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Reflection of climate changes in the structure and morphodynamics of talus slopes (the Tatra Mountains, Poland)

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ABSTRACT

Talus slopes beside glaciers are among the best objects to research on climate change. In the Tatra Mountains, the highest mountains of central Europe, no glaciers remain, only glacierets and permafrost. For that reasona complex investigation of talus slopes was conducted there in the years 2009–2010. This paper presents the results of GPR and lichenometric measurements of the talus slopes in six glacial cirques located in the High and Western Tatras. The thickness and internal structure of talus slopes were identified along with the variability and conditions of their development. Maximum thickness of the talus slopes ranges from 20 to 35 m, reaching higher values in the High Tatras. The diversity of the thickness of the talus slopes within the Tatras is mostly explained by differences in the relief conditioned by lithology. The diverse altitudinal locations of the talus slopes, and the exposure and inclinations are not reflected in the size and thickness. The thickness of the studied slopes depends primarily on the activity of the processes supplying rock material and on the size and shape of the sediment supply area. The results of the lichenometric testing together with the analysis of the long-term precipitation data imply a several hundred-year-long deterioration of the climate during the Little Ice Age, which is reflected in the increased activity of morphogenetic processes on the talus slopes across the whole massif of the Tatras. In the last 200 years, the talus slopes of the Tatras were most active in three periods: at the end of the Little Ice Age, in the 1930s and 1940s, and in the early 1970s.

1. Introduction

Talus slopes are among the most widespread elements of high-mountain relief. Their development is conditioned by the geological structure and relief of the sediment supply area (e.g., Densmore et al., 1997; Hetu and Gray, 2000) as well as by the processes of weathering, erosion, transport and accumulation of rock debris (e.g., Kotarba and Strömquist, 1984; Kotarba et al., 1987; Sass and Krautblatter, 2007; Fort et al., 2009; Krautblatter et al., 2012). The former factors are generally invariable over time, whereas the dynamics of geomorphological processes depend on hydrometeorological conditions whose variability reflects characteristics of the climate. Therefore, the thickness, internal structures and microrelief of the surface of debris slopes in glacial cirques may reflect the type of morphogenetic processes and changes in climatic conditions since deglaciation; however, the knowledge of this record is still limited (Sass and Krautblatter, 2007). More about the activity of the slopes may be concluded on the basis of the results of lichenometric dating of the surface of talus slopes, whose time frame is limited, however, merely to the recent centuries (e.g., Bull et al., 1994; Kotarba, 1989, 1995, 2004; McCarroll et al., 2001). On the other hand, correlating the results of lichenometric dating is possible with the results of instrumental meteorological measurements and thus deducing the impact of climate change on the morphodynamics of the slopes.

Frost weathering is the major process responsible for the supply of debris material building the talus slope. This material may be transported to talus slopes by various processes, mainly by debris flows, topplings and snow avalanches. These processes, along with weathering, determine the development of talus slopes. Therefore, various models of the accumulation on debris slopes have been developed to match the conditions in different regions of the world (e.g., Caine, 1969; Statham and Francis, 1986; Rapp and Nyberg, 1987). However, the model of accumulation designed for one region only may be different in time and within a limited space, i.e., at different places located in close vicinity to each other (Whitehouse and McSaveney, 1983).

Many researchers express the opinion that the development of debris slopes is the consequence of the interaction of the processes of detachment by rockfall and redistribution of debris material mainly by debris flows (e.g., Francou, 1991; Kotarba, 1992; Hinchliffe et al., 1998). Over time, the segmentation of rockwalls with couloirs also favours reworking of talus by channeling the water flow onto the talus,

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which causes debris flows and gully incision and increases sediment connectivity (Fryxell and Horberg, 1943; Becht et al., 2005; Sass and Krautblatter, 2007; Heckmann et al., 2012). The beginning of the formation of talus slopes is also disputable. According to Ballantyne (2002), they were formed in the periglacial climate during the period of deglaciation, and later on they were given mere 'retouches' mainly by the debris flow activity. Also Curry and Morris (2004) pointed out a change in the processes shaping high mountain slopes in the period between the Late Glacial and Holocene, emphasizing the increasing role of debris flows in the transformation of the slope and the decreasing importance of microgelivation in the temperate climate zone in the Holocene.

The impact of the last, conspicuous cooling of the climate during the Little Ice Age on the development of debris slopes manifested itself in an increased supply of the material through rockfall (e.g., McCarroll et al., 2001; Kotarba and Pech, 2002) or an increase in the activity of debris flows (e.g., Nyberg and Lindh, 1990; Strunk, 1992; Kotarba, 1995;). The subsequent climate warming resulted in an escalation of catastrophic processes in the mountains (e.g., Evans and Clague, 1994; Fort et al., 2009). Many researchers in recent years have shown the impact of climate changes, including an increase in the number of intense rainfall events and/or in temperatures on the activity of debris flows (e.g., Haeberli et al., 1990; Zimmerman and Haeberl, 1990; Kotarba, 1997; Rebetez et al., 1997; Jomelli et al., 2004; Pelfini and Santilli, 2008).

In the Tatra Mountains, talus slopes have been the subject of numerous studies (e.g., Kotarba et al., 1983, 1987; Kaszowski et al., 1988; Klimaszewski, 1988; Krzemień et al., 1995; Ferber, 2002; Kotarba, 2004). They concerned the origin and contemporary morphodynamics of the slopes. Models of the formation of the talus slopes of this region were presented by Kotarba et al. (1987).

Currently in the Tatras, the most extensively modified slopes are those within an altitude of about 1500 m asl, the upper limit of timberline, and about 1950 m asl. The highest accumulation of loose gravitational deposits (mainly in the form of talus slopes) and glacial and glacifluvial sediments occurs in this area, which are easily displaced under the influence of extreme hydrometeorological events. Because of the morphogenetic role (evidenced by new landforms) and the amount of dislocated material, debris flows belong to the most important processes (Kotarba, 1992, 1995, 1997; Rączkowska, 2006; Rączkowska et al., 2012; Kotarba et al., 2013). Above and below the aforementioned range of altitude, the intensity of geomorphological processes becomes lower partly because of less frequent freeze–thaw cycles (Kotarba et al., 1987).

However, because the thickness and internal structure of talus slopes were unknown, it was not possible to recognize the variability and conditions of the development. This paper presents the results of GPR and lichenometric measurements of the talus slopes in six glacial cirques located in the High and Western Tatras (Fig. 1). The slopes are different regarding altitude, size, and exposure as well as the geological structure and relief of the sediment supply area (Table 1). It was the first time that in such diverse sites within one mountain range of the temperate zone, information concerning internal structure, morphogenesis, and morphodynamics of the talus slopes had been collected; and for the purpose of its interpretation, multidecadal data on the intensity of rainfall had been employed. This allows us to present the complexity of the post-glacial evolution of high mountain talus slopes in the temperate zone, which is the main objective of this work.

