second one by a decisively longer period, during which glaciers retreated farther compared to the warming between the second and third maximum (Holzhauser 1983; Holzhauser et al. 2005).

Despite the fact that the cold maximum in Europe took place in the 17th century, during the so-called Maunder minimum (1645–1715 AD; Eddy 1976; Fig. 1), in most cases the 19th century advance was the most extensive one. Other cases are, for instance, represented by fluctuations of Mer de Glace and Lower Grindelwald glacier, which had had their maximum LIA extent in the first half of the 17th century (Nussbaumer et al., in press; see Fig. 1). The Alpine glaciers attained the 19th century maximum around 1850–1860 AD (Holzhauser 1982). In some areas of the Alps, like the Mont Blanc region, in turn, this event took place during an advance of the 1820s; therefore, the 1850 moraines are placed on the inner side of the 1820s moraines (cf. Aeschlimann 1983). The 1820 advance took its appearance in the remaining parts of the Alps; however, the stronger transgression of the 1850s surpassed the older one (Holzhauser and Zumbühl 1996; see Fig. 1 — fluctuations of Mer de Glace and Lower Grindelwald glacier).

In the forefield of some of Alpine glaciers, morainic ridges dated to the 17th and even 14th centuries did preserve (Patzelt and Bortenschlager 1973; Holzhauser 1984). The chronology of glacier advances in the 17th and 19th centuries and associated moraines are recognized considerably better compared to the first maximum, not only due to well preserved geological and geomorphic records, but also owing to greater number of data obtained from direct observations, like written records, drawings, paintings, photographs, and maps (Holzhauser 1983, 1984; Zumbühl et al. 1983; Zängl and Hamberger 2004; Brunner 2006).

Glacier fluctuations in the Alps between the 17th and 19th centuries used to be called in German literature as "*Neuzeitlichen Gletscherstande*" ("*Neuzeitlichen Gletscherschwankungen*" — modern age moraines; Holzhauser 1982), and the area comprised between the recent glaciers and the 17th and 19th centuries ones is precisely defined as the "*Gletscherforfeld*" — glacier forefield (Kinzl 1932; Holzhauser 1982). This zone is clearly different from the surrounding area in respect to geomorphic setting, pedological characteristics, floristic succession, and the degree of weathering. Moreover, lichenometric method can easily be applied in this zone, due to relatively young age of deposition. A few systems of moraines younger than 1850 AD are situated inside the area in question. These moraines, dated to 1890, 1920, and 1980 AD, were formed during short oscillations or episodes of glacier standstill, which broke the general trend of glacier retreat persisting until now (Patzelt 1985; Zemp et al. 2007). These oscillations were well recorded owing to direct measurements of ice volume, area and glacier length changes, which have been monitored in the Alps during the past 120 years (Zemp 2006). Moreover, constant monitoring of meteorological conditions together with estimation of ice volume balance make it possible to correlate in detail climatic and glacier geometry changes, providing a base for prediction of the future Alpine glaciation (cf. Zemp 2006; Zemp et al. 2006; Paul et al. 2007).

It is worth noting, however, that individual glaciers in the Alps could oscillate asynchronously, despite good concordance of the first-order oscillations . This results both from palaeoclimatic differences among individual Alpine regions, and from topographic conditions and variable sizes of Alpine glaciers, controlling their mode and time of reaction to climatic changes. The reaction time of glacier front positions to climatic changes is not simple and linear. It primarily depends on the size of glaciers, but also on the climate of a number of previous years. For the most of medium size Alpine glaciers it is about 14–18 years, whereas in case of larger ones it could be about 40 years (Hormes et al. 2006). According to T. Johanesson et al. (1989), it takes 15–60 years for changes in mass in the accumulation zone to reach the snout of maritime temperate glaciers. However, the time in which glacier geometry reaches new equilibrium after abrupt climatic change (response time) is much longer; for Grosser Aletsch Glacier it is in the order of 50 to 100 years (Haeberli and Holzhauser 2003; cf. Fig. 1).

From 1850 to the 1970s, 5,150 glaciers from the entire Alpine mountain chain lost about 35% of their areas, and almost 50% by 2000. In 1850, there was about 4,470 km^2 occupied by glaciers, while in 2000 only 2,270 km^2 remained (Zemp et al. 2007). Glacier recession is a result of the increased altitude of the equilibrium line of mass balance (ELA — equilibrium line altitude; see Gross et al. 1977; Benn and Lehmkuhl 2000). It is estimated that ELA shifted upwards between 1850 and 1973 AD by 114 m. This was primary triggered by a rise in temperature by about 0.6°C, which was partially compensated (about one fourth) by an increase in precipitation by ca 10% (Greene and Broecker 1999). Mean elevations of recent glaciers, which roughly equate to the equilibrium line altitude (ELA, here assumed as the steady-state ELA) is, for the European Alps, 2,945 m a.s.l. ± 214 m (Hoelzle et al. 2007). The altitude of equilibrium line during LIA maximum in the Alps was strongly variable depending on geographic location and slope aspect. Taking into account that ELA of individual glaciers is situated above their lateral moraines (cf. the MELM method — maximum elevation of lateral moraines; (see Benn and Lehmkuhl 2000), it was possible to estimate the altitude of eqilibrium line during the 1850's maximum in the central part of Eastern Alps at ca 2,600–3,100 m a.s.l., (Gross et al. 1977). In the northern fringe of the Alps, in Karwendel mountains, ELA was placed even lower, at ca 2,300 m a.s.l. (Kerschner and Ivy-Ochs 2007). Regional differences in ELA altitudes, known also from older, Late Glacial glacier advances like Egesen (Younger Dryas), result mainly from regional differentiation in precipitation, since differences in temperatures throughout the Alps are not so great (Kerschner et al. 2000).

