Chapter 5
Recent Landform Evolution in the Polish Carpathians

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Abstract The key drivers of relief transformation in the Polish Carpathians (the best studied region of the Carpathians) have been local downpours, continuous rainfalls, and rapid snowmelt. Threshold values are quickly exceeded during such events and powerful morphological processes are initiated. However, due to human activities, geomorphic processes are often accelerated and intensified with serious consequences. Thus, increased human impact is a key factor in the recent geomorphic evolution of the Polish Carpathians. Over the last two centuries deforestation, intensive agriculture, mining, housing developments on slopes, channelization of streams, and construction of reservoirs all have contributed to changes in the rate of the particular geomorphic processes. The intensity of sheet erosion depends on vegetation cover, land use, and cultivation techniques (terracing, contour tillage, crop rotation, etc.). Slopes bearing poorly constructed infrastructure have become susceptible to mass movements. At higher elevations debris flows, dirty avalanches,

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and extreme floods are crucial in slope evolution. Over the last 20 years the ratio of agricultural land and the rate of sheet erosion and deflation have decreased, while gully erosion on slopes and the incision of river channels have intensified and the reactivation of shallow landslides has become more common. Increasingly frequent extreme weather events may reverse the stabilizing trend of landform evolution.

Keywords Mass movements • Sheet and rill erosion • Ephemeral gully erosion • Land use • Piping • Aeolian processes • Periglacial processes • Karst • Fluvial processes • Biogenic processes • Human impact • Polish Carpathians

5.1 Introduction

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The Polish Carpathians are the best-studied area in the entire Carpatho-Balkan-Dinaric Region. Geomorphic processes and landforms here have been the subject of intensive research since the beginning of the twentieth century (Margielewski et al. 2008; Rączkowska 2008).

The first studies of mass movements in the Outer Carpathians dealt with single *landslides* (Zuber and Blauth 1907; Łoziński 1909; Sawicki 1917). Complex landslide analysis in selected Carpathian regions began during the interwar period (Świderski 1932; Teisseyre 1929, 1936). In the second half of the twentieth century a novel attitude to mass movements emerged and researchers began to accept the view that mass movements effectively shaped the surface of the Carpathians during the Late Glacial and the Holocene (Flis 1958; Starkel 1960; Ziętara 1964). Research also addressed contemporary landslides as fundamental geomorphic processes and their negative impact on human economic activities (Jakubowski 1965; Ziętara 1968; Mrozek et al. 2000; Bajgier-Kowalska 2004; Gorczyca 2004). The Polish Geological Institute has been recording landslides in the Polish Carpathians for years (Poprawa and Rączkowski 2003; Rączkowski 2007). Researchers of the Research Station of the Institute of Geography and Spatial Organization, Polish Academy of Sciences (IGiPZ PAN), in Szymbark (Beskid Niski Mountains), identified precipitation thresholds that trigger mass movements (Gil 1997; Gil and Starkel 1979; Starkel 1996). In addition, selected Carpathian landforms are being closely monitored and allow researchers to identify complex hydrogeological triggering mechanisms and to predict future occurrences of landslides (Zabuski 2004).

Research on gravitational and other geomorphic processes in the Tatra Mountains began in the 1950s and led to a better understanding of the morphodynamics of high mountain slopes (Kotarba 1976, 1992; Kłapa 1980; Kotarba et al. 1983, 1987; Rączkowska 1995, 1999). *Debris flows* were added to research topics in the late

1980s with the identification of precipitation thresholds and changes of debris flow activity, especially for the past several centuries (Krzemień 1988; Kotarba 1992, 1997, 2004, 2005).

In the 1960s measurements of *sheet erosion* using special traps (Gerlach 1966, 1976b) as well as of *splash* (Gerlach 1976a) were performed in the Beskid Sądecki and Gorce Mountains, and both surface runoff (Słupik 1970) and sheet erosion (Gil 1976) were monitored at the IGiPZ PAN research station in Szymbark. Thus, the station, located on the boundary line between the Beskid Niski Mountains and the Ciężkowice Foothills, possesses the longest sheet and *rill erosion* measurement record on plots for the Polish Carpathians, continuous since 1968 (Gil and Słupik 1972; Słupik 1973, 1981; Gil 1976, 1986, 1994, 1999, 2009). Simultaneous measurements of soil splash, wash, and supply to fluvial transport were carried out at the IGiPZ PAN research station in Frycowa located in Beskid Sądecki Mountains (Froehlich and Słupik 1980b; Froehlich 1986), and splash and sheet wash were studied for different types of crops in the Beskid Niski Mountains (Śmietana 1987). Field measurements of splash, sheet, rill, and ephemeral gully erosion are also performed in the Wiśnicz Foothills (eastern part of the Wieliczka Foothills) as part of a research program run by the Institute of Geography and Spatial Management, Jagiellonian University (IGiGP UJ), at the research station in Łazy, near the town of Bochnia (Święchowicz 2002a, b, 2008, 2009, 2010).

The effects of *aeolian deposition* on snow were first studied in the Tatras (Kłapa 1963). Quantitative analyses of aeolian processes have been performed in the Jasło-Sanok Depression (Gerlach and Koszarski 1968, 1969; Gerlach 1986), the Western Bieszczady Mountains (Pękala 1969), the Beskid Niski Mountains (Janiga 1971, 1975), the Tatras (Izmaiłow 1984a, b), the Wiśnicz Foothills (Izmaiłow 1995a, b, c), as well as in the boundary zone between the Beskid Niski Mountains and the Ciężkowice Foothills (Welc 1977).

Less attention has been devoted to *piping*. The studies involved the mapping of landforms and the description of relief evolution, primarily in the Outer Eastern Carpathians in Poland (Czeppe 1960; Starkel 1965; Galarowski 1976).

Periglacial processes have been investigated in the Polish Tatra Mountains since the 1950s (Jahn 1958; Gerlach 1959; Kotarba 1976; Dobiński 1997; Gądek and Kędzia 2008; Gądek et al. 2009). This includes research performed at the IGiPZ PAN research station at Hala Gąsienicowa (Kłapa 1970, 1980; Rączkowska 1992, 1993, 1995, 1999, 2007; Kędzia et al. 1998; Mościcki and Kędzia 2001), where processes affecting slope evolution, especially nivation and solifluction, have been studied.

Large-scale research efforts on *fluvial processes* and landforms began in the 1960s and 1970s and have focused on the structure and dynamics of stream channel features (Kaszowski and Kotarba 1967; Kaszowski 1973; Niemirowski 1974; Kaszowski et al. 1976; Kaszowski and Krzemień 1977, 1979; Klimek 1979, 1987; Baumgart-Kotarba 1983; Krzemień 1984a, b, 1991) in light of the role of floods (Ziętara 1968; Froehlich 1975; Krzemień 1985; Malarz 2002; Izmaiłow et al. 2006) and river engineering (Wyżga 1993a, b; Krzemień 2003; Kościelniak 2004; Zawiejska and Krzemień 2004). Particular attention has been paid to the development of fluvial landforms affected by dams (Klimek et al. 1990; Łajczak 1996, 2006). The influence of accumulations of large woody debris is also being investigated with respect to stream channel evolution (Kaczka 2003, 2008; Wyżga et al. 2003). The Holocene evolution of stream channels and floodplains is also in the focus of research (Starkel 1965, 2001; Klimek and Trafas 1972; Klimek 1974; Froehlich et al. 1977). Fluvial geomorphologists are particularly interested in determining the rates of contemporary fluvial processes (Froehlich 1975, 1982; Krzemień 1984a, 1991; Krzemień and Sobiecki 1998; Łajczak 1999; Święchowicz 2002c; Wyżga et al. 2003).