2. Regional setting

2.1. Geology and relief

The Tatra Mountains are of alpine character, elevated up to 2655 m asl (Gerlachovský štít peak). The Tatra Mountains are composed of a crystalline core formed of intrusive Carboniferous granitoids of the High Tatras and metamorphic rocks (Paleozoic rocks: gneiss, amphibolite, metamorphic shale) occurring mainly in the Western Tatras. The core is rimmed by allochthonous High Tatric Nappe and Sub-Tatric Nappes, consisting of quartzites, dolomites, limestones, marls, shales, and sandstones of the Triassic-Middle Cretaceous age (Książkiewicz, 1972; Nemčok et al., 1994; Oszczypko, 1995). The entire Belianske Tatras are built up of nappes of Mesozoic sedimentary rocks (Bezák et al., 1993; Nemčok et al., 1994).

In the Neogene, the Tatras were dissected by fluvial–denudational valleys, which underwent rejuvenation in the Pleistocene mainly through glacial and periglacial processes (Klimaszewski, 1988; Baumgart-Kotarba and Kotarba, 2001). As a result, in the High Tatras a system of glacial cirques was formed, often arranged in tiers, and glacial troughs, both with very steep and rocky slopes, deeply incised by couloirs. At the outlets of the couloirs, large talus cones developed, which are currently modified mainly by debris flows (Kotarba 1992, 1997). In the Western Tatras only the upper sections of the valleys underwent glaciations. The bottoms of the cirques are not overdeepened, and the rocky slopes and rock walls are shorter than in the High Tatras and often replaced by debris-mantled slopes (Klimaszewski, 1988). Currently, in the Tatras only glacierets and

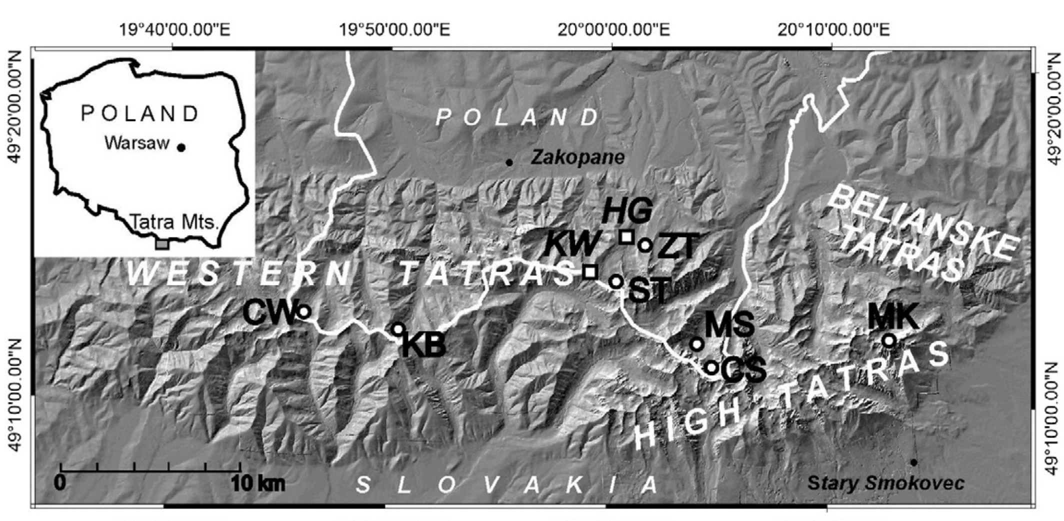

Fig. 1. Study area. Dots — location of studied slopes: CW — Szeroki Piarg on N slope of Wołowiec peak, KB — N slope of Błyszcz peak, ST — N slope of Skrajna Turnia peak, ZT — W slope of Żółta Turnia peak, MS — Szeroki Piarg over Morskie Oko lake, CS — Wielki Piarg over Czarny Staw pod Rysami lake, MK — Medená kotlina valley. Quadrats — location of meteorological stations: KW — Kasprowy Wierch, HG —Hala Gąsienicowa.

Table 1
Characteristics of study sites.

Symbol	Location	Landform	Geology	Relief of vicinity	Dominant geomorphic processes	Vegetation cover
CS	High Tatra, Wielki Piarg over Czarny Staw pod Rysami lake	Debris cones	Granitoids	Rocky slopes with partly developed hanging glacial cirque above the cone	Debris flows, gravitational processes, snow avalanches	Sprase alpine meadow
MS	High Tatra, Szeroki Piarg over Morskie Oko lake	Debris cone	Granitoids	Hanging glacial cirque with high rockwalls and rocky step above the cone	Debris flows, snow avalanches, gravitational processes	Upper part —lack, lower part — sprase alpine meadow and dwarf pine
ZT	High Tatra, W slope of Żółta Turnia peak	Debris slope	Granitoids	Rocky slope above the cone	Debris flows, snow avalanches	Sprase alpine meadow
ST	High Tatra, N slope of Skrajna Turnia peak	Talus cone	Granitoids	Rockwall above the cone	Gravitational processes, snow avalanches	Lack
KB	Western Tatra, N slope of Błyszcz peak	Talus cone	Gneiss, metamorphic shale	Debris-mantled slope and rocky slope above the cone	Gravitational processes, snow avalanches	Sprase alpine meadow
CW	Western Tatra, Szeroki Piarg on N slope of Wołowiec peak	Debris slope	Gneiss, metamorphic shale	Debris-mantled slope and rocky slope above the cone	Debris flows, snow avalanches, gravitational processes	Sprase alpine meadow

perennial snow patches exist (Gądek, 2008; Kędzia, 2015). The largest glacieret has ca. 2 ha of area (Gądek, 2014). In some shaded areas above 1900 m asl, small patches of sporadic permafrost occur; however, currently, its impact on periglacial processes is negligible (Dobiński, 1998; Kędzia et al., 1998; Rączkowska, 2006; Gądek and Kędzia, 2008; Gądek et al., 2009; Mościcki and Kędzia, 2001).

2.2. Climate

The Tatra Mountains are located in the temperate climate zone. Mean annual air temperature (MAAT) changes in the vertical profile from more than 3 °C in the bottoms of valleys to about −4 °C on the highest peaks (Table 2). At an altitude above 1500–1600 m asl, frost may occur at any time of the year; whereas on the highest peaks merely a few days occur without frost or freezing cold (Hess, 1965, 1996; Konček, 1974). The highest number of days with freeze–thaw cycles occurs in the range of altitude from ~1700 to about 2050 m asl, i.e., where the MAAT isotherm of 0 °C runs (Klimaszewski, 1971).

The mean annual total precipitation (MAP) increases along with the altitude, but the highest amounts of precipitation occur on the northern slopes between 1500 and 1900 m asl, and this is mainly associated with rainfall of the warm half-year (Table 2). In the cold half-year, however, the largest amounts of precipitation have been recorded on the highest peaks (Hess, 1965, 1996; Konček, 1974; Niedźwiedź, 1992).

The number of days with snow cover ranges from about 100 at the foot to nearly 300 on the highest peaks (Hess, 1965, 1996). The thermal climatic snowline runs at an altitude of about 2500–2600 m asl (Zasadni and Kłapyta, 2009).

3. Data and methods

3.1. Geomorphological mapping

To determine the main features of the relief of the studied areas, geomorphological mapping was employed. In the years of 2009–2010, geomorphological maps of the studied areas were drawn at the detailed scale of 1:5000 on the background of orthophotomaps of 2007, using the modified key of Klimaszewski (1968).

