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Precipitation as a factor triggering landslide activity in the Kamień massif (Beskid Niski Mts, Western Carpathians)

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Abstract. On the landslide slope in the Beskid Niski Mts (Western Carpathians) 48 silver firs were cored for dendrochronological samples. Tree-ring widths were measured for the upslope and downslope sides of each stem. Events of landslide activity were dated using the method of the eccentricity index. The tree-ring record of landsliding was compared with the occurrence of precipitation in the study area. The nature of the relation between precipitation and landsliding is complex. We have found a statistically significant correlation between landsliding and the number of days with 24-hour precipitation totals above 20 mm and high 3-, 5-, and 10-day precipitation totals during winter half-years. Thus landsliding in the Kamień massif is triggered mainly by high precipitation totals in the preceding winter period. No such relation was found for annual precipitation totals and different types of precipitation totals in the summer period. Single landsliding events related to high summer precipitation totals were found, but the correlation is not statistically significant. In addition some landsliding events are 1–2 years lagged after the occurrence of high long-term precipitation totals. It seems that the strongest landsliding events resulted from sequences of wet summer, wet winter and once again wet summer seasons directly following one another.

Key words
landslide,
precipitation,
dendrogeomorphology,
Beskid Niski Mts

Introduction

Three types of precipitation have been distinguished, depending on their amount and intensity, triggering different geomorphological processes (Starkel 1986, 1996). The first type is short-term local downpours with an intensity of 1–3 mm/min, responsible for triggering of soil ablation and down-flows. The second type is long-term downpours (150–400 mm during 2–5 days), resulting in the formation of shallow landslides and transformations of river beds. The third type is long rainy periods (100–500 mm in a period lasting up to several months), responsible for the activity of deep-seated rocky and regolith landslides.

However, the course of the particular precipitation events is usually complex and it is difficult to determine in an unambiguous manner the thresholds of the amount/intensity of precipitation above which geomorphological processes are triggered. The effectiveness of the given precipitation depends also on local relief, vegetation cover, soil type and its humidity in the period immediately preceding the occurrence of the process (Margielewski et al. 2008). It is especially difficult to determine the amount of precipitation triggering the landslides, mainly due to the fact that the landslide slopes are composed of various land forms, clearly separated from each other, e.g. landslide niche and tongue. There are smaller land forms within the niche and tongue, each of which can be subject to dislocation during various precipitation events (Migoń et al. 2014). Yet, there is a clear correlation between the amount of precipitation and the activity of Carpathian landslides dated dendrochronologically, as was indicated by previous studies in the area of the Polish Flysch Carpathian Mountains (Ziętara 1968; Gil, Długosz 2006; Starkel 2011).

Dendrochronological investigation can be of use in determining the activity of slight landslide movements not visible in the field. Previous dendrochronological studies allowed researchers to detect a high frequency of small-scale landslide movements occurring almost every year in the Carpathians (Wistuba et al. 2013). This was possible due to the fact that the trees which grow on landslide slopes are bent during the events of landsliding and annually record landslide events in their rings (Bramm et al. 1987; Corominas, Moya 2010). A landsliding event can be recorded in the same calendar year (annual ring) when the event occurred or with a one-year lag, in the next calendar year (Krąpiec, Margielewski 2000; Krąpiec et al. 2008). Application of the index of tree-growth eccentricity allowed researchers to calculate the intensity of tree reaction to landsliding in particular years (Malik, Wistuba 2012; Wistuba et al. 2013). This provides an opportunity to compare annual/semi-annual precipitation totals with the reaction of trees to landsliding (Wistuba et al. 2013).

The objective of the presented study was to examine the relation between the occurrence of different types of precipitation with totals calculated for various periods (annual precipitation, summer half-year precipitation, preceding winter half-year precipitation, maximum daily, 3-, 5-, 10- and 30-day cumulative rolling totals, number of days with daily precipitation exceeding 20 mm) and the reaction of the trees to landsliding on the slope of the Kamień massif (Beskid Niski Mts) expressed as an eccentricity index.

Study Area

The study was carried out on an eastern slope of the Kamień massif (714 m a.s.l.), in the Beskid Niski Mts (Outer Western Carpathians in Poland) (Fig. 1A, B). We have selected two sampling sites: the study site on a landslide slope, and the reference site on a stable slope, 1 km from the landslide slope. Bedrock of the Beskid Niski Mts is composed of flysch sandstones and shales. On the studied slope of the Kamień massif there is a stratigraphic border between sandstones, shales, claystones and cherts (lower part of the studied slope) and series of predominating sandstones interbedded with thin layers of claystones and mudstones (upper part of the studied slope). This provides favourable conditions for development of landslides.

The main scarp of the studied landslide is located near the ridge of the Kamień massif, approx. 650 m a.s.l. The main scarp is 25–35 m high, the length of the landslide is 1.4 km, the maximum width is 360 m and the area 35 ha. In the middle part typical landslide forms are common: hummocky relief and open landslide cracks. The accumulation zone is composed of colluvial blocks and descends to approx. 450 m a.s.l. (Fig. 1C). According to the modified classification by Varnes (Hungr et al. 2014) the analysed landslide is probably a compound slide with very low velocity of movement and a highly disturbed flow-like lower part of the landslide toe. Hummocky topography but without clear landslide scarps was observed also above the main scarp, up to the top of Kamień mountain ridge (Fig. 1).