Research on the formation and development of large *biogenic landforms* began during the second half of the twentieth century. Peat domes (Ralska-Jasiewiczowa 1972, 1980, 1989; Wójcikiewicz 1979; Żurek 1983, 1987; Obidowicz 1990) and human impact on the formation of peat bogs (Łajczak 2007b, 2011) are investigated in the Orawsko–Nowotarska Basin and in the Bieszczady Mountains.

5.2 Geological and Geomorphological Settings

Włodzimierz Margielewski

The Polish Carpathians is the northernmost section of the Alpine mountain belt. It is geologically divided into the *Inner (Central) Carpathians* (including the Tatras and the Podhale Basin) and the Outer (Flysch) Carpathians, separated by the Pieniny Klippen Belt (Fig. 5.1) (Książkiewicz 1972; Birkenmajer 1986; Oszczypko et al. 2006).

The *Tatra Mountains* are composed of an extensive crystalline core, including a Permian-Mesozoic sediment layer, the High-Tatric Nappe, and the north-verging

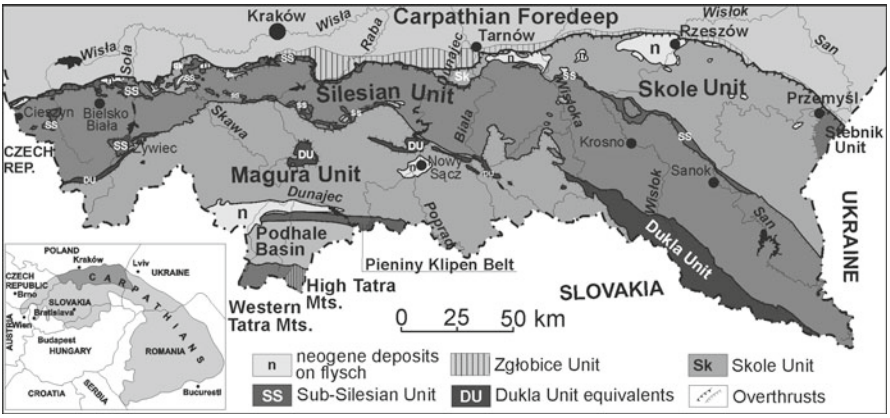


Fig. 5.1 Geological map of the Polish Carpathians (After Żytko et al. 1989)

Subtatric nappes (Książkiewicz 1972). The crystalline core is formed of intrusive Carboniferous granitoids of the High Tatras as well as metamorphic rocks (Paleozoic rocks: gneiss, amphibolite, metamorphic shale) found mainly in the Western Tatras. The crystalline core is unevenly covered with epicontinental autochthonous High-Tatra sediments such as quartzites, dolomites, limestones, marls, shales, conglomerates, and sandstones formed from the Triassic to the Middle Cretaceous. These High-Tatra rock series are covered from the south by the High-Tatric nappes of similar rock complexes. The following Subtatric nappes, formed between the Triassic and the Lower Cretaceous, are superimposed on the High-Tatric Nappe: the lower Križna Nappe, the middle Choč Nappe, and the upper Stražov Nappe. The Triassic formations developed here in a manner similar to that of the High-Tatric series. On the other hand, younger (Jurassic to Lower Cretaceous) formations are deep marine sediments (radiolarites, limestones – Książkiewicz 1972; Oszczypko 1995).

The *Podhale Basin* is filled with sediments of 2,500 m thickness from the Middle Eocene–Oligocene. Middle Eocene conglomerates are covered by nummulitic limestone, dominated by shale-sandstone sediments of the Podhale flysch type. The sediment fill is only slightly deformed and forms a wide syncline. Stronger deformations have only been found near the line of contact with the Pieniny Klippen Belt (Książkiewicz 1972; Oszczypko 1995).

The *Pieniny Klippen Belt* is a narrow geological structure, a suture formed in the collision zone between the North European Plate and the Adria-Alcapa terranes (microplates) (Oszczypko 1995; Oszczypko et al. 2006). Its constituent rocks form steep slide-folds and folds thrust in between the Podhale flysch formations in the south and between Magura-type nappes in the north. They consist of klippen belts built of resistant limestone and radiolarite as well as sheathing rocks such as less-resistant Upper Cretaceous to Paleogene marls, sandstones and shales (Birkenmajer 1986). Andesite intrusions associated with Miocene volcanic activity can be found along the northern edge of the klippen belt, along the line of contact with the Magura Nappe.

The *Outer (Flysch) Carpathians* are built primarily of Late Jurassic to Early Miocene flysch (turbidite) with occasional intercalations of marls and silicates (radiolarite, hornstones) (Fig. 5.1) (Książkiewicz 1972; Oszczypko 1995). Series of folded flysch sediments, disrupted from their substrate and displaced in a generally northerly direction, have formed discrete tectonic units (nappes) superimposed on one another and their foreland. These include the Magura Nappe, the Dukla Nappe (and its equivalents), the Silesian Nappe, the Sub-Silesian Nappe, and the Skole Nappe (Książkiewicz 1972; Oszczypko 1995). The Stebnik (Sambir) Unit in the northeast consists of folded Miocene formations (Fig. 5.1) (Oszczypko 1995; Oszczypko et al. 2006). The northern foreland of the Polish Carpathians contains Miocene Foredeep formations. The Zgłobice Unit is the folded part of these sediments, which appear in a narrow discontinuous belt in the Carpathian foreground (Fig. 5.1). The folded flysch formations of nappes are divided by a system of joints and numerous dislocations, while rock massifs are characterized by lithological variability.

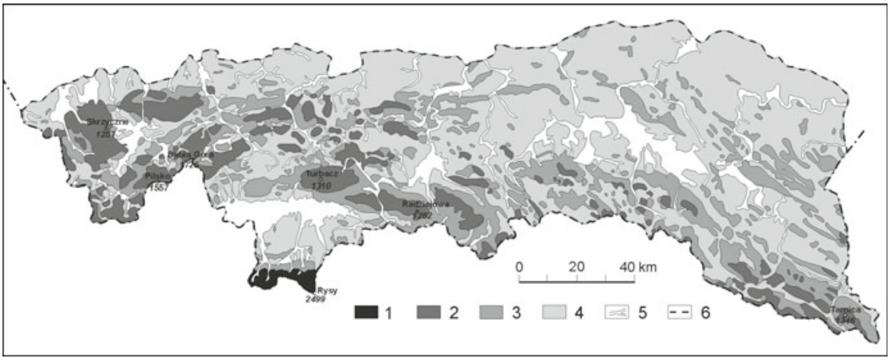


Fig. 5.2 Types of relief in the Carpathians (After Starkel 1980). 1, high mountains; 2, middle mountains; 3, low mountains and high foothills; 4, middle and low foothills; 5, bottom of valleys; 6, boundary of the Carpathians

Fig. 5.3 Geomorphological units of the Polish Carpathians (After Gilewska 1986; Starkel 1991). 1, Central Western Carpathians; 2, Outer Western Carpathians; 3, Outer Eastern Carpathians

The Carpathian relief is strongly linked to the lithology and tectonics of rock massifs and is subdivided into altitudinal belts (Starkel 1969). The Polish Carpathians encompass four dominant types of relief: high, middle and low mountains, foothills, and valley floors (Fig. 5.2 – Starkel 1972, 1991). *High mountains* (1,500–2,499 m a.s.l.) of postglacial and rocky slope type include the Tatra Range and the crest of Babia Góra Massif (1,725 m) in the Flysch Carpathians (Fig. 5.2 – Starkel 1991). *Middle mountains* (800–1500 m) with steep slopes (over 30%) on resistant sandstones include the compact Beskid Śląski, Beskid Żywiecki, and Beskid Sądecki mountain ranges and the isolated Beskid Wyspowy and Bieszczady ridges, ranging from 400 to 800 m in elevation (Figs. 5.2 and 5.3). *Low mountains* usually encircle medium-height mountains in the form of isolated ridges of resistant rocks, with steep slopes and a relative relief of 200–400 m, rising above depressions cut into less resistant rocks (like the Beskid Niski Mountains – Starkel 1991).