3.2. GPR surveys

For testing the thickness and internal structure of the talus slopes, the ground-penetrating radar (GPR) method was applied (Otto and Sass, 2006). It consists of emitting electromagnetic waves at a specific frequency into the ground and recording the waves reflected from the boundaries between structures of different dielectric properties. The velocity of the propagation of electromagnetic waves and the range of depth depend on dielectric properties of sediments (e.g., Berthling and Melvold, 2008).

For the purpose of our fieldwork, a RAMAC/GPR CUII pulse radar (Mala GeoScience) was used with unshielded antennas at a central frequency of 25 MHz. The distance between the transmitting and receiving antennas was 4 m. To establish a precise location of the recorded signals, a Leica 1230 GPS receiver was used.

The GPR measurements were carried out along and across the investigated talus slopes. The locations are presented in Fig. 1. The total length of the measurement lines was 2500 m. The radar traces were recorded every 10–35 cm (depending on the speed of the movement of the antennas). Time windows of the measurements ranged from 300 to 1000 ns. The velocity of the propagation of electromagnetic waves in the ground was determined by the method of common mid-point (e.g., Neal, 2004). The average velocity of the propagation of electromagnetic waves varied between 10 (MS, KB, CW) and 11 cm/ns (CS, ZT, ST). The calibration of GPR results was carried out on the talus slope of Kežmarsky štít peak in the Medená Kotlina valley (Fig. 1). This slope is dissected up to solid rock by a corrasion gully. This enabled the direct insight into the internal structure of the talus and measurements of its thickness by laser rangefinder. For digital processing and interpretation of the radar data, a RadExplorer v. 1.4 (DECO-Geophysical Ltd.) programme was employed.

Table 2
Characteristics of climate of the Tatra Mountains (Hess, 1965; Konček, 1974; Niedźwiedź, 1992).

Meteorological station		Kasprowy Wierch	Hala Gąsienicowa	Dolina Pięciu Stawów Polskich	Lomnický štít	Skalnaté Pleso	Štrbské Pleso	Hala Ornak
Altitude [m asl]		1991	1520	1670	2635	1786	1330	1109
Mean annual air temperature [°C]		−0.8	2.3	1.1	−3.8	1.6	3.4	3.2
Number of days with temp. Min <0 °C and max >0 °C		78	93	94	88	111	118	126
Number of days without frost		48	102	82	4	93	120	120
Precipitation [mm]	January	142	70	71				74
	July	215	247	260				206
	Mean annual	1889	1664	1692	1645	1323		1490

3.3. Lichenometry

The recent activity of the talus slopes, especially the debris flows, was examined in the summers of 2008 and 2009. To estimate the age of the surface of the talus slopes, the lichenometric method was applied, the basics of which were developed by Beschel (1950, 1957). In the Tatras, this method was used for the first time by Kotarba (1988), who developed separate lichen curves for the alpine and subalpine belts, using *Rhizocarpon geographicum*. These curves were used by the authors for dating the surfaces of the studied talus slopes. The measurements included only the largest lichen whose shape resembled a wheel. Next the percentage of the lichen of a certain age was determined in relation to the area of the surface of the talus slope or the length of the transverse profile of the talus slope, located in the central part of its vertical profile.

3.4. Precipitation data

To interpret the results of the lichenometric measurements performed on the studied talus slopes, precipitation data from Hala Gasienicowa (HG) and Kasprowy Wierch (KW) meteorological stations (Fig. 1) were applied. The stations are the only two that exist in the region. The first station is located at an altitude of 1520 m asl (upper timberline) in the bottom of Dolina Suchej Wody valley, and the second one is at an altitude of 1991 m asl on the top of Kasprowy Wierch (KW) above the valley. The distance between the stations is about 2 km. These data comprise the daily totals and intensity of precipitation in the years 1924–2004 (HG) and 1951–2001 (KW).

4. Results and discussion

4.1. Morphology of the slope system

In the Tatras, the system consisting of a rock wall or a rock slope and a talus slope is particularly well developed in cirques and glacial troughs, and therein are the talus slopes selected for the studies (Table 1), located on the seminival, alpine, and subalpine belts. Talus slopes, like rock slopes and rock walls, are diverse in terms of size, vertical extent, and gradient. The gradients of talus slopes usually range from 25° to 31°. Rock slopes are fragmented by couloirs, whereas talus slopes often have a diverse microtopography, reflecting the type of processes that shape them. Distinct differences of the relief of the discussed system are noticeable between the High and Western Tatras. This diversity is illustrated by exemplary geomorphological maps (Figs. 2 and 3) of the representative areas.

The study site of Szeroki Piarg over Morskie Oko lake (MS) in the High Tatras constitutes a vast talus cone of a north exposure inserted into the glacial lake basin (Table 1, Fig. 2). It is situated below the 300-m high rock threshold of the Kocioł Męguszowiecki glacial cirque, over which, in turn, rise the 400-m high rock wall of Pośredni Męguszowiecki Szczyt peak (2392 m asl) and Wielki Męguszowiecki Szczyt peak (2438 m asl).