Climate conditions in the study area are typical for lower mountain ranges of the Carpathians with precipitation of c. 800–900 mm per year. Monthly mean temperatures vary from -3°C in January to 17°C in July. Snow cover lasts 80–100 days per year. The study area is located in Magura National Park. Natural beech and fir mixed forests cover the area of the study. There are no large urban or industrial areas located less than 20 km from Kamień Mt. However, the study area is located within a former Central Industrial District (CID) which flourished in Poland in 1920–1940. After the 2nd World War CID begin to decline but other industrial districts, located further away from the studied landslide, developed, emitted harmful air pollution to the atmosphere from 1960 to 1990 and affected tree growth in southern Poland (Malik et al. 2012). After 1990 environmentally friendly technologies were introduced and air pollution decreased.

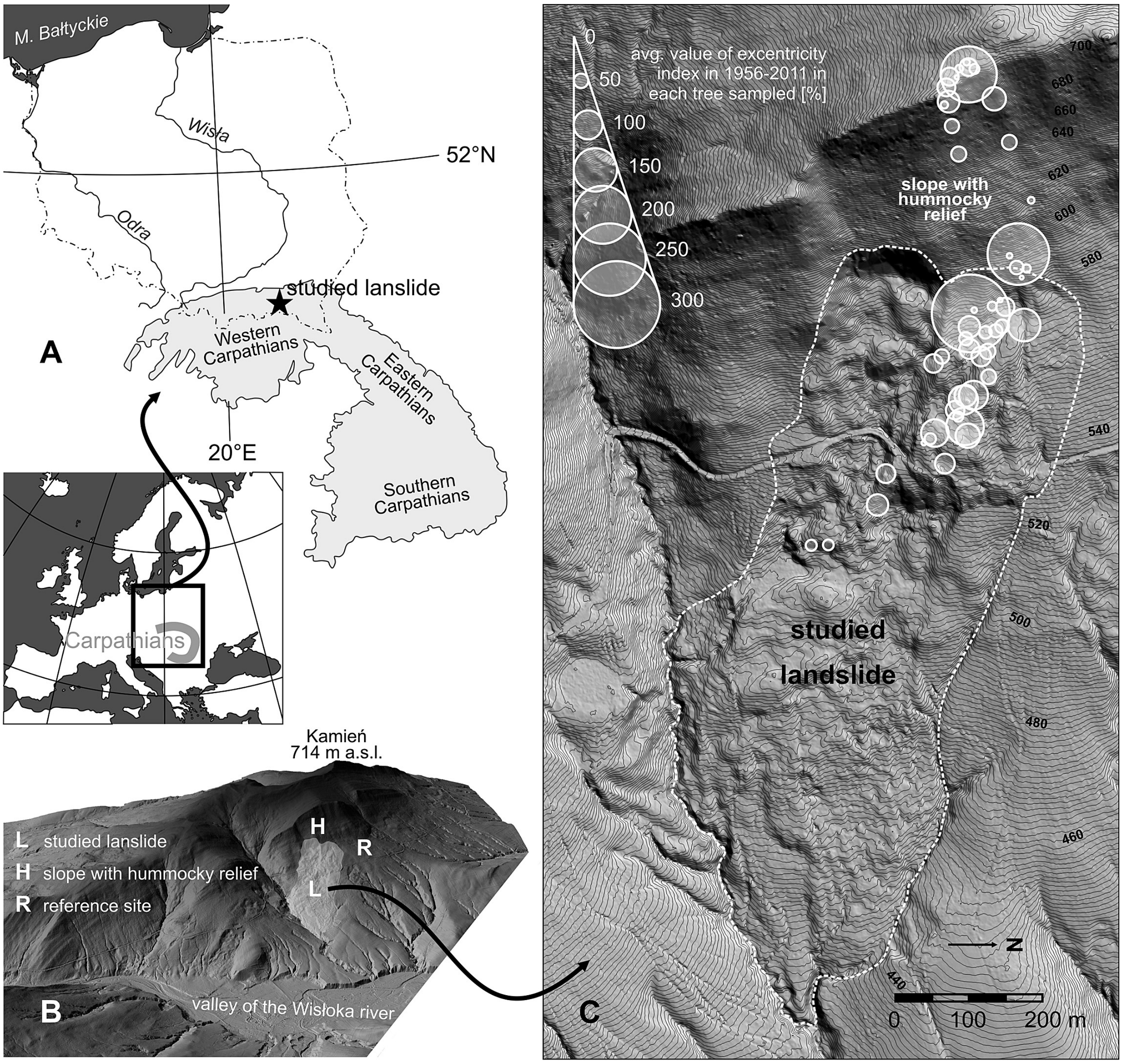

Fig. 1. A – location of the studied landslide in Poland and in the Carpathians; B – relief of the Kamień massif; C – relief of the studied landslide with the location of sampled trees and their avg. reaction to landsliding in 1957–2012 (digital terrain model of the studied landslide with contours each 1 m in m a.s.l.)

Methods of Study

Dendrochronological analysis of landslide activity

On the studied slope 48 silver firs (*Abies alba*) were sampled (Fig. 1C). We collected samples from trees which grow on or near landforms such as cracks, colluvial blocks, and the main scarp of the landslide. We also sampled trees above the main scarp, on the hummocky slope, up to the mountain ridge. Additionally 10 firs were sampled on a reference slope. We took two cores from each tree using a Pressler borer: one from the upslope side and the second from the downslope side of a stem. Cores were glued into wooden stands and polished in order to reveal the wood structure. Tree-ring widths were measured with 0.01 mm accuracy using the LinTAB measuring system.

Cores from the upslope and downslope side of each stem were compared (Fig. 2). An eccentricity index and its annual variation were calculated (Malik, Wistuba 2012; Wistuba et al. 2013). Using reference thresholds obtained on the reference slope (average level of eccentricity in 10 trees sampled on a stable slope) we dated events of landslide activity. An example of the application of the eccentricity index method in dating landsliding is shown in Figure 2. The percentage of trees showing a reaction to landsliding was calculated for each year of the studied period (1957–2012) as an indicator of general landslide activity of the studied slope.