Since the last maximum of 1850 AD, fronts of the following large glaciers have retreated considerably: Grosser Aletsch — about 3.4 km, Gorner (Alps of Valais) ca 2.6 km, lower Grindelwald (Bernese Oberland) ca 2.0 km, and Pasterze (Hohe Tauern) ca 1.8 km. It is worth noting, however, that during the Bronze Age climatic optimum (ca 1200 BC), the Grosser Aletsch glacier was shorter by ca1 km, and during the Roman climatic optimum (400 BC–400 AD) only a little bit shorter in respect to its present day length. Not all results appear to support a view that the present-day position of Alpine glaciers is far from their Holocene minimal extent. Accoording to Grosjean et al. (2007), the present-day warming is unprecedented one since at least 5,000 years, because melting ice on highly elevated Alpine passes discloses artefacts from the Neolithic, Bronze Age, Roman Period and Medieval Ages. These artefacts, due to their fragile nature, could not have been preserved when exposed during earlier warming periods. One should also bear in mind that the present-day extent of glaciers is very far from balance with climatic conditions, as shown by the results of climatic-glaciological modelling (Greene and Broecker 1999). It means that the glaciers have not managed to apply their geometry to the existing climatic conditions. The rate of glacier retreat in the last decades has been faster than long-term average, particularly after 1985 AD. This fact, together with dramatic changes in alimentary areas of glaciers points to massive downwasting rather than dynamic glacier response to a changed climate (Paul et al. 2004). For instance, in one, extremely warm and dry year 2003, the Alpine glaciers lost ca 5–10% of their previous volume (Zemp et al. 2005).

CLIMATIC BACKGROUND

Glaciers are extremely sensitive to climatic changes, their fluctuations depending on temperatures and precipitation totals as well as seasonal distribution of these factors (Pfister 1980, 1984; Denton et al. 2005). The mass balance of glaciers is particularly sensitive to summer temperatures and winter precipitation. Also important are summer snowfalls upon glaciers. Fresh snowfall during the ablation season can increase albedo and effectively protect pure ice against solar radiation, diminishing melting rate of the former.

Climatic fluctuations of the order of tens to hundreds of years during the last millennium and throughout most of the Holocene, are mainly induced by external factors related to solar activity (Eddy 1976; Stuiver et al. 1998; Mayewski et al. 2004). The North Atlantic Oscillation (NAO) and its interaction with the atmosphere, as well as inner, random climatic oscillations (Schulz and Paul 2002) are also very important.

The minima of solar activity reconstructed from atmospheric ^{14}C residual series (based on tree ring records; cf. Stuiver et al. 1998) are in good accordance with glacial advances. This relationship has also been confirmed by direct observations in the last decades. Some discrepancies are mainly due to long-term glacier response to climatic changes (Holzhauser et al. 2005) and stochastic processes within the climatic system (Hunt 2006), which do not allow directly correlate solar activity with glacier behaviour. Solar activity varies in the 11- (Schwabe cycle), 80- (Gleissberg cycle), and 200- year periodicities (Schove 1983), as is directly demonstrated in the sunspot number and, in consequence, changes in the solar constant.

Dendrochronological data point to a sharp cooling around 1570/80 AD over a large area of middle and northern Europe. This region has its southern boundary in the Alps (Briffa et al. 1999). The cooling, initiating LIA in a climatic sense as well as modern age glacier fluctuations, is associated with the Maunder and Dalton solar minima activity (see Fig. 1). A distinct advance of Alpine glaciers at the end of the 16th century (the second LIA maximum) was a result of predominantly low winter and summer temperatures, and summer snow falls in high Alpine area, while the last LIA maximum (around 1850 AD) was mainly triggered by low winter temperature and high amount of winter precipitation; lowering of summer temperatures was slightly less pronounced than that in the earlier period. During Maunder minimum (1645–1715 AD), despite low temperatures, no significant advance was recorded (Fig. 1), probably due to continental, very dry winters (Pfister 1984).

The last two distinct advances in the 19th century (in the 1820s and 1850s) resulted from cold and enormously wet conditions (Pfister 1980, 1984), although a role of volcanic activity can not be neglected. Big Plinian eruptions like, for example, those of Tambora 1815, Coseguina 1835, and Krakatoa 1883 (compare advances in Fig. 1), produced a large amount of dust propagated globally in the atmosphere, drastically influencing the climate by substantial lowering of temperature (for example, 1816 "the year without the summer", 1838 "the coldest year overall in Europe") and increasing cloudiness. Nevertheless, volcanically-induced climatic deteriorations are rather short (mostly 2–3 years; Robock 2000). It is hypothesized that the 20th century climatic amelioration was amplified by the dearth of volcanic eruptions between 1915 and 1960 AD. In addition, dust layers in ice core records point to a relatively lower number of volcanic eruptions between 1000 and 1500 AD, compared to the period of 1500–2000 AD (Robock 2000; Jones and Mann 2004, and references therein).