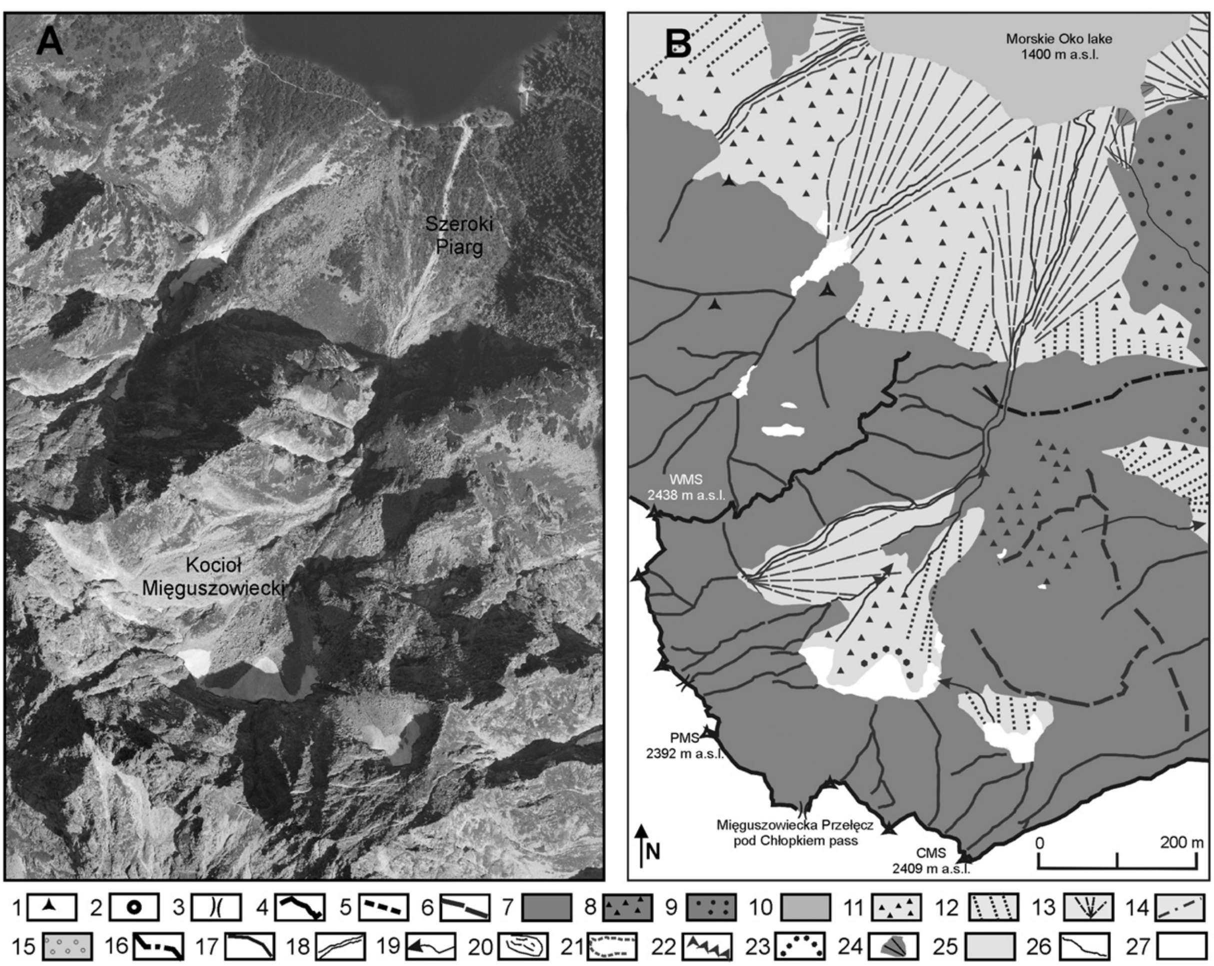

Fig. 2. The Szeroki Piarg over Morskie Oko lake studied slopes (MS). (A) Airphoto; (B) geomorphological sketch. Legend for geomorphological sketches on this figure and Fig. 3: 1 — sharp summits, 2 — rounded summits, 3 — passes, 4 — sharp rocky ridges, 5 — rounded ridges, 6 — convex break in rocky slope, 7 — rockwall and rocky slopes, 8 — rocky slopes covered by rockfall debris, 9 — rocky slopes covered by moraine, 10 — debris-mantled slopes, 11 — rockfall talus slopes, 12 — talus slopes, 13 — alluvial slopes, 14 — alluvial–avalanche slopes, 15 — moraine deposits, 16 — step of hanging glacial cirque, 17 — rocky chute, 18 — large debris flow gullies (>5 m width), 19 — debris flow gullies, 20 — debris flow tongue, 21 — erosional niches (avalanche erosion), 22 — front of fossil rock glacier, 23 — protalus rampart, 24 — delta, 25 — lake, 26 — streams, 27 — perennial snow patches. WMS — Wielki Mięguszowiecki Szczyt peak, PMS — Pośredni Mięguszowiecki Szczyt peak, CMS — Czarny Mięguszowiecki Szczyt peak.

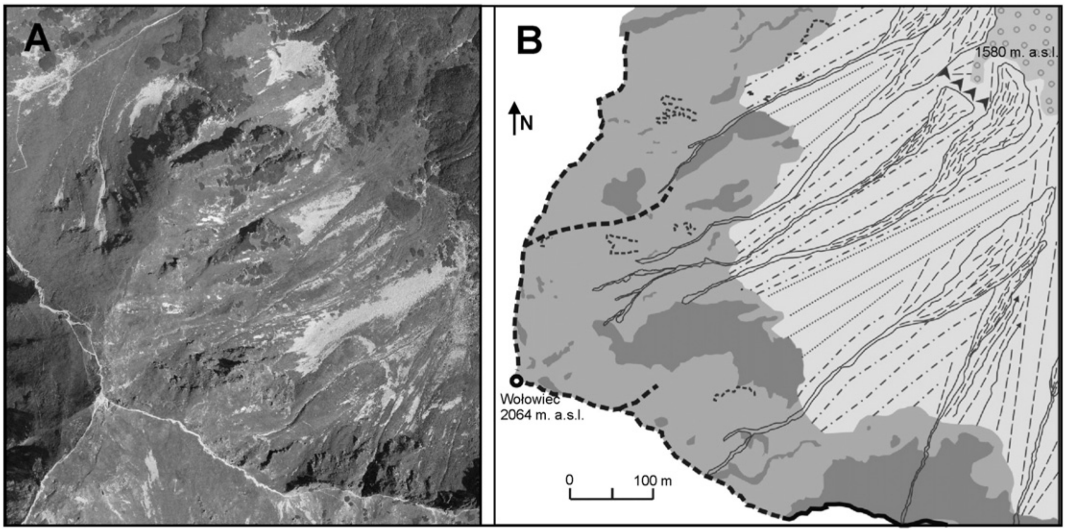

Fig. 3. The Szeroki Piarg on the northern slopes of Wołowiec peak in the Chochołowska Valley (CW). (A) Airphoto; (B) geomorphological sketch — for explanation see Fig. 2.

The total length of the talus cone is ~600 m (including 200 m under water), and its average gradient is close to 22°. The whole slope (from the ridge to the bottom of the lake) has a stepped longitudinal profile. The gullies with side levée occurring on the surface of the Szeroki Piarg talus cone imply that debris flows are the dominant processes that shape the slope (Fig. 2).

The study site of Szeroki Piarg in the Chochołowska Valley (CW) in the Western Tatras makes up a vast talus cone overlapping a subslope fossil rock glacier. It extends to the northern side of Wołowiec peak, below small rock walls (Fig. 3). Its length is ~450 m and an average gradient is close to 25°. This slope has been formed primarily by debris flows and snow avalanches. A smooth longitudinal profile and a relatively great length of this slope together with its gradient are conducive to the occurrence of dirty avalanches, which also has been confirmed by direct field observations. In contrast, a significant contribution of debris flows to the modification of the slope is indicated by the presence of three distinct debris flow gullies, which are accompanied by levées, as well as by numerous smaller forms of this type.

The relief of the other talus slopes selected for the studies is similar. The microrelief of the slope surfaces indicates that the formation largely results from rockfall and debris flows. A clear difference exists, however, between the development of the rock slope–talus slope systems in the High Tatras and the Western Tatras. In the High Tatras the systems are often tiered, and rock walls are much higher.

4.2. Structure and thickness of the talus slope

The radar echograms of the studied slopes revealed the thickness of talus deposits and the main elements of the internal structure. The vertical resolution of the images is 1.6 m, while the range of depth exceeds 40 m. The results of the calibrations of the GPR measurements are shown in Fig. 4. The average velocity of the propagation of electromagnetic waves inside the slope below Kežmarsky štit peak in the Medená Kotlina valley was 10 cm/ns. The structure of the echogram reveals a clear contrast between loose deposits of variable grain size and its granite substrate. The part of the echogram that shows loose talus material is made up of clear, ribbed reflections revealing features of the internal structure of the slope, whereas the part of the echogram depicting solid rock is lacking in radar reflection patterns. The most deeply located reflective horizon was recorded at a depth of ~15 m. It reflects the course of the substrate of the slope sediments, which was confirmed by the results of the direct measurements. Thus, the thickness of the other tested talus cones was determined in a similar way (Figs. 5 and 6). The average velocity of the propagation of electromagnetic waves inside the other slopes varied between 10 (MS, KB, CW) and 11 cm/ns (CS, ZT, ST).