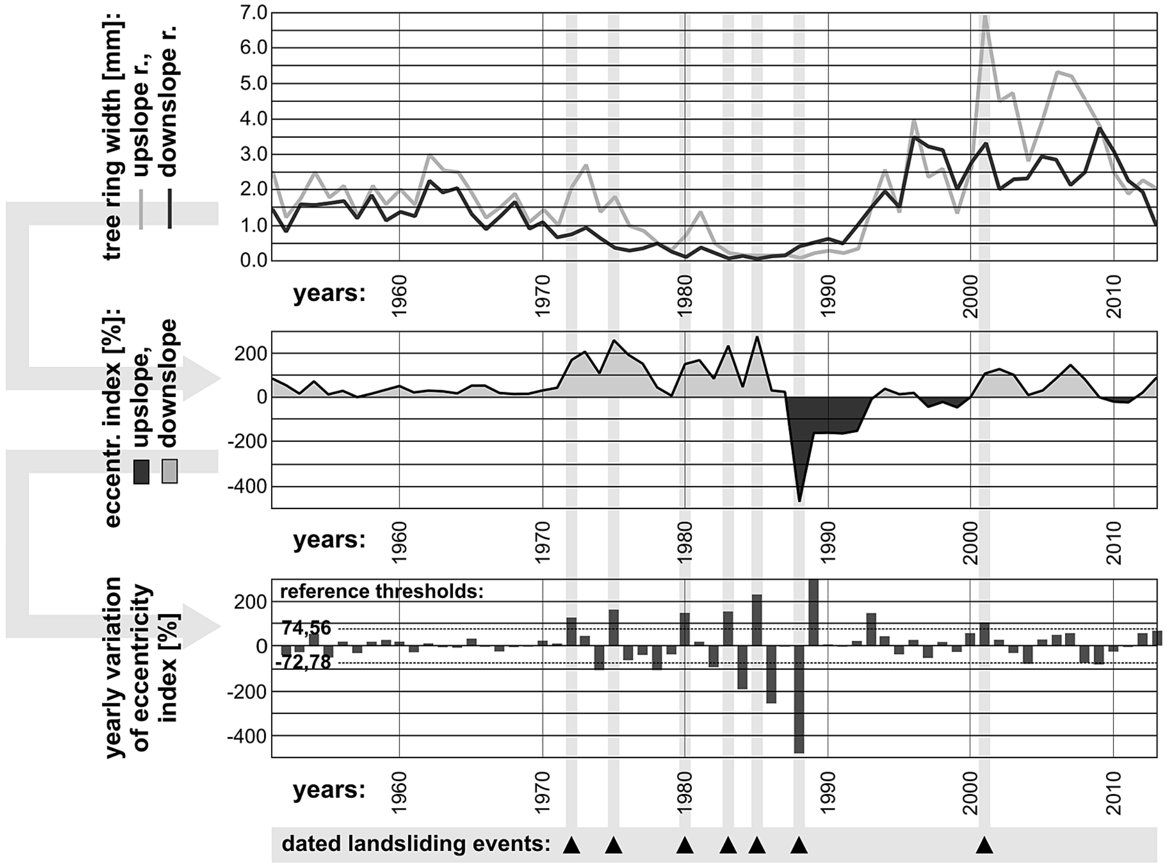

Fig. 2. Transformation of tree-ring widths into eccentricity index and its yearly variation with dated landsliding events (an example of tree growing on the studied landslide)

Comparison of the dendrochronological record of landsliding and the precipitation record

To determine the role of precipitation as a trigger of landsliding on the studied slope we have used data from the precipitation posts in Barwinek (16.8 km SE of the studied landslide, 450 m a.s.l.) and Krempna (3.7 km SW, 380 m a.s.l.). These data sets were selected because the two stations have the longest measurement periods (since 1956 in Barwinek) and the closest distance to the studied landslide (Krempna). Data from other precipitation posts in Nowy Żmigród (7.3 km N, 310 m a.s.l.), Dukla (11.7 km E, 325 m a.s.l.) and Wyszowatka (12.3 km SW, 505 m a.s.l.) were also used for the purpose of comparing the precipitation data and assessing their reliability. For all five posts a daily precipitation record was available for different periods (Barwinek: 1956–2012, Krempna: 1975–2012, Nowy Żmigród: 1985–

–1994, Dukla: 1986–2012, Wyszowatka: 1981–2012, all data from the Institute of Meteorology and Water Management – National Research Institute).

The values of the correlation coefficients between the data from all 5 posts in 1986-1989 (the only period fully covered in all posts) range from 0.67 to 0.86 (Table 1) and are all statistically significant. The data set for the Barwinek post was incomplete with gaps for some single months in 1990–1995. Missing data were supplemented using the quotient stability method (Pruchnicki 1987; Tuomenvirta 2001) based on the data from the other four precipitation posts, with priority for Krempna, showing the highest correlation coefficient with Barwinek.

Table 1. Correlation coefficients for examined records of precipitation in 1986-1989 in five posts surrounding the studied landslide.