The overall loss of Alpine glaciers after the 1850 maximum is coincident with the positive (warm/wet) NAO index, contrary to the Maunder Minimum when minimum solar activity together with negative (cold/dry) NAO index produced cold and severe climate. According to M. Schulz and A. Paul (2002), in the earlier part of the Holocene, between 8.5 and 3 ka, the climate system showed a regular, ca 900 years long, periodicity related to the North Atlantic thermohaline circulation. The quoted authors maintain that such climatic fluctuations resulted from an oscillation within the ocean-climate system, which was initiated by external cooling force in 8.2 ka cool events (outburst of Lake Agassiz). Later, after 3 ka, this 900 years climatic periodicity was not visible, but it is worth noting that in the late Holocene, after Hypsithermal warm interval (~9–5.7 ka BP), increasing frequency and intensity of the LIA-type events (Neoglacial) could have resulted from progressive decrease in solar radiation and summer temperature lowering due to external Earth's orbital forcing (Jones and Mann 2004; Matthews and Briffa 2005). Bond et al. (1999) concluded that the North Atlantic circulation and Holocene climate tended to oscillate in a 1–2 ka, Dansgard/Oeschger-like mode, what means that the LIA and LIA-type events originated like an abrupt full-glacial cold period events, although in the mild interglacial climate mode (Björck et al. 2001, and references therein).

Despite common belief that the LIA cooling marked its appearance first of all in the Northern Hemisphere, one should bear in mind that this event was strongly differentiated both in time and space. For instance, glacier advances in the Alps are asynchronous compared to those of Scandinavia and North America (for further discussion see Matthews and Briffa 2005).

Moraines of the maximum of LIA represent one of the best developed and clearly marked morainic systems in the Alps, occurring between <1,500 and ca 3,000 m a.s.l. The lowermost extent refers to the moraines of large glaciers, like the Grosser Aletschgletscher or Pasterze Gletscher, which during the LIA reached the upper timber line. This level in the Alps is at present at 2,100–2,300 m a.s.l.

The LIA moraines developed in front of the middle-size Alpine glaciers present a characteristic picture (Fig. 2). The lateral moraines, depending on suitable morphology of host valleys, attain a few tens to more than 100 m of height (Maisch et al. 2000), and are typified by extremely sharp profile. The frontal moraines, in turn, are usually small-size and frequently remodelled by fluvioglacial waters. There also occur, much more frequently than in lateral moraine zone, multiple ridges owing to frequent fluctuations of glacier fronts and strong erosion of proglacial waters, unfavourable to the formation of high and massive moraines. In the upper ablation zone, in turn, where the glacier maximum extent is more stable, massive lateral moraines can easily be deposited.

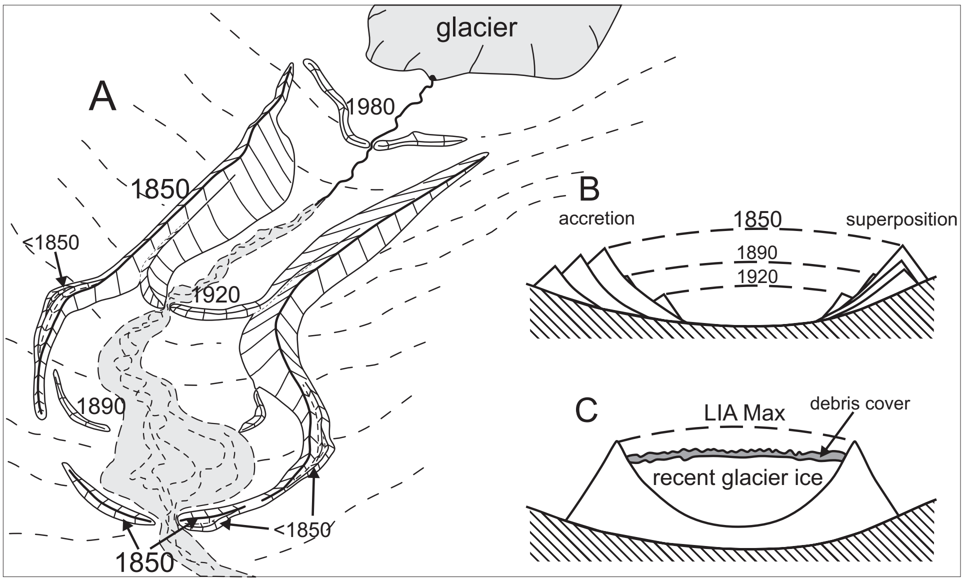

Fig. 2. A — Scheme of typical moraine formation in the forefield of a medium-size Alpine glacier, on valley bottom not affected by rough topography. Note massive (ca tens of metres high) lateral moraines and relatively small and multi-walled front moraines, partially destroyed by outwash fans. Glacier front position is usually more sensitive to minor climatic fluctuations than glacier surface in the higher part of the ablation zone, where a lateral moraine is deposited. Numbers indicate time of advances, "<1850", could correspond to the 17th century, late Medieval or early Holocene advances. B — Scheme of LIA moraine depositional structure; two modes of moraine building — by accretion and superposition. C — Scheme of recent debris-covered glacier; note strong lateral moraine and moraine accretion on the sole of glacier which rises the glacier above valley bottom