4.2.1. The Wielki Piarg over Czarny Staw pod Rysami lake study site(CS)

The thickness of loose material in the part above the lake level varied from ~10 m at the apex of the talus slope to ~35 m in the inshore zone. A clear reflective horizon, an extension of the surface level of Czarny Staw pod Rysami lake, was also recorded (Fig. 5A and 6A). It reflects the level of groundwater.

4.2.2. The Szeroki Piarg over Morskie Oko lake study site (MS)

The results of the radar sounding point to an incoherent internal structure of the whole talus slope. Its thickness in the above-water zone ranges from 15 m (in the upper part of the profile) to ~30 m (in the lower part). In the upper part of the slope, several artifacts were recorded from reflections from the rock walls. In the lowest part, however, between 360 and 420 m of the measurement profile, two horizontal reflective horizons, were recorded. The higher one coincided with the surface of the slope at 390 m of the profile, at the apex of the alluvial cone composed of fine material. In contrast, the reflective horizon situated at a lower level course corresponds to the position of the surface of Morskie Oko lake and represented the groundwater table (Figs. 5B and 6B).

4.2.3. The Żółta Turnia peak study site (ZT)

The average velocity of the propagation of electromagnetic waves was close to 11 cm/ns. The thickness of the slope sediments varied from 0 in the upper part of the profile to ~25 m at the base of the slope, where they overlap moraine formations (Figs. 5C and 6C). The echogram demonstrates a sequence of stratified debris, irregular reflection patterns, and bedrock (see: Sass, 2008), which are typical of such slopes.

4.2.4. The Skrajna Turnia peak study site (ST)

The thickness of the talus cone varied from 2 m at its apex to ~20 m in the central part and at its base (Figs. 5D and 6D). In view of these results, the swelling supporting the talus cone, overgrown with dwarf pine, is composed of debris material and is a fossil ridge of a nival moraine or fossil subslope rock glacier.

4.2.5. The Błyszcz peak study site (KB)

The thickness of the debris varied from 2 m in the upper part of the slope to 20 m at its base (Figs. 5E and 6E), where the slope deposits overlap the lobe of the fossil subslope rock glacier. This was clearly

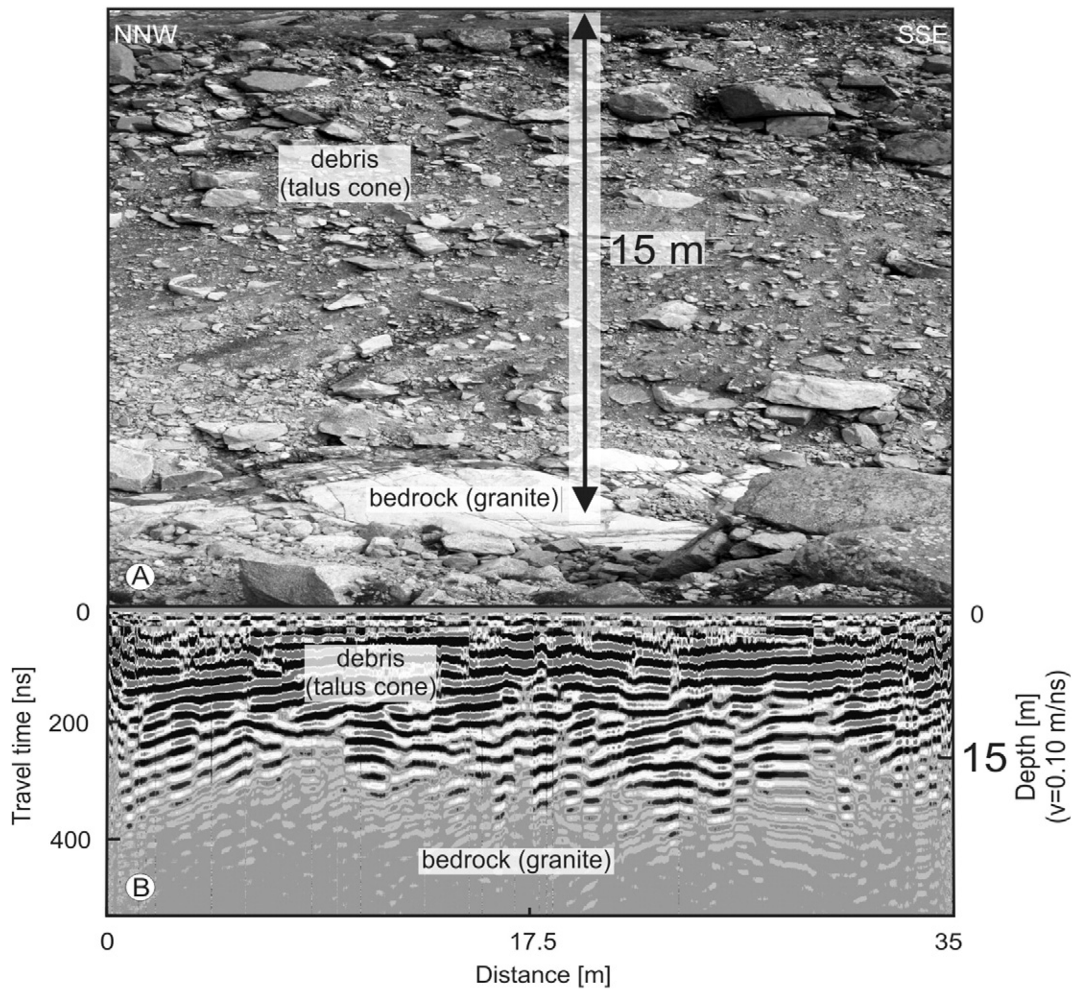

Fig. 4. The internal structure of the talus cone at Kežmarsky štit peak in the Medená kotlina valley: (A) natural exposure and (B) the GPR section (25 MHz).

reflected in the structure of the radar image (surface-parallel/irregular reflection patterns). The maximum thickness of the slope sediments is ~12 m.

4.2.6. The Szeroki Piarg on Wołowiec peak study site (CW)

The average velocity of the propagation of georadar waves in the debris substrate was 10 cm/ns. The registered reflective horizons imply differences in the physical characteristics of the upper, middle, and lower slope segments (Figs. 5F and 6F). The upper part of the slope cover (0–97 m of the profile), with an incline of ~33° and a thickness of up to 15 m, revealed similar dielectric characteristics within the whole of its mass. Below, in the section from 97 to 310 m of the measurement profile, where the incline of the slope surface is 28°, the thickness of the detrital material amounts to 25 m, and the registered reflective horizons indicate a change in the physical characteristics at depths of 10 and 15 m. A very distinct change in the dielectric properties of the substrate was also recorded in the lowest part of the measurement profile (between 310 and 350 m). At this flattening, ending with a swelling, the total thickness of the loose material decreases from ~20 to 10 m. The reflective horizon registered above the bedrock surface indicates that the formations that make up the debris swelling (fossil rock glacier; Fig. 3) have a thickness of ~10 m, and are buried in the talus material (Fig. 6F).