A common characteristic of relief in the *foothills* north of the Beskidy Mountains (see Fig. 5.3) is the presence of wide hills of various gradient (10–40%) with convex-concave slopes and flat valley floors. Three types of foothills are distinguished in the Flysch Carpathians: high foothills with crests at 200–300 m, middle foothills with crests at 120–200 m, and low foothills with crests at 40–100 m above valley bottoms, formed in the least resistant of rocks around the Polish Carpathians (Fig. 5.2). Protruding elements such as escarpments, associated with more resistant rocks on the margins of tectonic units, are remarkable features of the foothills in the Flysch Carpathians. The floors of valleys and basins are flat and often terraced (Starkel 1991).

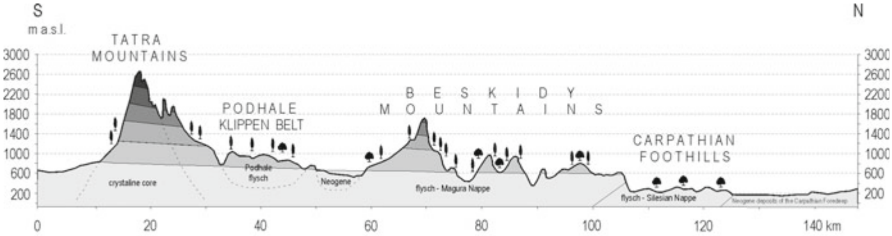
Based on their diverse geological structure and dissected topography, the Polish Carpathians can be divided into three major morphostructural units (sub-provinces): the *Central Western Carpathians* (the Tatras and Podhale) (Fig. 5.3: 1), the *Outer Western Carpathians* (western part of Polish Flysch Carpathians) (Fig. 5.3: 2), and the *Outer Eastern Carpathians* (eastern part of Polish Flysch Carpathians) (Fig. 5.3: 3) (Gilewska 1986; Starkel 1991; Kondracki 2000).

Quaternary formations cover the slopes of the Polish Carpathians and the sides of valleys. They are formed of silt and debris clays from periglacial rock weathering, gravitational rock displacement on slopes, and accumulation in foothill areas (Stupnicka 1960; Cegła 1963; Starkel 1984; Henkiel 1972). In the medium-height mountain zone, the slope deposits on sandstones are relatively thin (70–80 cm) and only thicken out on flat footslopes (Kacprzak 2002–2003). In the foothill zone, the thickness of the slope deposits on flysch schist is greater (3–20 m), especially at lower elevations (Dziewański and Starkel 1967; Starkel 1969; Henkiel 1972). Plenivistulian (Last Glacial Maximum) loess patches, up to a dozen metres thick, locally occur in the foothills (Cegła 1963; Starkel 1984; Gerlach et al. 1991). Deep solifluctional-deluvial mantles cover the Polish Carpathian depressions and basins as a result of sheet erosion, gravitational displacement, and fluvial transport of slope material (Stupnicka 1960; Cegła 1963; Henkiel 1972).

5.3 Climate and Hydrology

Jolanta Święchowicz

A key characteristic of the Carpathian climate is both *vertical* (from 350 to 2,500 m elevation – Fig. 5.4) and horizontal *variability*. Each year maritime-polar and continental-polar air masses predominate in the area over 65% and 20% of the time, respectively. The foothills and the lower sections of the Beskidy Mountains, up to about 700 m elevation, are classified as moderately warm (Hess 1965). The boundary of the moderately warm climate zone decreases about 60 m in the easterly direction. The higher regions of the Beskidy Mountains are moderately cool, while the highest elevations of the Beskid Sądecki, the Gorce, the Beskid Śląski, and the Beskid Żywiecki Mountains fall within the cool climatic belt. The very cool climate zone



	Vertical climate zone	Precipitation			Number of days with snow cover	Number of days with min. temperature <0°C	Wind		
		Yearly totals [mm]	Number of days with precipitation:				Mean velocity [m s⁻¹]	Number of days with wind	
			>1 mm	>10 mm				>10 ms⁻¹	>10 ms⁻¹
	Cold	2050-1250	145-210	45-70	310-270	105-60	5,2-6,8	140-220	55-110
	Temperate cold	2050-1350	155-210	45-65	280-230	120-65	5,1-6,8	135-220	50-110
	Very cool	2100-1350	160-215	40-60	240-195	105-75	4,8-6,7	115-215	40-105
	Cool	2300-1300	145-195	35-60	210-160	110-80	3,2-4,5	45-90	5-45
	Temperate cool	1800-1200	130-185	35-55	175-125	120-70	2,0-3,7	10-60	2-20
	Temperate warm	1350-950	110-150	20-35	140-90	120-60	2,3-2,8	10-35	1-10

Fig. 5.4 Vertical climate zones in the Polish Western Carpathians (After Hess 1965)

is restricted to the Babia Góra Massif above the upper timberline. The Tatra Mountains show five climate zones (Fig. 5.4) with boundaries located 100–200 m higher on southern-facing slopes than on northern-facing slopes.

The average *precipitation* gradient in the Polish Carpathians is 60 mm per 100 m of elevation. The 700 mm isohyet runs along the edge of the Carpathian Foothills, while the 900 mm isohyet encircles the Beskidy Mountains. Strong rain shadow effect is typical of intramontane basins and N-S aligned valleys. Average annual precipitation drops towards the east from over 1,400 mm in the Beskid Śląski Mountains to 1,000–1,300 mm in the Bieszczady Mountains. In the Tatras, maximum annual precipitation is 1,500–1,700 mm (Niedźwiedź and Obrębska-Starklowa 1991; Obrębska-Starklowa et al. 1995). Intense continuous rainfall events are mainly characteristic in June and July and short local downpours in May. The predominance of continuous rainfalls in the west extends to the Dunajec and Biała Dunajcowa valleys (Cebulak 1992). Three types of precipitation determine the type and intensity of slope and fluvial processes: (1) short local downpours (intensity of 1–3 mm min^{-1}), which induce intense sheet erosion and mudflows; (2) continuous rainfalls (150–400 mm per 2–5 days), which lead to landslides, floods, river channel modifications, and sediment deposition on floodplains; (3) rainy seasons with monthly precipitation in the 100–500 mm range, which cause deep infiltration and rockslides (Starkel 1986, 1996; Froehlich and Starkel 1995).