Precipitation post	Barwinek	Krempna	Nowy Żmigród	Wyszowatka	Dukla
Barwinek	x	0.73	0.67	0.70	0.68
Krempna	0.73	x	0.84	0.86	0.86
Nowy Żmigród	0.67	0.83	x	0.72	0.86
Wyszowatka	0.70	0.86	0.72	x	0.70
Dukla	0.68	0.86	0.86	0.70	x

The fact that precipitation records from posts surrounding the studied landslide are similar and correlate well one with another suggests that two selected data sets from Barwinek and Krempna can be used in further analyses of landsliding as reliable records, representative for the whole area and for the studied landslide. It was found that particular precipitation events recorded with daily resolution at five studied posts are clearly overlapping (Fig. 3). We have, however, analysed a detailed example of the daily precipitation record for the year 1987 at all five posts and found some differences between particular posts. The differences can result from the complex spatial distribution of precipitation in mountain areas influenced by such local features as topography, elevation a.s.l., slope aspect and by the local range of precipitation from singular storm units. Despite the above-mentioned we have found clear consistency of precipitation records from five posts surrounding the studied landslide. This allows reliable comparison of the dendrochronological record of landsliding with the precipitation record.

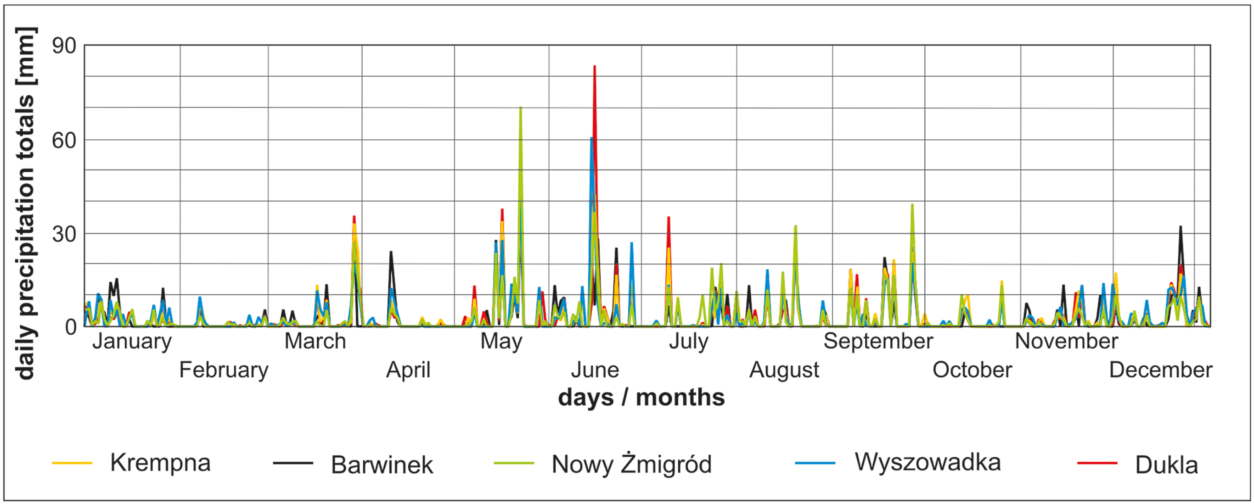

Fig. 3. Daily precipitation totals in 1987 at five precipitation posts surrounding the studied landslide

Results of the dendrochronological analysis were compared with precipitation totals for 1957–2012 in the case of the Barwinek post and 1975–2012 in the case of Krempna. We have analysed 3- 5-, 10- and 30-day cumulative rolling totals of precipitation and the number of days with daily precipitation exceeding 20 mm (Table 2). The analysis was conducted separately for the summer half-year (April-September: the vegetation period when the annual tree ring is developed) and for the preceding winter half-year (October-March, the dormant season preceding development of the annual tree ring) since landsliding active at the end of the vegetation period and during the dormant season can be recorded in the ring developed in the following year. To determine the relation between precipitation and landsliding recorded in tree-ring eccentricity we have calculated the Pearson correlation coefficient for the dendrochronological record of landsliding and the following precipitation indicators for Krempna and Barwinek posts:

- the highest daily, 3-, 5-, 10- and 30-day cumulated rolling precipitation totals during summer and preceding winter half-years,
- precipitation totals for summer and preceding winter half-years,
- annual precipitation totals,
- the number of days with daily precipitation exceeding 20 mm during summer and preceding winter half-years.

Table 2. Correlation coefficients for the activity of the studied landslide and precipitation data from Barwinek and Krempna posts: >20 – number of days with daily precipitation totals exceeding 20 mm; 3, 5, 10, 30 – maximum 3-, 5-, 10-, 30-day cumulative rolling precipitation totals; C – precipitation total for given period (preceding winter, summer); S – levels of statistical significance

Precipitation post	Pearson correlation coefficients for particular types of precipitation data													S
	Preceding winter half-year						Summer half-year						Year	
	>20	3	5	10	30	C	>20	3	5	10	30	C	C	
Barwinek 1957–2012	0.21	0.06	0.11	0.14	0.06	0.04	-0.16	-0.12	-0.14	-0.13	-0.18	-0.11	-0.26	0.26
Barwinek 1980–2012	0.29	0.10	0.18	0.23	0.00	0.07	-0.08	-0.21	-0.18	-0.04	-0.12	-0.13	-0.21	0.34
Krempna 1975–2012	0.31	0.24	0.28	0.26	0.01	0.18	0.04	-0.17	-0.23	-0.13	-0.08	-0.03	0.20	0.32
Krempna 1980–2012	**0.42**	**0.34**	**0.37**	**0.38**	0.09	0.25	0.14	-0.11	-0.19	-0.07	-0.07	0.02	-0.04	0.34

bold – statistically significant correlations

We have checked the statistical significance for calculated correlation coefficients.

Previous studies have shown that industrial air pollution strongly influenced tree growth in Poland and the whole of Europe (e.g. Schweingruber et al. 1985; Elling et al. 2009). High emissions of harmful compounds into the atmosphere resulted in significant reduction in tree-ring widths (Danek 2007) (Fig. 2). Trees frequently formed only single rows of cells instead of full growth rings (Malik et al. 2012). Some studies suggest that in the case of such narrow rings the reaction of trees to tilting can be supressed and growth eccentricity may not be developed despite active landsliding (Malik et al. *in press*). Because in southern Poland industrial pollution resulted in particularly supressed growth prior to the 1980s we have analysed tree rings developed after 1980 separately (Table 2).