4.3. Recent dynamics of the talus slope

The results of the lichenometric tests show that the talus slopes in the Tatras are being continuously modified, however, with variable intensity in time and space (Fig. 7). The analysis of the results of the lichenometric dating of the studied slopes (Fig. 7) reveals that in the High Tatras and the Western Tatras major intensification of slope processes was characteristic of a period of the Little Ice Age. Taking into consideration the ubiquity of modeling slopes, periods of increased activity of geomorphological processes were determined. What constituted the basis was either a small (below 10%) percentage of the surface of a certain age, however, on most of the examined slopes or a greater percentage of the surface of a certain age on merely a few slopes, even in the absence of such a surface on the others. The periods determined on this basis were the years of 1820–1830, 1850–1860, 1880–1900, and 1910–1920. Generally, they correspond to the periods of glacier advances in the Alps reported, inter alia, by Vivian (1975), Bachman (1979), Röthlisberger et al. (1980), and Zumbühl et al. (1983) and with the periods of slope alluviation in the Tatras specified by Kotarba (1995, 2004). On many of the talus slopes, the percentage of the surface formed during the Little Ice Age is still very large.

Another period of an increased intensity of modeling the slopes was the 1930s and 1940s. These were relatively warm years, with a few severe winters in the Tatras and a large flood in the Western Carpathians in 1934 (Niedźwiedź, 1996, 2000, 2004). Then, on most of the talus slopes a period of relative stability occurred until the 1970s, when a period of intense slope modifying, mainly by debris flows, occurred. This resulted from a relatively large annual precipitation in the years of 1968–1980. After 1980, annual totals of precipitation decreased (Niedźwiedź, 1996), but the number of days with daily precipitation exceeding 50 mm increased (Fig. 8). The variability and diversity of the

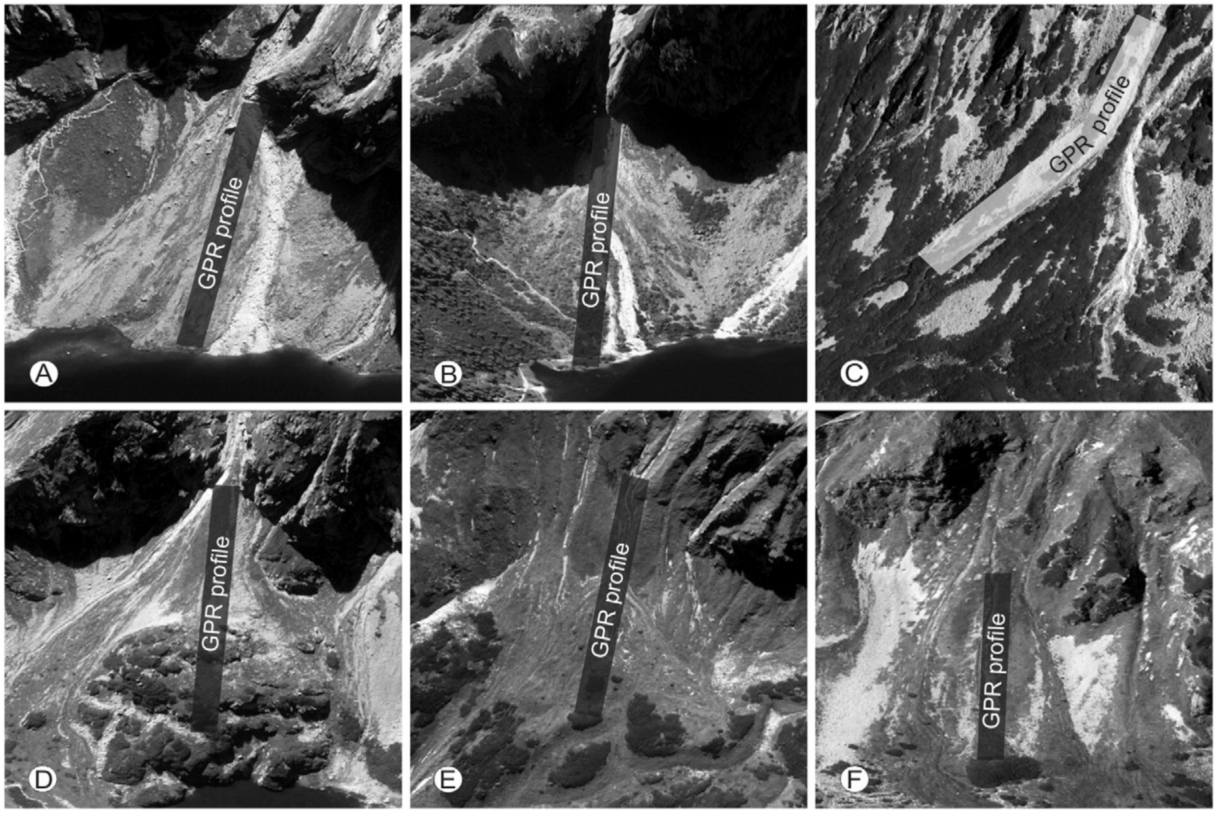

Fig. 5. The study talus cones and location of the GPR profiles: CS (A), MS (B), ZT (C), ST (D), KB (E), and CW (F).

modeling of the slopes in the last period, i.e., the second half of the twentieth century, observed in the Tatras and in the French Alps (Jomelli et al., 2004) are similar.

The aforementioned phases, in particular the oldest ones, are not noticeable on all of the studied talus slopes. The cause of this lies mainly in the spatial variation of the intensity of precipitation, as well as in overlapping older formations by younger debris material. Nevertheless, on some of the slopes, the transformation of the slopes has been occurring rather evenly over the last 200 years. The most conspicuous among the aforesaid periods are: the end of the Little Ice Age (mainly the years