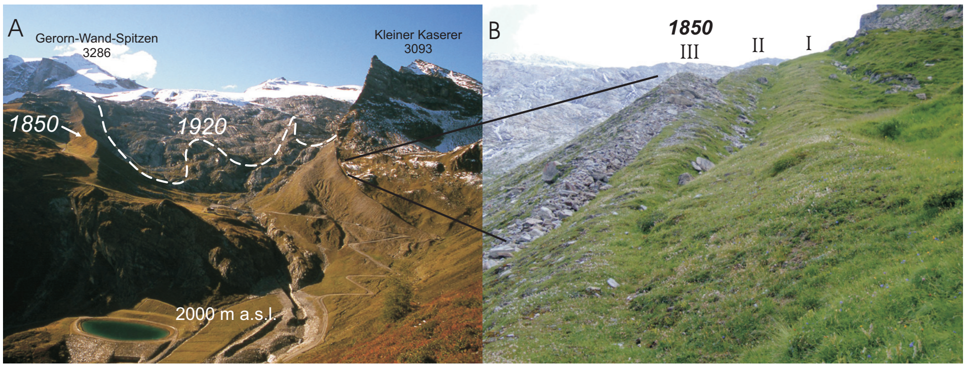

Fig. 3. A — Moraines on the forefield of the Hintertux Gletscher (Tuxer Ferner) in Zillertal Alps. Glacier front during LIA maximal advance (1850) was 1400 m down the valley from recent front position on the rock threshold. Frontal moraine is only preserved below left lateral moraine (right on the picture); note a three-folded morainic sequence. The right lateral moraine (on the left) is 35 m high, 1,300 m long, and 1.46 mln m^3 in volume. The 1920 recessional position is reconstructed mainly on the base of old photos; moraines are only partially preserved. B — accretion depositional structure of the left lateral moraine. Older moraine ridges (I and II) are more affected by denudation processes

Moraines marking the maximum extent of the LIA (1850 AD), occupy comparable positions to those of the previous LIA maxima, as well as glacier advances in the entire Holocene. At places, the sedimentological and geomorphic record of such moraines perfectly reflects their complex and long-term process of deposition. In certain circumstances, both along frontal and lateral moraines, there are preserved moraines of glacier advances preceding 1850 AD, occupying a more outer position. Successive moraines became deposited close to the older ones, forming multi-ridge moraines (Figs. 2B, 3). In most cases, morainic material of successive maxima used to upbuild previously deposited ridges, forming a single ridge. In the inner part of such moraines, frequently strongly eroded due to slopewash and landsliding, one can observe layering, in which morainic material of individual advances is separated by fossil soil horizons (cf. Holzhauser and Zumbühl 1996).

Among moraines younger than 1850 AD, the most clearly marked are those of the 1920s, although none of them is comparable, as far as the material volume is concerned, to the maximal moraines. Moraines dated to the end the 19^{th} century (1890s) are marked less clearly, as are those of the 1980s. These moraines are well developed in case of debris-abundant glaciers, while below debris-free glaciers such an oscillation left no moraines at all. Moreover, the lack of recessional moraines below large glaciers can result from their prolonged reaction and absence of more important oscillations of the glacier front.

A certain regularity exists regarding the presence of debris within glaciers: following diminishing glacier size and its retreat towards cirques and rock walls, the glacier area diminishes in respect to the alimentary area. Glacier melting and intensive physical weathering of rock walls produce a wealth of ablation debris and that supplied by rockfalls. In such a case, even minor oscillations of the glacier front leave a certain amount of debris in the form of moraines. When topographic and lithological conditions do not allow for intensive debris supply, even marginal glaciation leaves no recessional moraines at all.

Moraines of those glaciers which during the LIA represented small cirque glaciers are usually massive ones and accompanied by periglacial landforms, like rock glaciers. Such moraines resemble high debris ramps (Fig. 4) (cf. M a i s c h 1981, p. 102). Great thickness and/or volume of such moraines results from the fact that debris supplied from cirque walls during successive Holocene advances became accumulated on a relatively small area, on cirque thresholds (M a i s c h 1981). In case of larger glaciers, the amount of debris derived from the surrounding rock walls was not considerably larger, although it was deposited on a larger area. Moraines of small and remnant glaciers are usually built up of angular and poorly rounded debris, resulting from a short, usually supraglacial, transport route.

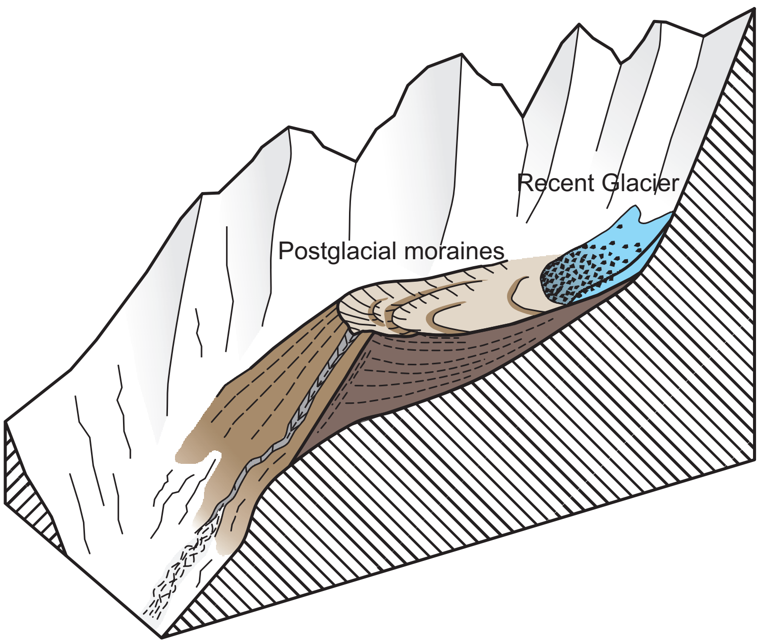

Fig. 4. Scheme of morainic depositional complex of a small Alpine cirque glacier. In such topoclimatic conditions, glacier deposits were accumulated during a number of narrow-range fluctuations in the entire Holocene, and produced huge latero-frontal moraine on a cirque threshold