Results and Discussion

Dendrochronological record of landsliding on the slope of the Kamień massif in 1957–2012

The highest average reaction of trees to landsliding in 1957–2012 was observed in the central part of the studied slope, but sampled trees also recorded some landslide events in both the upper and lower part of the studied slope (Fig. 1C). The change in landslide activity over time was analysed as the percentage of trees reacting to landsliding in the sampled population in particular years of the studied period. The highest percentage of trees showing a reaction to landsliding was recorded in 1957, 1986 and 1987 (27% in each year). Significant landslide activity (≥24% of sampled trees) was also recorded in 1975, 1977, 1988 and 1996. Some weaker landslide events (≥ 15% of sampled trees) were also recorded in half of the analysed years (Figs 4–6).

The results obtained suggest that the studied landslide was inactive and stable in 1958–1959, 1964, 1967–1968, 1974, 1976, 1998, 2004, 2005 and 2011, as proven by the lack or low level of a dendrochronological record of eccentricity in these years. The examined landslide shows high variability of activity over time. No longer periods of active landsliding and stability were recorded. Several-year-long periods of increased activity are separated by short periods without any significant tree-ring record of slope movement.

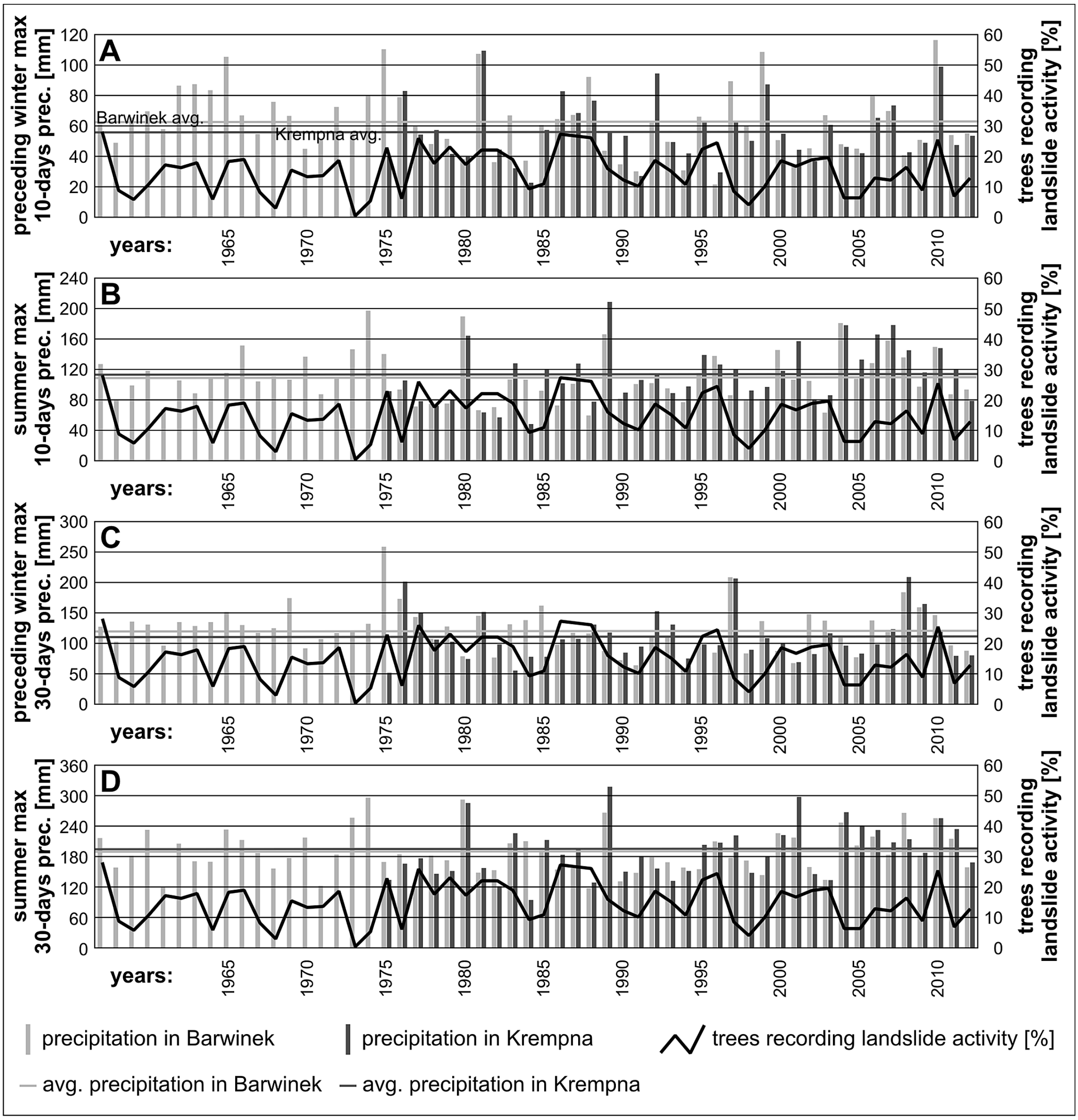

Fig. 4. Dendrochronological record of landsliding on the studied slope in 1957–2012 compared with various precipitation data from Barwinek and Krempna posts: A – landslide activity and maximum 10-day precipitation totals during preceding winter half-year; B – landslide activity and maximum 10-day precipitation totals during summer half-year; C – landslide activity and maximum 30-day precipitation totals during preceding winter half-year; D – landslide activity and maximum 30-day precipitation totals during summer half-year