Two *hydrological macroregions* are identified in the Polish Carpathians: the western and the eastern macroregion (Ziemońska 1973; Dobija 1981; Dynowska 1995). The drainage basins west of the Dunajec River are characterized by a rain-snow *precipitation type*, while those east of the river by a snow-rain type. The *river*s of the Polish Carpathians show substantial seasonal and year-to-year variability of *discharge*. Low discharge occurs in the autumn-winter months (September to

February), while high discharge is typical in spring and summer (March to August) (Chełmicki et al. 1998–1999). The incidence of floods in different regions depends on precipitation type. Two discharge maxima occur during the spring-summer season in the western Polish Carpathians (in the Beskid Śląski and Beskid Żywiecki Mountains): (1) a spring maximum (mainly April) resulting from snowmelt and (2) a summer maximum (mainly June) resulting from rainfall. Only the Dunajec and Poprad rivers have an extended flood season, due to a combination of belated and prolonged snowmelt in the Tatras and summer rainfalls. Rivers east of the Dunajec River experience major snowmelt-induced spring floods and secondary rainfall-induced summer floods (Dynowska 1972, 1995; Ziemońska 1973; Chełmicki et al. 1998–1999). The former do not develop particularly rapidly as snow cover in the Carpathians melts gradually. Flash floods induced by downpours tend to be localized and occur mainly in June and July. Summer floods due to prolonged (3–5 days of) rainfall build up slowly, but usually affect large areas. Daily precipitation can total 200 mm in the Soła, Skawa, Raba, and Dunajec catchments, 150 mm in the Wisłoka basin and 100 mm in the San basin (Osuch 1991; Dynowska 1995; Punzet 1998–1999). During flood events, precipitation is normally higher in the western Polish Carpathians than in the east (Cebulak 1998–1999). For this reason, rivers from the western Carpathians (Soła, Skawa, Raba, Dunajec) contribute more substantially to flood waves on the Vistula River than do tributaries from the eastern part (Wisłoka, San). Fortunately, catastrophic floods never affect all Carpathian tributaries of the Vistula River simultaneously (Punzet 1998–1999).

5.4 Mass Movements

Włodzimierz Margielewski
Zofia Rączkowska

The shaping of slopes by mass movements in high mountains differs fundamentally from that taking place in medium-height and low mountains or in foothills. In the Tatras, mass movements are controlled by slope type and climate belt. Rockwalls and rocky slopes develop mainly as a result of weathering processes and rockfalls of various size. As confirmed by measurements of *rockwall retreat* and debris accumulation (Kotarba et al. 1983; Kotarba 1984), the conditions are most favorable in the cold climate zone and on western-facing walls (Klimaszewski 1971; Kotarba 1976). Debris accumulation rates on *talus slopes* range from 0.1 to 10 cm year^{-1} (Table 5.1), much higher on carbonates than on granite if elevation is the same. Debris mostly accumulates at the top of talus slopes and convex midslope sections (Kotarba et al. 1983). Twelve models of talus slope formation (Fig. 5.5) have been defined according to the morphometry of the rockwalls and slopes to the process which displaces debris (Kotarba et al. 1987).

Weathered surface material on debris-mantled slopes and Richter-denudation slopes stabilized by vegetation is displaced by slow mass movements, including

Table 5.1 Dynamics of present-day geomorphological processes in the Tatras

Geomorphological process	Rate of process in climatic belts D	G	K	H	T	Study site	Author
Chemical denudation, $m^3\ km^{-2}\ year^{-1}$	85.6	95.1	37.7	35.8		Limestones, dolomites, Mała Łąka valley	Kotarba (1971)
Output of mineral material, $m^3\ km^{-2}\ year^{-1}$		12				Metamorphic rocks and granites, Starorobociański stream	Krzemień (1985)
Chemical denudation, $t\ km^{-2}\ year^{-1}$			10.3–10.9	10.2–10.4		Granites, upper Sucha Woda valley	Kot (1996)
Chemical denudation, $m^3\ km^{-2}\ year^{-1}$		14				Metamorphic rocks and granites, Starorobociański stream	Krzemień (1984a)
Retreat of rockwalls, $mm\ year^{-1}$			0.028–0.26			Granites, Pięć Stawów Polskich valley	Rączkowski (1981)
			0.0004–0.26 mean 0.04			Metamorphic rocks and granites, Starorobociański stream	Koszyk (1977)
			0.7			Metamorphic rocks and granites, Starorobociański stream	Kotarba et al. (1987)
				0.4		Granites, Chochołowska valley	Mamica (1984)
			0.1–3.0; mean 0.84			Limestones, dolomites, Mała Łąka valley	Kotarba (1972)
			0.0004	2.5			Kotarba et al. (1987)
		0.01–0.43		0.003–0.019		Limestones, dolomites, Bélanske Tatry	Midriak (1983)
	0.1–0.95					Western Tatras	Midriak (1983)
Frost creep, $cm\ year^{-1}$		0.4–0.5	0.6–0.7	1.7		Mała Łąka valley	Kotarba (1976)
Frost creep and solifluction, $cm\ year^{-1}$			0.1–0.3	0.3–2.0		Metamorphic rocks, Goryczkowa Świńska valley	Baranowski et al. (2004)

Plowing boulders, cm year^{-1}			0.14–3.25 mean 1.3			Limestones, dolomites, Mała Łąka valley	Kotarba (1976)
			2.7			Granites, Sucha Woda valley	Kotarba et al. (1983)
Needle ice, cm year^{-1}			38			Granites, Sucha Woda valley	Gerlach (1959)
Deflation, g m^{-2} year^{-1}			87.9; max. 163.7				Izmaiłow (1984a)
Eolian deposition, g m^{-2} year^{-1}			1–265				Izmaiłow (1984b)
Denudation by tree uprooting, m^{3} ha^{-1}	50,000					Mała Łąka valley	Kotarba (1970)
Max. accretion of talus slopes, cm year^{-1}		2–6	6–8	8–10	2–8	Mała Łąka valley	Kotarba (1976)
Accumulation on talus slopes around snow patches, cm year^{-1}				0.03–0.06		High Tatras	Rączkowska (1999)
Nival erosion, cm year^{-1}			0–5	0.05–4.2		Sucha Woda valley	Rączkowska (1993)
						Bélanske Tatry, Kopske Sedlo pass	Midriak (1996)
Water erosion, kg ha^{-1} year^{-1}	9–33	12–253					Kotarba (1976)
Length of bedload transport, m per flood	100–120		> 20			Chochołowski stream	Krzemień (1991)
	160 to few hundreds					Biały stream, tributary streams of the Chochołowski stream	Kaszowski (1973), Krzemień (1991)

D temperate cool, (forest) 900–1,100 m above sea level, *G* cool (forest) 1,100–1,450 m, *K* very cool (subalpine) 1,450–1,850 m, *H* temperate cold (alpine) 1,850–2,200 m, *T* cold (seminival) >2,200 m

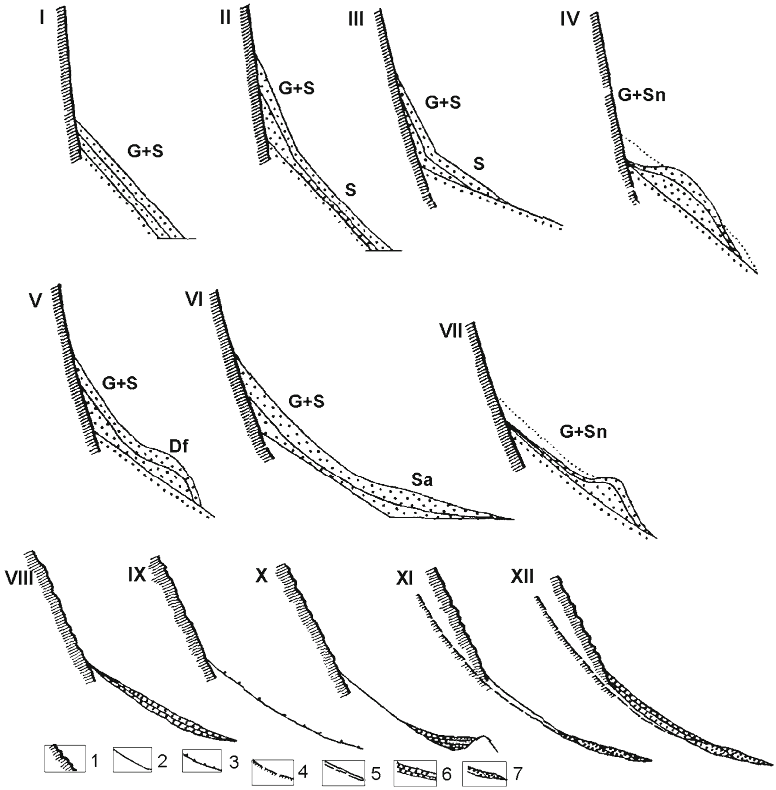


Fig. 5.5 Models of accumulation of material on debris slopes. The High Tatras – I–VII; the Western Tatras – VIII–XII; *G* supply of material by gravity, *S* sliding on debris surface, *Sn* sliding on snow surface, *Df* by debris flows, *Sa* slush avalanche. 1, rocky slope; 2, debris slope surface stabilized by vegetation; 3, sliding of single particles downslope; 4, rocky chute; 5, debris flow gully; 6, deposition of material by gravity; 7, deposition of metrial by debris flows (Kotarba et al. 1987)