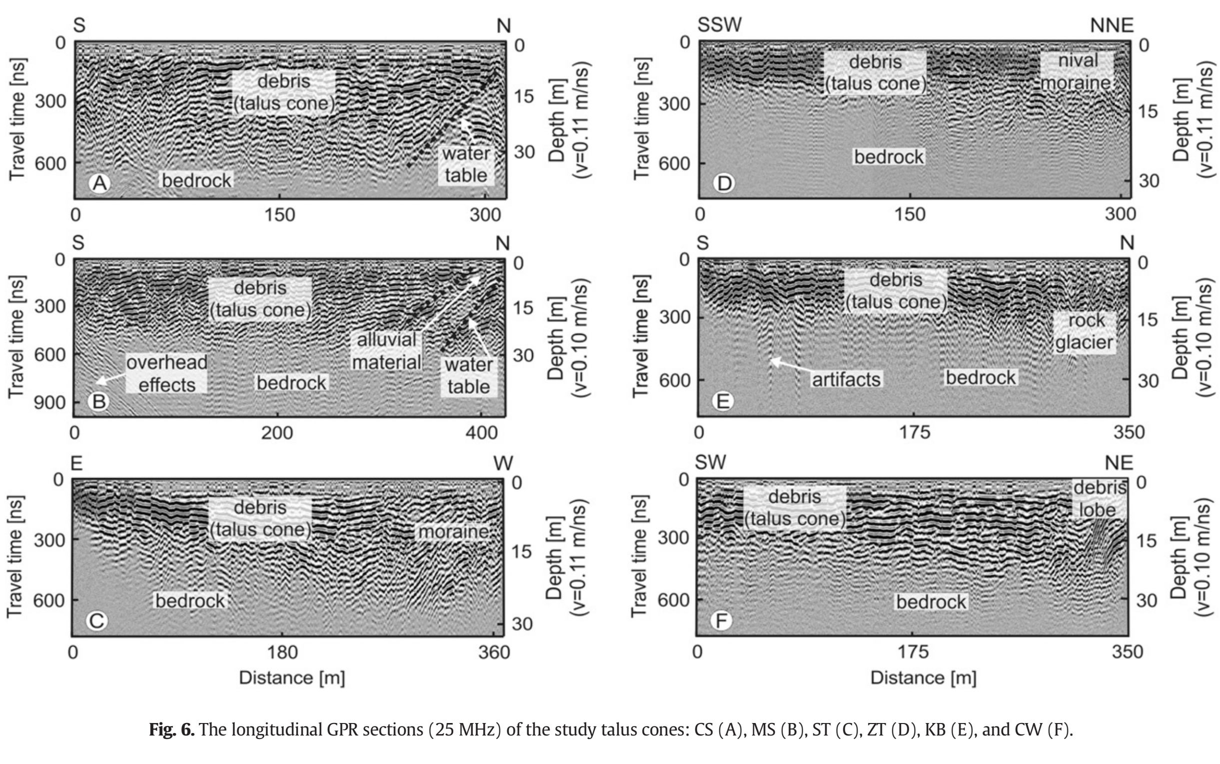

Fig. 6. The longitudinal GPR sections (25 MHz) of the study talus cones: CS (A), MS (B), ST (C), ZT (D), KB (E), and CW (F).

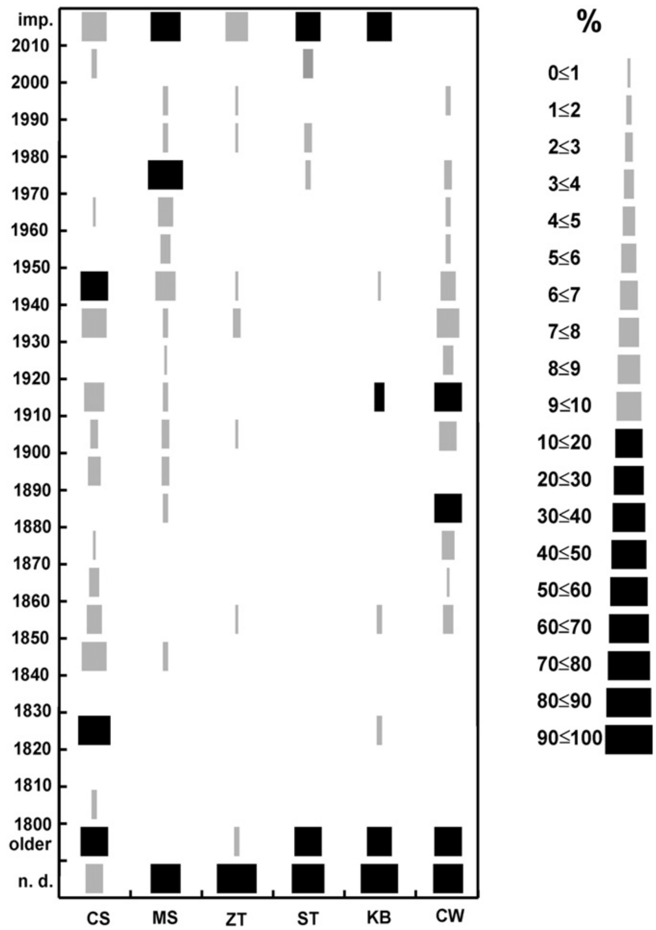

Fig. 7. The percentage of the surface of a certain age in the entire surface of the talus slope. Abbreviations: imp. — fresh surfaces without lichen; older — surfaces dating before 1800; n.d. — surfaces whose age cannot be determined because, e.g., the appearance of vegetation (dwarf pine, alpine grasslands). CS, MS, ZT, ST, KB, CW — study sites.

1910–1920) and the youngest two periods (1930–1950 and the period after 1970).

4.4. Variability of precipitation

The intensity and duration of rainfall play an important role in triggering debris flows (Caine, 1980; Kotarba, 1992, 1995; Van Steijn, 1996; Blijenberg, 1998). According to Kotarba (1997), rainfalls of more than 30 mm/h intensity triggers debris flows in the Tatras. Clear spatial heterogeneity of precipitation in alpine areas, however, was proved (e.g., Cebulak, 1983; Sevruk, 1997).

In the analyzed period, the highest daily precipitation totals were recorded on Hala Gąsienicowa (HG). Also the number of days with precipitation exceeding 50 mm was higher there than on Kasprowy Wierch; however, only some of them resulted in debris flows. Even very heavy rainfall in July 1997, with a maximum daily total of 223.5 mm on 8 July, which brought about significant changes in the morphology of the channels, caused relatively little changes on the slopes. At the same time, debris flows within the talus slopes were triggered by rainfall of a daily total of merely 24.3 mm, which was the case on 20 June 1986 (Kotarba, 1989). On the other hand, the research (lichenometric dating and interpretation of aerial photographs) on the talus cone below Żółta Turnia peak (Fig. 1) showed no activity of debris flows in the years 1945–1985 (Kędzia, 2010), despite that within this period, daily precipitation totals exceeded 50 mm (Fig. 8). The reason for the stabilization of the talus slopes during this period, documented (among other) by Kotarba et al. (1983), was the low intensity of the rainfalls. Convective precipitation of high instantaneous intensity amounting to 17 mm/20 min triggered a debris flow below Żółta Turnia on 20 June 1986 (Kotarba, 1989). The debris flows on 16 August 1988 and 9 August 1991 were also caused by high-intensity rainfall of over 1 mm/min. In contrast, the rainfall on 8 July 1997, despite a very large total amount, did not cause major changes in the morphology of the debris flow gullies that had already existed in the area of Hala Gąsienicowa (Kotarba 1998). The reason was because of a relatively low intensity of the precipitation. On 8 July 1997, rainfall of 0.8 mm/min intensity lasted merely a few minutes, and the maximum hourly intensity was about 30 mm. The analysis of the available pluviogrames also showed that during the aforementioned period of the stabilization of the talus slopes, the hourly intensity of the rainfall reached a maximum of about 20 mm.

Periods of a greater number of days with the daily precipitation exceeding 50 mm (2 days or more) correspond to the periods of increased activity of the slope processes determined on the basis of the results of the lichenometric tests (Figs. 7 and 8).

As all of the debris flows registered in the past several dozen years took place during the summer, snow cover should be ruled out as a factor triggering debris flows. The only exception may be considered high hanging glacial cirques, because of the delayed melting of snow cover (June). On the other hand, the presence of snow cover prevents the triggering of debris flows even by suitably high rainfall (Kotarba et al., 2013).

5. Environmental factors in the development of talus slope

The results of the GPR measurements on the slopes of the Tatras confirm all the observations and conclusions of similar work in the Alps (Otto and Sass, 2006; Sass and Krautblatter, 2007; Sass, 2008). This applies to the scope of the propagation velocities of GPR waves and to the reflection patterns of debris sediments of different genesis and the debris/bedrock interface. Moreover, in topographical situations similar to those described by Sass (2008), we registered groundwater tables and interference associated with overhead effects (secondary reflections of rock walls).