Given suitable topographic conditions, even large glaciers can become covered with debris. A typical place of occurrence of debris-covered glaciers are mountain chains of Asia, such as Himalaya and Karakoram, where, due to young high relative relief, intense accumulation of supraglacial debris supplied by avalanches can occur. The debris-covered glaciers are characterised by well-marked moraines with slopes of the angle of repose and a few metres thick debris mantle on ice surface in the ablation zone (Fig. 2C). This cover masks glacier ice significantly and influences surface energy fluxes and discharge of meltwater, leading to a reduction of the ablation rate and producing an alteration of response of this kind of glaciers to climatic changes (Dagata and Zanutta 2005). The abundance of supraglacial debris helps to produce complex lateral-frontal moraines that may record one or more glacial cycles (Benn and Owen 2002).

The debris-covered glaciers are not common in the Alps. The two best known, Miage and Brenva glaciers, are situated in the western Italian Alps, on the southern side of the Mont Blanc massif. They gained their debris cover during one large rock fall in 1920 AD. After that, they were advancing till 1940 AD, and finally achieved their LIA maximum starting from 1818 AD. Before the rock fall event they behaved like other "clean" glaciers on the northern side of Mont Blanc (Dagata and Zanutta 2005). On the contrary, like all the remaining glaciers in the Alps, the debris-covered glaciers on the southern slope of Mont Blanc in the second half of the 20th century (1959–1997), revealed a net increase in volume and thickness (Dagata and Zanutta 2005). These glaciers lean on the LIA maximum moraines (cf. Fig. 2C). In such a kind of glaciers, climatic induced mass balance change in the accumulation zone does not imprint on the glacier front position, but is rather conducted by kinematic wave, during which ablation zone resulted in a rise of glacier surfaces (Pelfini et al. 2007).

FORMATION OF ROCK GLACIERS

Rock glaciers were originally defined as landforms produced due to creep of alpine permafrost (Haeberli 1985; Barsch 1987, 1988), although both origin and classification of such landforms have been a matter of animated debate. A widely known controversy exists between advocates of periglacial origin of rock glaciers (Haeberli 1985; Barsch 1987, 1988), and those who favour glacial origin ("pure ice" glaciers becoming successively covered with rock debris; cf. Johnson 1987; Ackert 1998; Whalley and Martin 1992).

Rock glaciers occur in the periglacial belt below ELA. The vertical extent of this belt depends on the degree of climatic continentalism (Haeberli 1985; Barsch 1988). Formation of rock glaciers requires specific climatic conditions and debris delivery; therefore, the distribution of such landforms is controlled by local topographic, topoclimatic, and lithologic conditions (Haeberli 1985; Barsch 1988; Humlum 1998; Ikeda and Matsuoka 2006).

These landforms are ubiquitous in the Alps, and their number can attain a few thousand. Active rock glaciers accompany moraines and "glacial" glaciers of Holocene (LIA) age. Huge landforms attain 1.5 km in length, ca 200–300 m in width, and a few tens of metres of thickness. There occur prevailingly two genetic types of rock glaciers *sensu* D. Barsch (1988): debris rock glaciers —