solifluction mainly in the alpine and subalpine belts and by *creep* occurring overall. The highest rates of surface material displacement of 1.7–2.0 cm year^{-1} have been detected in the moderately cold zone (Table 5.1), decreasing in the forest belt (Kotarba 1976). Such processes smooth slopes with dome-shaped or step-type microtopography.

The most important geomorphic agent in the Tatra Mountains is abrupt mass movements, which displace and transport significant quantities of material (Midriak 1984; Krzemień 1988; Kotarba 1992, 1995, 2002; Rączkowska 2006, 2008). Such rapid and high-energy movements include debris flows, avalanches, and rockfalls (Kotarba 1992, 1995; Rączkowska 2006). *Debris flows* and *avalanches* take place across multiple altitudinal belts (Kotarba 2002) and often transport weathered material from crests to valley floors, thus connecting the slope system with the channel system (Kotarba et al. 1987). They create new landforms such as debris flow gullies, several meters deep and several hundred meters long, and debris flow levees, over 1 m in height or significantly change the morphometry of existing forms. The changes in relief are linear (debris flows, avalanches) or of point-like manner (rockfalls).

Among abrupt mass movements, debris flows are the most common and most influential in relief formation (Kotarba 1992, 1995, 1997). A single flow event can transport up to tens of thousands of cubic meters of material, often to the bottoms of cirques and glacial valleys, permanently transforming their topography and extending debris slopes (Midriak 1984; Kotarba et al. 1987; Krzemień 1988, 1991; Kotarba 1992; Rączkowska 1999). Debris flows are caused by short but powerful downpours. Precipitation events around 25 mm most often trigger debris flows (Krzemień 1988). They only affect the top of talus slopes. Precipitation intensities 35–40 mm h^{-1} generate debris flows that run across the full length of talus slopes (Kotarba 1992, 2002). The momentary intensity during such rainfalls is ≥1 mm min^{-1}.

The greatest changes in relief are caused by downpours during continuous and widespread rainfalls. In such cases, in addition to debris flows above the upper timberline, *mudflows* and *shallow landslides* occur in forests, affecting the entire length of the slope and removing all weathered material, leaving behind nothing but bare rock (Kotarba 1999, 2002). Lichenometric research has shown that debris flows and rockfalls used to be more active during the Little Ice Age (Kotarba et al. 1987; Krzemień 1988; Kotarba 1992, 1993–1994, 1995, 2004; Kotarba and Pech 2002). In the past 20 years precipitation intensity has increased and extreme geomorphic processes have become more widespread in the Tatra Mountains (Kotarba 1997, 2002, 2004; Kotarba and Pech 2002) than between 1958 and 1978 (Niedźwiedź 2003).

Surface mass movements commonly occur in the Outer Carpathians in the medium-height and low mountains and foothills (Fig. 5.6.: 1). In some mountain areas, they are of primary importance (Starkel 1960; Kotarba 1986; Ziętara 1988; Bober 1984). The major landslides date back to the Late Glacial and Holocene (Alexandrowicz 1997; Starkel 1997; Margielewski 2002, 2006a). Numerous contemporary mass movements are local phenomena and their intensity is associated with extreme precipitation events (Figs. 5.7: 1-4 and 5.8) (Ziętara 1968; Mrozek et al. 2000; Rączkowski and Mrozek 2002; Poprawa and Rączkowski 2003; Gorczyca 2004). Gravitational movements usually displace slope surface material in small and shallow *rotational landslides* (Fig. 5.7: 1–3) (Jakubowski 1965; Ziętara 1968; Poprawa and Rączkowski 2003; Mrozek et al. 2000; Gorczyca 2004). Mass movements that involve bedrock are usually shallow *translational* displacements over bedding planes, fractures and faults, called *structural landslides* (Fig. 5.7: 4 and 5.8) (Ziętara 1968; Bober 1984; Margielewski 2006b). Far less numerous are *deep-seated rockslides*, which tend to be composed of several types of gravitational movements (Fig. 5.7: 5) (Ziętara 1968). A common occurrence is the reactivation of old landslides by new generations of mass movements (Ziętara 1964; Jakubowski 1967; Bajgier-Kowalska and Ziętara 2002). In fact, some landslides exhibit signs of continuous development (Gil and Kotarba 1977). Less common types of landslides include mudflows and debris-and-mud flows (German 1998; Poprawa and Rączkowski 1998; Ziętara 1999; Gorczyca 2004). *Creep* is observed primarily in weathered (Starkel 1960; Jakubowski 1965; Gerlach 1966) and block-type material (Pękala 1969).

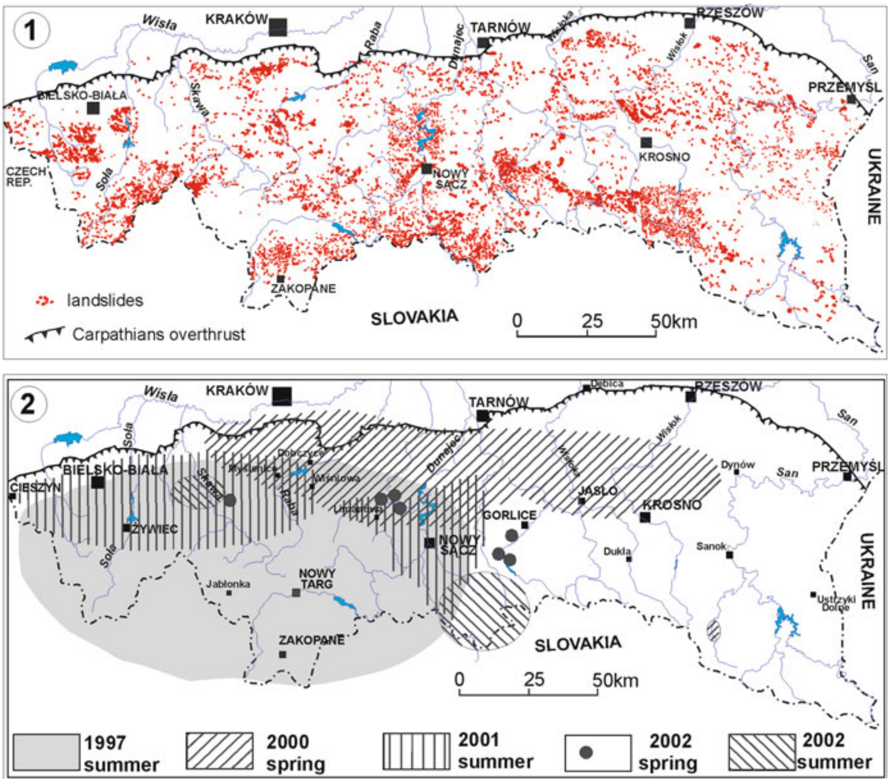


Fig. 5.6 1, Distribution of landslides registered by Polish Geological Institute in the Polish Carpathians (state in 2000) (After Poprawa and Rączkowski 2003, supplemented); 2, Occurrence and range of recent landslides in the Carpathian area, generated by various hydrometeorological factors (snow melting, downpours, heavy rainstorms) (After Poprawa and Rączkowski 2003, supplemented)