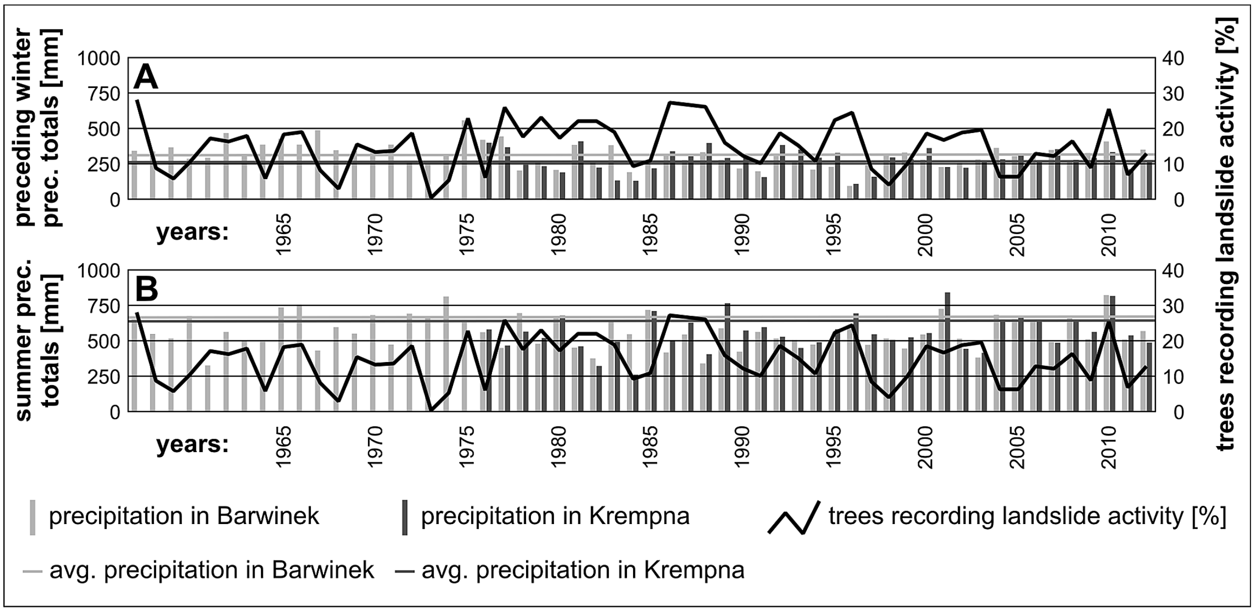

Fig. 5. Dendrochronological record of landsliding on the studied slope in 1957–2012 compared with various precipitation data from Barwinek and Krempna posts: A – landslide activity and precipitation totals during preceding winter half-year; B – landslide activity and precipitation totals during summer half-year

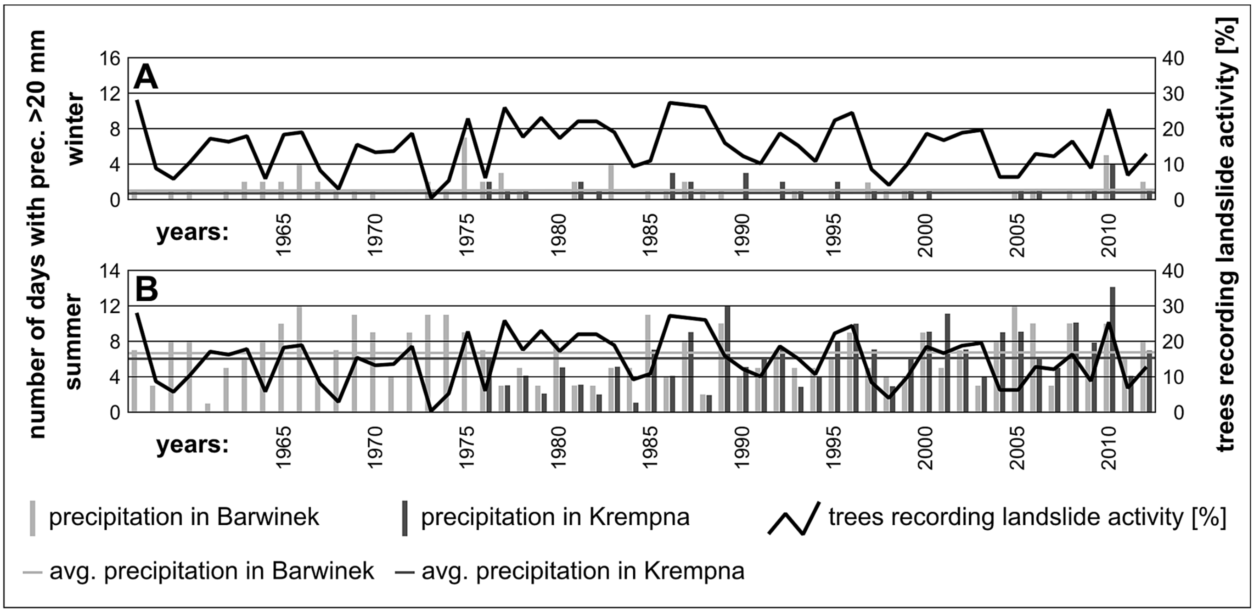

Fig. 6. Dendrochronological record of landsliding on the studied slope in 1957–2012 compared with various precipitation data from Barwinek and Krempna posts: A – landslide activity and number of days with daily precipitation totals > 20 mm during preceding winter half-year; B – landslide activity and number of days with daily precipitation totals > 20 mm during summer half-year