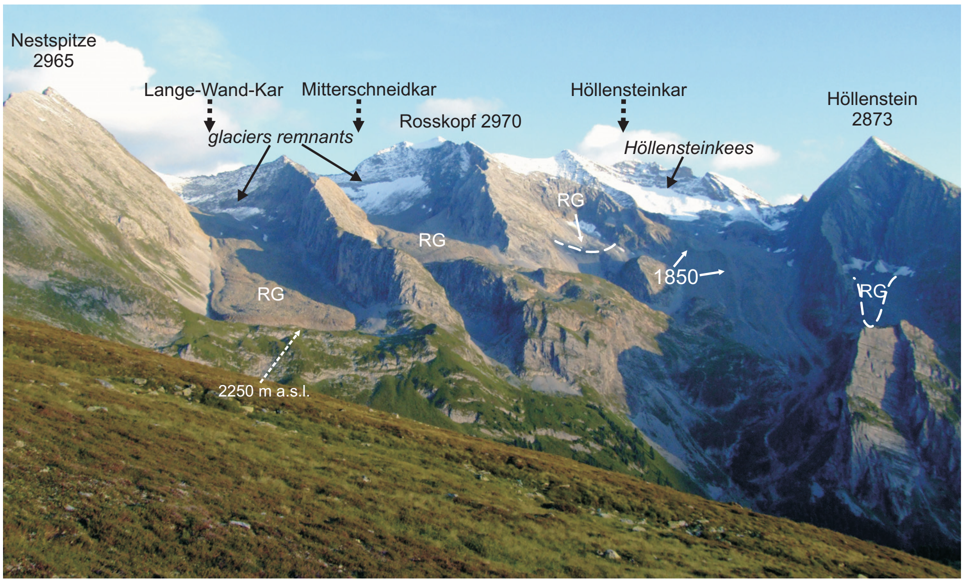

Fig. 5. Active rock glaciers and LIA moraines on the northern side of the Tuxer Hauptkamm (Zillertal Alps, Austria). In the same location and aspect, note coexistence between these two different modes of debris transport and deposition. Both landforms originated in the Holocene; the Lateglacial moraines (Egesen stadial — Younger Dryas) are situated down the valley below bare rock thresholds. In this area rock glaciers dominate, as cirque bottoms are substantially below the LIA snow line (2,650–2,700 m a.s.l.). Exception is Höllensteinkar, where the cirque floor near head walls reaches elevation permitting glaciation during the LIA. Moreover, this cirque is also more oval than other neighbouring cirques, which favour rather snow than debris accumulation. In the Lange-Wand-Kar and Mitterschneidkar rock glaciers rooting zone, below small glaciers, spoon-shaped depressions are situated (cf. Fig. 6). 1850 — huge LIA latero-frontal moraine in Höllensteinkar, RG — rock glaciers

associated with moraines and "glacial" glaciers (Fig. 5), and talus rock glaciers — generated from talus fans.

The best conditions for rock glacier formation are provided by dry, continental climate. Nevertherless, these landforms occur in all major mountains systems of the Earth (Barsch 1988). They frequently accompany regular glaciers, particularly in partly glaciated areas, occupying the locations which are more dry and prone to debris delivery (Humlum 1998). The key role in formation of the rock glaciers plays the proportion between snow accumulation and the input of debris (Humlum 1998). In other words, under small delivery of snow and abundant debris influx, rock glaciers are formed, while under reciprocal conditions pure ice glaciers tend to be produced. The balance is a delicate one, and can easily change both in time and space (climatic changes, topographic differences). An important role in rock glaciers distribution plays the relationship between cirques bottom and ELA (see Fig. 6), as well as the number, height and lithology of rock walls surrounding the cirque (Ikeda and Matsuoka 2006). When the entire cirque bottom is situated well below ELA, it will provide suitable conditions for rock glacier formation.

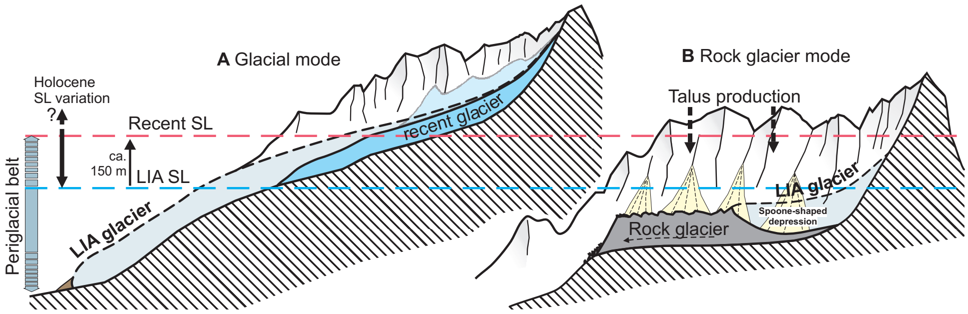

Fig. 6. Schematic relation between cirque bottom and snow line and its consequences in glacial/rock glacier formation. A — glacial mode: glaciers originate where topographic conditions allow for accumulating snow above the snow line. B — rock glacier mode: rock glaciers form where most of the cirque bottom is substantially below the snow line and/or sufficient debris load from head walls is available. SL — snow line, LIA SL — snow line during the Little Ice Age maximum (Holocene minimum elevation)

Small cirque glaciers are frequently being transformed into rock glaciers, particularly during their melting phase. First, owing to debris covering of the "pure" glacier, a debris-covered glacier originates, then, the prograding process leads to the formation of an ice-cored rock glacier (a glacigenic rock glacier — cf. Ackert 1998; Whalley and Palmer 1998; Berger et al. 2004). During individual advances, "glacial" glaciers can advance upon rock glaciers (alpine permafrost) (Haeberli 1979; Whalley 1979; Whalley and Palmer 1998). According to D. Barsch (1988), debris supplied by glaciers during their advances can supply rock glaciers of the debris rock glacier-type, and farther transport of the material proceeds due to permafrost creeping (Fig 6B).

During suitable topoclimatic conditions, frequently recurring glacier oscillations in the late Holocene led to formation of large, tongue-shaped rock glaciers. In the root zone of such rock glaciers, spoon-shaped depressions left after melt-out "glacial" glaciers are frequently preserved (Fig. 6B) (Barsch 1988; Avian et al. 2005). These depressions used to be filled with ice during the Holocene maxima, recently in the 1850s. Other landforms indicating glacial advances and occurring in the root zones of rock glaciers include: moraine walls, lodgement tills (Berger et al. 2004; Avian et al. 2005), flutes, fresh glacial striations (Whalley and Palmer 1998), and push moraines (Haeberli 1979). Spoon-shaped depressions can also be formed in the rooting zone of glacial originated rock glaciers (ice-cored rock glaciers) as a result of melting of debris-free part of the "glacial" ice body below head walls (cf. Berger et al. 2004).