Local downpours, heavy continuous rainfalls, and abrupt snowmelt can exceed threshold values of gravitational processes and *trigger* mass movements. However, it is difficult to identify universal *precipitation thresholds,* which heavily depend on slope susceptibility to infiltration (Ziętara 1968; Gil and Starkel 1979; Starkel 1996, 2006; Gil 1997). It has been assumed that continuous and widespread rainfalls in the 150–400 mm range in 2–5 days generate earthflows and weathered material flows. It has also been assumed that rainy seasons with monthly totals of 100–500 mm lead to deep-seated landslides (Starkel 1996). Repeated hydrometeorological events are the most effective in geomorphological terms. Other very effective events include sudden downpours during continuous and widespread rainfalls (as observed in July, 1997 in the Beskid Wyspowy Mountains – Gorczyca 2004; Starkel 2006). The intensity of geomorphic transformation by mass movements depends on the incidence of intense precipitation events, which never affect the entire area of the Polish Flysch Carpathians. Such events can occur repeatedly

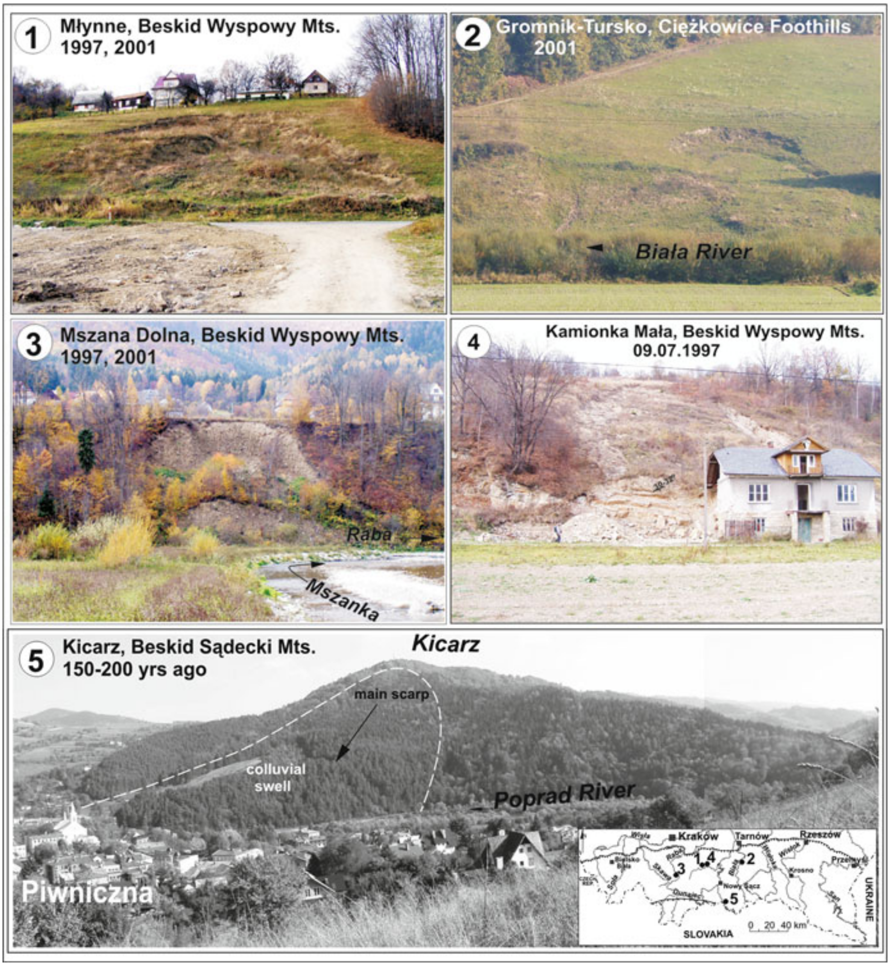


Fig. 5.7 Landslides formed and rejuvenated during current heavy rains (date shows the age of main stage and rejuvenation of each form): 1–3, shallow landslides formed in weathering (slope) mantles (forms 1–2) and fluvial deposits (form 3); 4, Kamionka Mała (Beskid Wyspowy): shallow weathering-rocky landslide (structural, slip consequence); 5, deep seated rocky landslide formed ca 150–200 years ago (see Pulinowa 1972) on Kicarz Mountain, Beskid Sądecki Mountains (Photo: W. Margielewski)

over the course of several consecutive years, as in 1958–1960 and 1997–2002 (Fig. 5.6: 2) (Mikulski 1954; Ziętara 1968; Cebulak 1992, 1998–1999; Poprawa and Rączkowski 2003; Rączkowski 2007) and cause pulses (uneven activity) in geomorphic processes (Ziętara 1968). The landforms produced by new and reactivated mass movements over the past 200 years are dated by lichenometry (Bajgier-Kowalska 2008) and dendrochronology (Krąpiec and Margielewski 2000). Both methods confirm the close correlation between mass movement activity and extreme precipitation events causing floods across the Carpathian region, while