Comparison of tree reaction to landsliding on the studied slope of the Kamień massif and the precipitation record

Most of the analysed records of precipitation do not show any statistical correlation with the course of the landslide process on the studied slope. There is a slight correlation between the dendrochronological record of landsliding and precipitation of the preceding winter half-years (October-March) in Krempna in 1980–2012. No correlation was found for the annual totals and precipitation of summer half-years (Table 2, Fig. 5B). Results suggest that the activity of the landslide depends mainly on the amount of precipitation during the previous winter. Also Ziętara

(1968) suggests that catastrophic landsliding is associated with spring rather than with summer precipitation. Gorczyca (2008) suggests that factors and conditions triggering development of landslides are complex including both spring thaws, spring and summer precipitation, long-term rainfalls and short, torrential downpours. Also Starkel (2014) pointed out that landslides in 2006 in the foothills of the Carpathians and in the Moravian Carpathians may have been triggered by snow thaws.

The results obtained, which suggest the importance of winter precipitation in triggering landslide activity, are also compliant with investigations by Gil et al. (2009), who compared data from inclinometers with amounts of precipitation of summer and winter in 2007–2009. Studies were conducted in the Beskid Niski Mts, where the Kamień massif is also located. Gil et al. (2009) found that colluvial bodies of landslides were moving mainly during winter half-years due to high rates of precipitation in winter and melting of the snow cover.

Excluding data from before 1980 from the correlation increased the values of correlation coefficients obtained, in the case of the 3-, 5- and 10-day precipitation totals during preceding winters in Krempna even above the level of statistical significance (Table 2). Thus it seems that growth suppression in the period of air pollution (Fig. 2) concealed the reaction of trees to landsliding – tree-ring eccentricity was not as clear as later. A similar phenomenon was observed by Opała and Mendecki (2014) in a dendroclimatic study conducted for the Silesian Upland. They have also excluded the 1960–1980 period from the analyses since the dendrochronological signal was mainly a reflection of the atmosphere pollution changes and not climatic changes (Opała, Mendecki 2014).

The strongest events of landsliding and events of slope stability recorded in tree rings in the Kamień massif compared with precipitation data

Because of the complex nature of landslide triggers and the lack of an unambiguous relation between landsliding events on the studied slope of the Kamień massif and the precipitation record we have analysed precipitation conditions in single years with a particularly strong dendrochronological record of landslide activity.

In 1957 when 27% of sampled trees recorded landsliding there were two relatively strong precipitation events (data from Barwinek post): at the beginning of May (c. 150 mm in 10 days) and in mid-July (c. 200 mm in 20 days). The summer half-year was humid (90 mm above the 1957-2012 average), whereas the precipitation of the preceding winter was average (341 mm, 31 mm above the average). However on May 13th 1957 an earthquake occurred (Rączkowski 2007). The role of the 1957 earthquake in triggering the landslide in Lipownica near Dukla (11 km E of the Kamień massif) was described by Gerlach et al. (1958). By contrast, Gil and Długosz (2006) indicated that landsliding in 1957 when an earthquake took place could have been triggered by rainfall and snowfall exceeding average values and by an intensive thaw at the turn of April and May.

In 1977 landsliding was recorded by 26% of sampled trees. According to precipitation data from Barwinek and Krempna it was quite a dry year. Precipitation during the summer half-year term did not exceed the average values at both posts. The previous winter was, however, wet, with precipitation exceeding average values. The maximum 30-day precipitation total was 143 mm in Barwinek and 149 mm in Krempna (Fig. 4C), compared to average values: 121 mm and 111 mm, respectively. Precipitation totals of the preceding winter half-year in 1977 were 442 mm (average 311 mm in the Barwinek post) and 366 mm (average 275 mm in the Krempna post) (Fig. 5A). High precipitation amounts were also recorded in previous years. In 1975, when landslide activity was slightly lower (reaction of 24% of sampled trees) precipitation rates were the highest during the whole period of 1957–2012 (Figs 4–6). Total precipitation one year earlier, in 1974, reached 1382.7 mm, 296.2 mm of which occurred during 30 consecutive days in summer (rainfall total of July: 282.2 mm). There were 11 days with daily precipitation exceeding 20 mm (compare with 6.6 on avg.) during the 1974 summer half year and 7 such days during the 1974/75 winter half year (1.1 on avg.) (Fig. 6). The precipitation total of the 1974/75 winter half-year was 555 mm (with 310 mm on avg.) with maximum 10- and 30-day precipitation totals also twice as high as average values (Fig. 4). The landslide activity dated to

1975 has, therefore, a substantiated background of extreme precipitation in July 1974 and a very humid preceding winter term. We assume that the debris mantle and bedrock were oversaturated with water for many months, which resulted in intensified landsliding in 1975. The long-lasting increased precipitation rates in the subsequent years explain landslide activity recorded in 1977. Also some previous studies indicate the complicated nature of precipitation as a factor triggering landslide activity, due to its temporal and spatial variability and its complex effect on ground water level (Gil, Słupik 1972; Starkel 1976, 2011).

A similar longer landslide event took place in the years 1986–1988 when 26–27% of trees showed a reaction to landsliding. This period was characterised by average annual precipitation totals, low precipitation totals of summer half years in relation to the average value as well as increased precipitation during previous winter half-years. Landsliding was also preceded by high amounts of precipitation in the summer of 1985. Thus, probably similarly to 1975, the wet period started during the preceding summer. Since 1980 the correlation coefficient between dendrochronologically recorded landsliding and maximum 3-, 5-, 10-day precipitation totals in Krempna is statistically important (Table 2). Therefore, it seems that in the discussed years 1986-88 the precipitation from the preceding winter could have been the main triggering factor for landslide activity. The 10-day precipitation totals of preceding winters were increased for the whole 1986–88 period, reaching 82.6 mm in 1986 in Krempna, and 91.9 mm in 1988 in Barwinek. Winter half-year precipitation totals reached 302.6 mm in 1987 and 396.4 mm in 1988 in Krempna.