The age of active Alpine rock glaciers does not exceed 10 ka (Barsch 1988), as shown by their position versus the youngest, Late Glacial, moraines (Egesen — Younger Dryas). Most frequently, however, their age is estimated at 3 to 4 ka (Neoglacial; cf. Avian et al. 2005). The debris rock glaciers distinguished by D. Barsch (1988) are of such an age, whereas ice-cored rock glaciers could have originated due to one-time transformation of “glacial” glaciers, mostly during a recession postdating 1850 AD (Whalley and Palmer 1998; Berger et al. 2004).

SUMMARY AND CONCLUSIONS

The climatic and LIA-rank glacier fluctuations occurred throughout the Holocene every 1–2 ka; in the late Holocene this regular periodicity was so regular. The LIA (1300–1950 AD; Matthews and Biffa 2005) represents the youngest cold period in this series of fluctuations. Climatic conditions in the Alps during LIA were comparable, or slightly more glacier-friendly, to the earlier Holocene LIA-type events. Changes in altitude of climatic zones and different ecosystems (timber line, periglacial belt, snow line) in the Holocene were relatively small, of the order of ca 200 m, compared to the Late Glacial. These changes resulted in i.a., fluctuations of glacier fronts from a few hundred metres to more than 3 km, depending on glacier size. The present-day warming reveals that glacier-occupied surfaces in the Alps can diminish by more than half between cold and warm periods. The LIA-type events were not only colder by ca 1–2°C than the intervening warmings, but also relatively humid (particularly owing to high winter precipitation), providing suitable conditions for glacier growth and high lake level stands.

The most important role in shaping Holocene climate is played by solar activity, together with its interactions with the Atlantic circulation and internal, random climatic fluctuations. Increasing intensity of cool periods in the Late Holocene (Neoglacial) could have been induced by gradually diminishing solar irradiance in high latitudes, resulting from the Earth’s orbital forces, mostly Milankovich cyclicity. The glacier-friendly climate culmination during LIA (1570–1900 AD) was associated with several episodes of reduced solar activity, particularly during the Maunder and Dalton minima, being amplified by the negative index of NAO circulation and relatively high volcanic activity.

In the Alps, the three principal LIA glacier maxima of ca 1350, 1650, and 1850 AD attained extents comparable to those of previous Holocene maxima. Apart from sedimentological, palynological or dendrochronological pieces of evidence, the long-term and multi-stage deposition of LIA moraines is documented by well-preserved geomorphic record in the form of massive and multi-ridge moraines. One can infer that Holocene conditions of moraine formation were entirely different from those of the Late Glacial, when every successive advance was of smaller extent within the valley. Moreover, the Late Glacial glacier advances were much more shorter than the total time of moraine formation during the Holocene. Direct observations of the youngest LIA events indicate that interactions between climatic changes and glacier behaviour are reflected in fluctuations of glacier fronts and moraine deposition at the scale of even few decades.

Transport and deposition of material in Alpine cirques and valleys varied depending on palaeoclimatic, topographic, topoclimatic, and lithological conditions. A certain continuity between landforms can be inferred between moraines of debris-free glaciers, ablation landforms of debris-covered glaciers, and glacigenic and cryogenic rock glaciers. It does not mean, however, that every landform should have experienced transformation according to the above sequence, but rather conditions prone to origin of given landforms could have gradually changed with time and in space.

Formation of rock glaciers is only possible when a certain proportion between debris supply and snow alimentation is maintained. That is why continental climate is particularly favourable for the develoipment of such landforms. In the Alps, however, the great number of active, Holocene, rock glaciers contrasts with relatively small degree of climatic continentalism. Hence, the most important factor in this case appears to be the relation between altitude of cirque bottom, where debris is being accumulated, to the altitude of snow line, i.e. the topoclimatic factor (Fig. 6), as well as great amount of debris accumulation within cirques throughout the Holocene, i.e. the palaeoclimatic factor.

Some Alpine massifs, rising not very high above the regional Holocene snow line, either were or are glaciated. Therefore, large portions of cirques and valleys are situated below this line, although still within the periglacial belt, provoding suitable conditions for rock glacier formation (Fig. 6). Those massifs which rise high above the regional snow line are covered by large, debris-fee glaciers, and are dominated by glacial transport and moraine deposition.

The conditions suitable for rock glacier formation have changed with time. Transformation of regular glaciers into rock glaciers took place usually during glacier recessions, when glacier ice balance was negative (little snow accumulation) and rock debris supply increased, owing to more intensive physical weathering. In case of small glaciers, which even during the LIA maximum represented cirque glaciers, debris accumulated during successive oscillations tended to build massive morainic ramps (Fig. 3). When the accumulated morainic material was voluminous enough, periglacial conditions led to its slow creeping and rock glacier formation. Successive glacier oscillations supplied rock glaciers in their root zones (debris rock glaciers *sensu* Barsch 1988; Fig 6B). One can conclude that the key role in origin of such rock glaciers was played by high productivity and/or prolonged duration of glacial transport, as well as comparable amplitude of climatic changes throughout the Holocene. It should be added, however, that the response time of rock glaciers to climatic changes is uncomparably longer and completely different than that of regular glaciers (Haeberli 1985). The age of Alpine rock glaciers is frequently estimated at a few thousands of years, implying that they survived warm periods separating the LIA-type events, and are still active during the present-day warming, i.e. they still preserve interstitial ice and are able to creep.