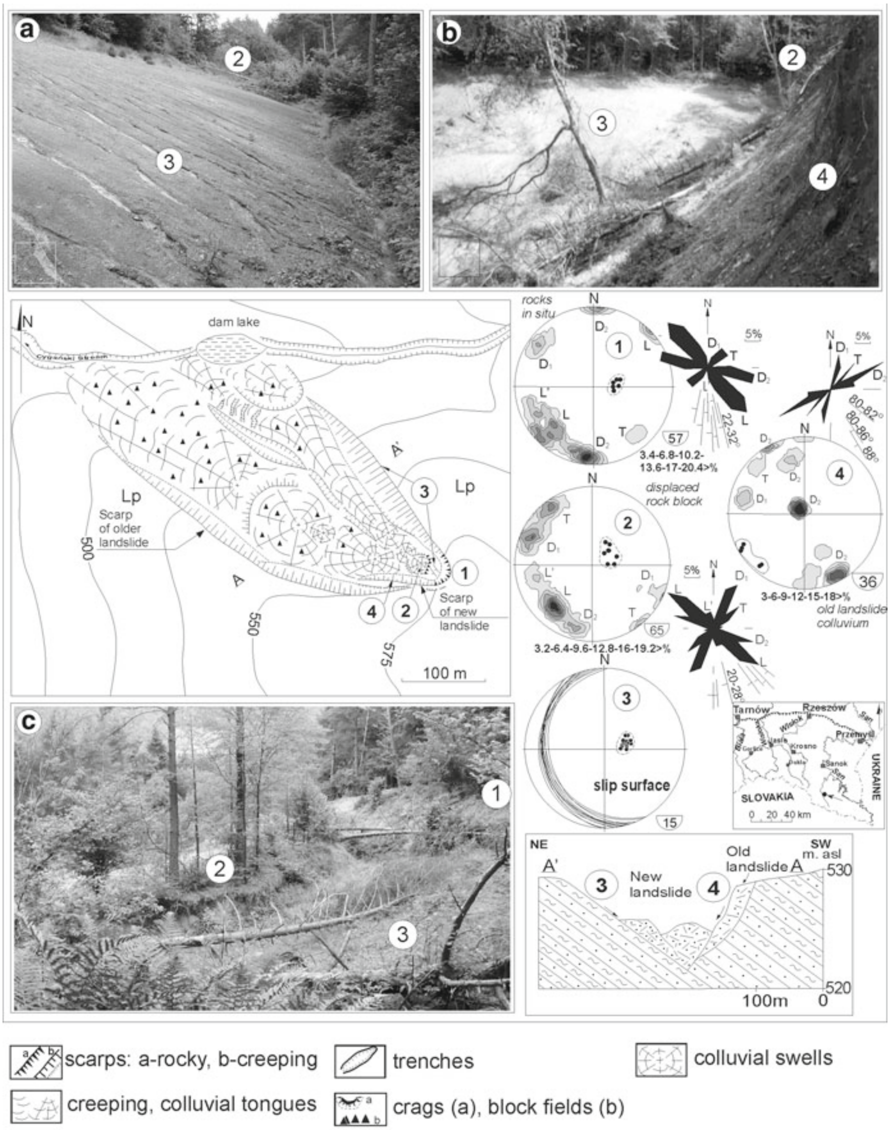


Fig. 5.8 Example of typical structural landslide, slip consequent, displaced along variegated shale (*Lp*) surface. Landslide in the Prełuki (Cygański stream), Bieszczady Mountains, formed in spring 2000 (After Margielewski 2004). Landslide successively damaged older form (signed on map and cross section). Map, cross section, and photographs of various parts of the landslide: *A* and *B* slip surface, *C* displaced rock mass. Joints on rose diagram, and contour diagram (equal area plot, pole projection on lower hemisphere, number of measurements and contour interval near each diagram). Bedding position (and slip surface) on pole point diagram (projection on lower hemisphere). Joints system; *L* longitudinal, *D1*, *D2* diagonal, *T* transversal (After Mastella et al. 1997) (Photo: W. Margielewski)

seismic tremors (Pagaczewski 1972; Schenk et al. 2001) triggering mass movements in the Polish Carpathians have not yet been properly documented (Gerlach et al. 1958; Poprawa and Rączkowski 2003). In addition to precipitation, other major factors of mass movements are slope susceptibility to landslides (geological structure), the thickness of slope deposits, vegetation cover, and human impact. Mass movements of different rate affect the sides of river valleys, valley heads, and slopes in the Beskidy Mountains and the Carpathian Foothills.

Flood-driven *lateral erosion* of valley sides is the most important trigger of mass movements, leading to abrupt displacements of slope cover and bedrock (Ziętara 1968; Kotarba 1986; Mrozek et al. 2000; Gorczyca 2004; Rączkowski 2007). Mass movements often reactivate old landslides (Fig. 5.7: 3). The resulting extensive landforms are composed of surface material and bedrock of a different age that remain subject to the influx of new material during subsequent flood events (Ziętara 1968). On a local scale, such processes lead to a gradual upslope extension and steepening of valley sides (Fig. 5.7: 3) (Ziętara 1968), accompanied by meander shifts (Starkel 1972). The colluvial material reaching the river channels is rapidly removed by floods. However, in the case of smaller valleys, an abrupt delivery of colluvia can agitate water to the point where it causes intensified fluvial erosion, local channel scour, widening of the valley floor, and steepening of valley sides (German 1998).

Mass movements can also be triggered by *headward erosion* (Starkel 1960; Ziętara 1968; Kotarba 1986). Intense precipitation leads to incision into slopes. Existing landslide surfaces become significantly larger as new material is brought down by repeated mass movements, resulting in concentric landforms (Ziętara 1968). In mountains of medium height, mass movements extend to headwater areas upslope, form stepped slope profiles, and, over thousands of years, lengthen valleys and create small tributary valleys (Ziętara 1968). In foothill areas, where valley heads are smaller, landslides help propagate them upslope. In effect, landslides shape the upper parts of slopes, and in the long term, cut through crests and form narrower ridges (Kotarba 1986).

Landslides occur as a result of the extra loading of masses of rock and regolith by rainwater. In the Carpathian Foothills they tend to be shallow and small in area (Fig. 5.7: 1–2) (Kotarba 1986), while in the Beskidy Mountains, extend high upslope and deep into the bedrock. Such landslides are called rockslides, sometimes also affecting weathered material. They can be quite extensive but of local occurrence (Fig. 5.7: 5) (Ziętara 1968; Pulinowa 1972; Oszczypko et al. 2002; Rączkowski and Mrozek 2002). In some cases, older landslides are reactivated by newer generations of mass movements and, in other cases, debris and mud flows occur in colluvia (Ziętara 1999). In the Flysch Carpathian region, under the current climatic conditions, shallow surface material and rock-weathered material landslides predominate. They take place primarily in the Carpathian Foothills and in medium-height mountains and are quite typical of deforested areas (Fig. 5.7: 1–2). Most contemporary mass movements develop in areas of previous landslide activity. They usually remove slope surface material, although there have been locally recorded cases of so strong a shearing action that bedrock became exposed (Poprawa and Rączkowski 1998). Such landslides normally do not produce remarkable topographic changes, but on a local scale, given their large number and density, even shallow movements have substantial effect: producing steeper (Fig. 5.7: 3) and also smoother valley sides (Fig. 5.7: 1). They eliminate or smoothe morphological irregularities like edges of old landslide scarps and road cuts and agricultural terraces (Ziętara 1968; Kotarba 1986; Mrozek et al. 2000; Gorczyca 2004). Repeated landslides can locally broaden and reshape valleys. An undulating or step-like slope profile with numerous depressions results. In effect, slopes affected by mass processes are characterized by an irregular shape both along their longitudinal and transversal sections (Starkel 1960; Ziętara 1968; Kotarba 1986; Gorczyca 2004; Rączkowski 2007). An important feature of landslides in regolith is their tendency to planate relief – even if the landslides in question tend to recur. Shallow landslides are moderately and occasionally influential in shaping the slopes of the Beskidy Mountains and Carpathian Foothills (Ziętara 1968; Mrozek et al. 2000; Gorczyca 2004).

Mass movements *in bedrock*, although only occuring locally, tend to be deeper and exert a more permanent impact on relief (Figs. 5.7: 5 and 5.8). Repeated landslide activity on slopes leads to a gradual fragmentation of straight slopes and the formation of side valleys. Side valleys in valley heads become elongated as a result of regression promoted by subsequent mass movements (Fig. 5.8) (Kotarba 1986). On the other hand, mass movements generate concave slope profiles (Starkel 1960). Both in the Beskidy Mountains and in the Carpathian Foothills, however, it takes thousands of years to produce perceptible topographic changes.

Rockslides can be traced for a very long time, particularly in resistant sandstones, where contemporary rockfalls of various dimensions accompany the retreat of landslide-created rock scarps (Jakubowski 1967; Ziętara 1968; Bajgier-Kowalska 2008). Large-scale landslide features produce distinct relief in resistant rocks. However, these types of deep-seated landslides have only been sporadically recorded in the past 200 years (e.g. Fig. 5.7: 5) (Zuber and Blauth 1907; Sawicki 1917; Pulinowa 1972).