In 1996 when 24% of sampled trees recorded landsliding high precipitation occurred in July (monthly total: >150 mm) and September (>200 mm), and the maximum daily rainfall total reached 78 mm. There is no proof that the landslide event can be connected to precipitation of the preceding winter, which, according to gathered data from Krempna and Barwinek, was at the average level or below it (Fig. 5A). It seems that landslide activity in 1996 resulted mainly from increased summer precipitation.

The year 2010 was extraordinarily wet in the whole of the Western Carpathians. The tree-ring record of landsliding on the studied slope in 2010 is strong. 26% of sampled trees showed a reaction to slope movements then. A very high amount of precipitation during the summer half-year occurred along with some events of long-term downpours in May-June (100–150 mm each). Heavy precipitation occurred also in the second half of June. The preceding 2009–2010 winter was also distinctly wet. Very high short-term precipitation totals and numerous days with daily precipitation exceeding 20 mm were recorded both during the summer in 2010 and the previous winter season (Fig. 6); also the maximum 10-day precipitation totals significantly exceeded the average value (Fig. 4A, B).

Also the periods of slope stability dated by means of dendrochronology were compared with the precipitation record. After the landsliding event in 1957, in 1958–1959 the tree-ring data obtained suggest that the studied slope was stable (only 6–9% of sampled trees recorded slope instability). The period was characterised by average precipitation, both during the summer and the preceding winter half-years. Analysed indicators of precipitation were lower than the average; only the longer-term totals were slightly higher (Figs 4–6). The lack of strong precipitation probably resulted in stabilisation of the studied landslide.

The studied slope of the Kamień massif was also stable in 1967–68 (only 3–8% of sampled trees recorded landsliding) and 1973–1974 (0–5%). The first of the above-mentioned periods was relatively wet, both in terms of the winter and summer half-year precipitation. Still, the preceding years when the landslide was highly active were even more humid. In 1974 extremely high precipitation occurred in the summer (Fig. 4B, D). Precipitation of the summer half-year reached 813 mm (compared to 562 mm on avg.) (Fig. 5B). Despite such high precipitation amounts in summer the dendrochronological record of landsliding in 1974 is poor (5% of sampled trees show a reaction to slope instability). This may result from very poor precipitation of the winter half-years during the period (265.4 mm for the winter preceding 1973 and 243.5 mm for 1974).

In the 1980s and 1990s there was a significant increase in the activity of the studied landslide. In the period no long-term, distinct stability of the slope was recorded. Only in 1998 was the reaction of the trees poor (4% of sampled trees recorded landsliding). In this year precipitation both in summer and preceding winter half-years was average (Figs 4 and 5). In the Beskid Niski Mts, opposite to areas of the Carpathian mountain belt located more to the west, the year 1997 was also dry (annual precipitation total lower than avg. by 170 mm in Barwinek and 80 mm in Krempna). In 1997 landsliding on the studied slope was recorded only by 4 trees (9% of the sampled population). Similarly in 2004 and 2005 landslide activity was low (6% of trees recorded landslides). These were humid years, with a particularly high share of the summer term precipitation, but also with particularly low 10-day precipitation totals during winter half-years (Fig. 4A). Also tree-ring data obtained for the year 2011 indicate slope stability in the period (7% of sampled trees showing a reaction to landsliding). The reason is probably the fact that the 2010/2011 winter season was particularly dry (Fig. 5). The analysis suggests that the stability periods of the examined landslide body are related to the relatively low precipitation, especially during the preceding winter term. Still, we cannot exclude that, for example in 2011, the studied landslide was inactive since all the available material had already been dislocated during the strong event in 2010.

Conclusions

a) The studied landslide in the Kamień massif is contemporarily active showing high frequency of small-scale movements of colluvia. In 1957–2012 movements of the landslide body occurred almost every year but with different strength. In particular years landsliding occurred in different parts of the studied slope. Increased landslide activity was recorded in tree rings in 1957, 1975, 1977, 1986–1988 and 1996.

b) Neither the importance of high precipitation totals during the whole summer half-year nor intensive short term precipitation occurring during the summer appeared to be significant as factors triggering slope instability in the Kamień massif. Still, in the case of several landslide events they have probably played some role (e.g. 1957, 1996, 2010). Precipitation of the preceding winter half-year seems to be the most important factor in triggering landslide activity on the study slope, despite the fact that the precipitation totals of winter seasons are generally lower than during summer. During winter seasons reduced evaporation, water retention in snow cover and mid-winter and spring thaws cause saturation of bedrock with water, which becomes more susceptible to landsliding.

c) The nature of the relation between precipitation and landsliding is complex. The strongest landsliding on the studied slope resulted from sequences of wet summer, wet winter and once again wet summer half-years directly following one another. Long-term saturation of the ground with water (lasting several months) can decrease the threshold of precipitation necessary for finally triggering landslide activity. Therefore long-term rainfalls are important as factors preparing a slope for landslide events. The analysis additionally suggests that also earthquakes may have been a factor triggering the instability of the studied slope.

d) Analysis of the relations between the dendrochronological record of landsliding and precipitation occurrence have revealed some difficulties associated with the annual record of eccentricity. We have found large reductions in ring widths caused by harmful air pollution in 1960–1985. Strong reductions overlap eccentric growth caused by landsliding. This may limit the possibility of dating landslide activity from dendrochronological records.

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