The maximum thickness of the talus slopes in view of the GPR soundings ranges from 20 to 35 m, reaching higher values in the High Tatras (Table 3). The diversity of the talus slope thickness within the Tatras refers mainly to the differences in the relief conditioned by lithology. A greater resistance of the granite rocks in the High Tatras determined better conditions for the development of glacial landforms, glacial cirques, and troughs in this area, surrounded by extensive rock walls and rock slopes. They constitute the sediment supply area for the talus slopes, which is more extensive here then in the Western Tatras. This limits the possibility of determining the rate of slope retreat in the Holocene derived from sediment thickness (e.g., Otto and Sass, 2006).

The lower part of all studied talus slopes overlaps periglacial or glacial forms of different ages. Because most of the researched slopes are located in glacial cirques, this indicates that the talus slopes may have been formed after the retreat of the glaciers, which contradicts the paraglacial theory (Ballantyne, 2002) and the opinions that most of the talus slope deposits have persisted unaltered over the Holocene (e.g., Sass and Krautblatter, 2007).

The diverse altitudinal positions of the talus slopes, as well as the exposures and inclinations, are not reflected in the size and thickness (Table 3). This also testifies that the thickness of the studied slopes depends primarily on the activity of the processes supplying rock material. Moreover, the size and shape of the sediment supply area determine the amount of delivered weathered material and the type and dynamics of slope processes (Kotarba et al., 1987).

6. Relation between climate and talus slope development

The results of lichenometric testing together with the analysis of long-term precipitation data imply a several-hundred-year long deterioration of the climate during the Little Ice Age, which is reflected in the

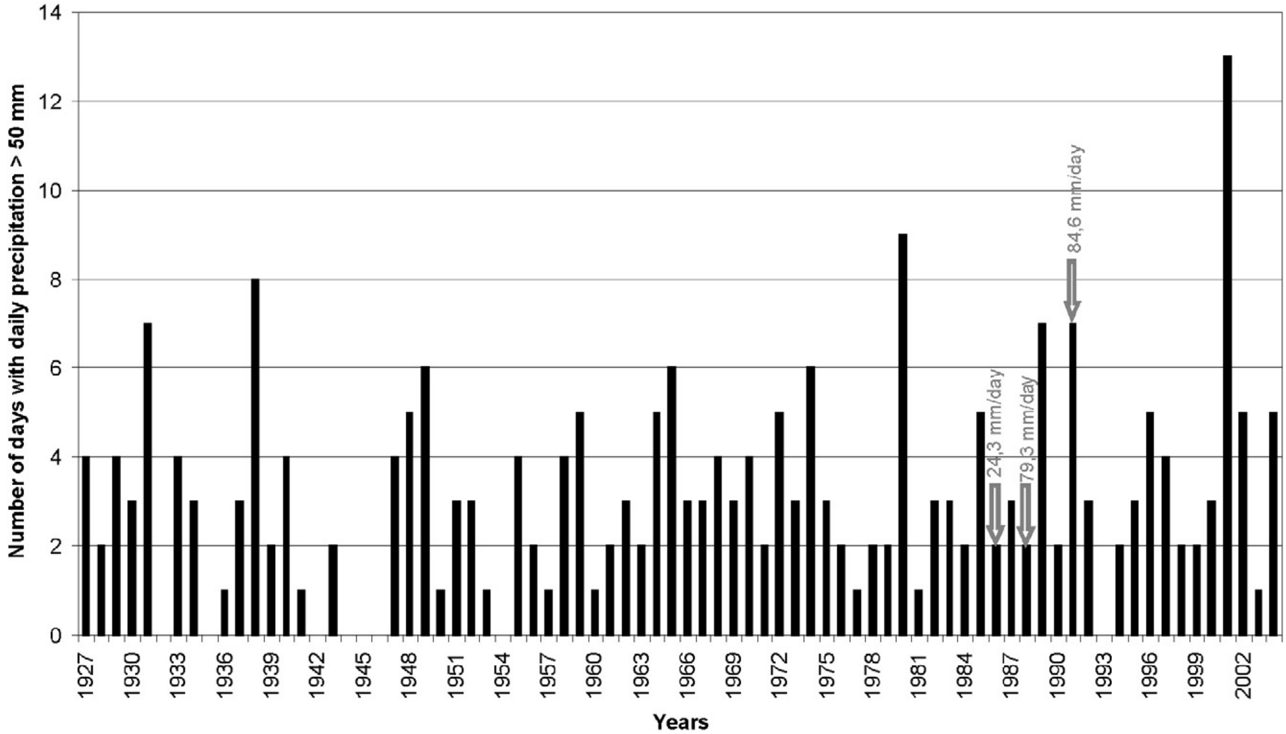

Fig. 8. The number of days with daily totals exceeding 50 mm at Hala Gąsienicowa meteorological station. The arrows indicate the years within which debris flows occurred in the area of Hala Gąsienicowa. Daily rainfall totals on the days when the debris flows occurred are presented over the arrows.

increased activity of morphogenetic processes on talus slopes across the whole massif of the Tatras (Fig. 7). It terminated in the 1920s. In the later period, a lack of a uniform record of the activity of the talus slopes for the whole massif testifies to its high temporal variability and spatial diversity. Occurring after the Little Ice Age, periods of increased Atlantic influence manifested, above all, in a greater number of local downpours or the occurrence of single wet years. As an example, in 1974, rainfall extended over the whole region of the Polish Carpathians. At the same time, the record of slope activity preserved in the lichenometric data is incomplete because the slopes have overgrown unevenly and older deposits are overlain by younger. A large percentage of the surface of the talus slopes deprived of lichen testifies to an ongoing activity of morphogenetic processes.

7. Conclusions

The maximum thickness of the talus slopes depends primarily on the activity of the processes supplying rock material and the size and relief of the sediment supply area. Altitudinal position and slope exposure are not of such great significance. No relationship was found between the thickness of the talus and its vertical extent.

Variations in the activity of morphogenetic processes on the talus slopes, which is much larger and continuous in the High Tatras, are noticeable even on a local scale. This underlines the significance of the size and relief of the sediment supply area.

In the last 200 years, the talus slopes of the Tatras were most active in three periods: at the end of the Little Ice Age, in the 1930s and 1940s, and in the early 1970s. The aforementioned periods were not reflected identically on each of the talus slopes, which may be attributed, inter alia, to the spatial diversity of rainfall.

During the aforesaid periods, the activity of the talus slope resulted from an increased supply of debris material from the rock walls and slopes to the talus slope. Debris flows had the greatest share in material transportation. They also made the most significant contribution to remodelling the talus slope in these periods by relocating the material within their range.

The analysis of the rainfall data collected so far indicates that the daily total of rainfall is much less important than the intensity of rainfall. The greater the intensity of rainfall, the shorter the time needed to trigger debris flows. We can assume that with an intensity of 1,0–1,5 mm/min, debris flows may occur about 15 min after beginning of precipitation. Together with an increased amount of rainfall of high intensity, the frequency of debris flows increases, hence an increase of their participation in transporting material from rock walls to the talus slope as well as in shaping it.

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Table 3
Morphometric characteristics of the studied talus slopes in the Tatra Mountains.

Name/symbol of study site	Height range [m asl]	Exposure	Vertical extent [m]	Surface slope [°]	Maximum thickness [m]
CS	1583–1770	N	187	25	35
MS	1395–1645	N	250	25	30
ZT	1650–1850	W	200	31	20
ST	1700–1900	N	240	31	20
KB	1600–1750	N	150	26	20
CW	1600–1800	NE	200	25	25

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