The LIA depositional landforms in the Alps illustrate very well cold epoch processes in a high-mountain environment. Careful examination of these landforms and their origin helps to understand glacial history of those areas, where such landforms are inactive or fossil. Great diversity of landforms produced by LIA in the Alps, an area of relatively uniform palaeoclimatic conditions, provides opportunity to conclude about similarly high diversity in other glaciated mountain massifs in the Carpathians, including the Tatra Mountains. Drawing simplified conclusions on comparable age and origin of rock glaciers throughout the massif and implying palaeoclimatic suggestions is not justified, though. It seems obvious that in mountain massifs like the Tatra Mts, rock glaciers and moraines were formed, depending on local topoclimatic conditions of a cirque or valley, during all climatic deteriorations from the Last Glacial Maximum to the last glacial–Holocene transition. It does not mean that no certain periods existed being particularly favourable to a given type of debris transport and deposition, either glacial or periglacial; however, palaeoclimatic interpretation of individual phases requires careful identification of related landforms and, what is particularly difficult, separation of topoclimatic and topographic factors.

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STRESZCZENIE

Jerzy Zasadni

MAŁA EPOKA LODOWA W ALPACH: JEJ ZAPIS W OSADACH GLACJALNYCH I FORMOWANIU LODOWCÓW GRUZOWYCH

Mała Epoka Lodowa (MEL) była globalnym ochłodzeniem klimatu obejmującym kilka ubiegłych wieków. Ochłodzenie to znacząco odcisnęło swoje piętno w środowisku naturalnym oraz cywilizacji człowieka, szczególnie w rejonie Północnego Atlantyku, w rejonach polarnych i górskich. Dzięki temu, że okres ten obejmował czasy nowożytnie, z metodologicznego punktu widzenia jest to wydarzenie bez precedensu, gdyż, oprócz dużego spektrum metod związanych z naukami o Ziemi, możliwe było zastosowanie w jego badaniu wyników bezpośrednich obserwacji, jakie celowo lub nieświadomie prowadził człowiek. MEL najlepiej poznana została w europejskich Alpach. Istnieją duże rozbieżności co do początku i końca tej epoki. Częściowo rozwiązaniem tego problemu może być przyjęcie za pracą J. A. Matthews i K. R. Briffa (2005) dwóch oddzielnych przedziałów: w sensie glacjologicznym 1300–1950 r. i klimatycznym 1570–1900 r.

Mała Epoka Lodowa była najsilniejszym ochłodzeniem spośród kilku, jakie miały miejsce w holocenie. Poprzednie nasunięcia lodowców rangi MEL jednak nieznacznie tylko ustępowały zasięgowi awansów MEL. Intensywność i długość tych zimnych i wilgotnych epizodów wzrastała począwszy od tzw. optimum klimatycznego holocenu. W Alpach podczas MEL lodowce osiągnęły trzy główne maksima w latach ok. 1350, 1650 i 1850. Te i inne mniejsze oscylacje lodowców podyktowane były fluktuacjami klimatycznymi wynikającymi z cyklicznych zmian aktywności Słońca, wpływem cyrkulacji Oceanu Atlantyckiego i jego interakcji z atmosferą, aktywnością wulkaniczną oraz wewnętrznych, losowych fluktuacji klimatu. Czoła lodowców alpejskich cofnęły się po 1850 r. nawet 2–3 km. Związane jest to z ociepleniem rzędu 1°C i podniesieniem linii wieloletniego śniegu o ok. 150 m. Obecne warunki są bliskie lub porównywalne z tymi okresami holocenu, które uważane są za najcieplejsze.

Poza Skandynawią, w Europie najlepiej wykształcone formy i osady związane z MEL znajdują się w Alpach. Specyficzna sytuacja, jaką były powtarzające się holoceńskie awanse lodowców rangi MEL oraz fluktuacje w obrębie samej MEL, doprowadziła do formowania dość charakterystycznych form morenowych oraz lodowców gruzowych. Moreny określane jako maksymalne stanowią sedymentologiczny zapis większości poprzednich awansów w holocenie, co tłumaczy ich wyjątkową masywność, a także wielogrzbietowy charakter. Wykształcenie form akumulacji glacjalnej w dużej mierze zależy od wielkości lodowców, ich lokalizacji oraz sytuacji topograficznej i topoklimatycznej. Ze tego względu, w Alpach możemy obserwować całe spektrum form: od typowych moren formowanych przez "czyste" lodowce, poprzez kompleksy ablacyjne lodowców pokrytych gruzem, do lodowców gruzowych glacigenicznych i kriogenicznych.

Lodowce gruzowe występują tam raczej w obszarach o marginalnym (karowym) zlodowaceniu. Warunkiem, który decyduje o formie transportu i depozycji gruzu w karach (transport glacjalny a ruch lodowców gruzowych) jest stosunek dostawy materiału skalnego do zasilania śniegiem, co często przekłada się na relację między wysokością linii wieloletniego śniegu związaną z MEL, a dnami karów. Wielokrotnie powtarzające się awanse małych karowych lodowców podczas MEL oraz we wcześniejszych zimnych etapach holocenu doprowadziły do nagromadzenia gruzu i uformowania lodowców gruzowych typu *debris rock glaciers* (*sensu* Barsch 1988). Dość często występują także w Alpach lodowce gruzowe z jądrem lodowym (lodowce gruzowe glacigeniczne), które powstały w wyniku jednorazowej transformacji "czystych" lodowców podczas ostatnich faz MEL.