5.5 Water Erosion

Jolanta Święchowicz

5.5.1 Sheet and Rill Erosion

For hydrological and geomorphological reasons, sheet and rill erosion in the flysch Carpathians is of a different nature than in the Beskidy Mountains and the Carpathian Foothills (Fig. 5.9) (Starkel 1972, 1991; Słupik 1978; Gil 1979). An *infiltration-evapotranspiration* water circulation pattern with a predominance of throughflow has been identified in middle mountains. In the case of straight and convex slopes covered by clays with abundant rock debris, rainwater tends to infiltrate, leach the regolith, and create piping tunnels. On complex slopes, steep on sandstones at the top and schist at the bottom, percolating waters into the fractured sandstone induce

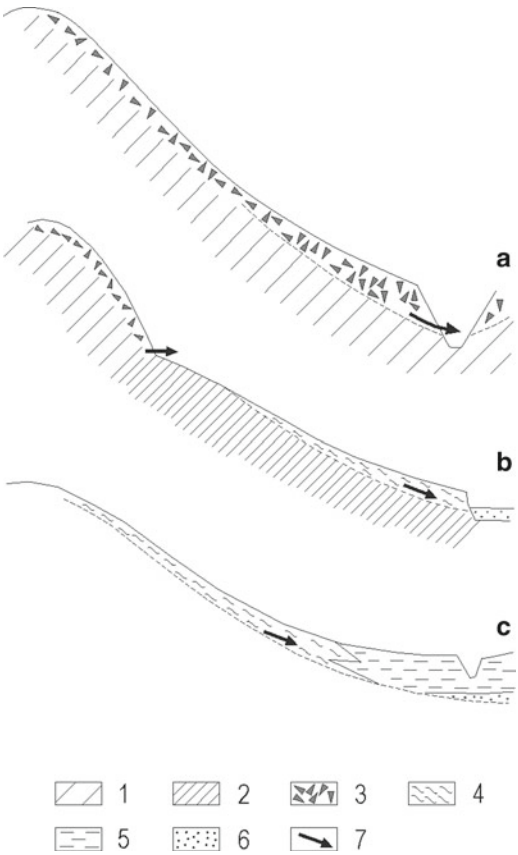


Fig. 5.9 Slope types and slope cover types in the Polish Carpathians (After Starkel 1991). (**a**) undercut mountain slope with accelerated throughflow, (**b**) bi-segmented mountain slope with limited throughflow, (**c**) foothill slope (upland); 1, sandstone series; 2, shale series; 3, debris slope cover (permeable); 4, solifluction clayey-skeletal cover; 5, alluvial clays/soils; 6, sand and pebbles; 7, principal groundwater flow directions

landslides and incision at footslopes. In mountain catchments, especially in upstream sections, slopes come into direct contact with stream channels, which directly receive water and material (Starkel 1972; Froehlich and Słupik 1977; Froehlich and Starkel 1995). In Carpathian Foothill catchments, infiltration-evapotranspiration is the dominant type of water circulation pattern with substantial input from surface runoff, which favors sheet erosion (Fig. 5.9) (Słupik 1973, 1978; Starkel 1991; Gil 1999). In the Carpathian Foothills, slopes are separated from stream channels by wide valley floors, which render direct delivery of weathered material impossible, and, thus most of it is deposited at footslopes (Święchowicz 2002c).

Vegetation cover is another key factor in the effectiveness of sheet and rill erosion. The forested slopes of the Beskidy Mountains and the Carpathian Foothills are not subject to significant sheet erosion (Table 5.2) (Gerlach 1976b; Gil 1976; Święchowicz 2002c). Forests, however, are dissected by a network of roads, footpaths, and logging tracks. While road density on forested slopes in the Beskidy Mountains is far lower than that in agricultural areas, roads play a key role in water and sediment transport. Roads facilitate flow and linear erosion and serve as delivery routes

Table 5.2 Rate of sheet and rill erosion (kg ha^{-1} year^{-1}) in different parts of the Polish Carpathians

Area	Author	Research period	Sheet and rill erosion [kg ha^{-1} year^{-1}]				
			Forest	Meadow	Fallow land	Potatoes	Cereal
Jaworki (Beskid Sądecki Mountains/ Pieniny Mountains)	Gerlach (1966, 1976a, b)	1956–1958 1968–1971		3.1–7.1 2.7–11.8		63,195.0	17,022.0
Jaszcze (Gorce Mountains)	Gerlach (1976b)	1969–1971	2.1–12.9				
Szymbark (Beskid Niski Mountains/ Ciężkowice Foothills)	Gil (1994, 1998, 1999, 2009)	1969–2000		3.5–466.1		371.5–99,139.9	10.5–8,495.1
Polanka Hallera (Wieliczka Foothills)	Drużkowski (1998)	_	25.0	50.0		75,000.0	1,250.0
Łazy (Wiśnicz Foothills)	Święchowicz (2010)	2007–2008	–	2.3–41.9	2.9–47,340.2[2]	0–43,396.0	8.0–31.4

Based on Gerlach (1966, 1976b), Gil (1994, 1999, 2009), Drużkowski (1998), and Święchowicz (2010) for suspended matter to stream channels (Froehlich and Słupik 1980a, 1986; Froehlich 1982; Soja and Prokop 1996).

Land use changes from season to season on arable lands by crop rotation and tillage operations in the Beskidy Mountains and the Carpathian Foothills. Along with changing weather conditions, this affects the rate of sheet and rill erosion (Table 5.2) (Gerlach 1966, 1976a, b; Gil 1976, 1979, 1986, 1998, 1999, 2009; Święchowicz 2002b, c, 2010). Areas of sparse or no vegetation are characterized by soil erosion several hundred times more intensive than areas with a dense vegetation cover (Table 5.2). In winter 30–40% of surfaces can be either free of vegetation or possess a sparse winter plant cover. Such conditions favor concentrated runoff and thaw facilitates rill erosion. Sheet erosion patterns are similar across fields planted with winter crops, and arable land in general. Intense water runoff takes place primarily during advective-type thaws often associated with rainfall (Słupik 1978; Gil 1999).

In summer runoff and soil erosion rates and variability are a function of land use, precipitation amounts, and rainfall intensity (Słupik 1973, 1978; Gil and Starkel 1979; Gil 1994, 1999, 2009; Starkel 1986, 1996; Święchowicz 2002a, b, 2008), as well as rainfall erosivity (EI_{30}) and 30-min maximum intensity (I_{30}) (Demczuk 2009; Święchowicz 2010). Intense sheet and rill erosion takes place on fields where plant cover is not dense enough to protect the ground surface from splash and sheet flow. Slopes and stream channels are primarily transformed by short local downpours and continuous rainfalls. Short but intense summer downpours produce precipitation amounts up to 100 mm at a rate of 1–3 mm min^{-1}. The infiltration capacity of the parent material is exceeded and heavy splashing ensues. The short-duration surface runoff transports soil over short distances to accumulate on densely vegetated slopes or on valley floors in the form of deluvial fans (Gil and Słupik 1972; Gerlach 1976b; Święchowicz 2002a, 2008). Stream channels only receive material transported by field roads from adjacent fields (Froehlich and Słupik 1980a, 1986; Gil 1994, 1999). Heavy sheet erosion mainly affects arable lands and may be catastrophic for the plowed layer (Gil 1998, 2009; Święchowicz 2008, 2009). Sheet erosion is light on fields with crops of dense surface cover and on meadows. Continuous rainfall amounting to 100 mm or more with a rate of 10 mm·h^{-1} saturates the parent material and surface runoff starts once the infiltration capacity of the soil is exceeded or precipitation intensity increases abruptly for a moment. The amount of surface runoff is then determined by the properties of the parent material and the water capacity of the slope material. At that point, land use and plant cover are less important (Słupik 1973, 1978). If the parent material becomes fully saturated with water, surface runoff restarts across affected surfaces. Surface runoff lasts significantly longer than precipitation. Sheet and rill erosion measured during continuous rainfall events is usually much more limited than during downpours (Gil 1986, 1999).

Measurements obtained on experimental fields suggest significant sheet and rill erosion on agricultural slopes. The mechanism of soil erosion is, however, complex and its high rates are not reflected directly in the *transport* and delivery *of material* from slopes to river channels. Material is transported in discrete steps over short distances and its amount varies along the length of the slope. Accumulation affects fields with dense vegetation, gentler slope sections, and footslopes (Gerlach 1966;

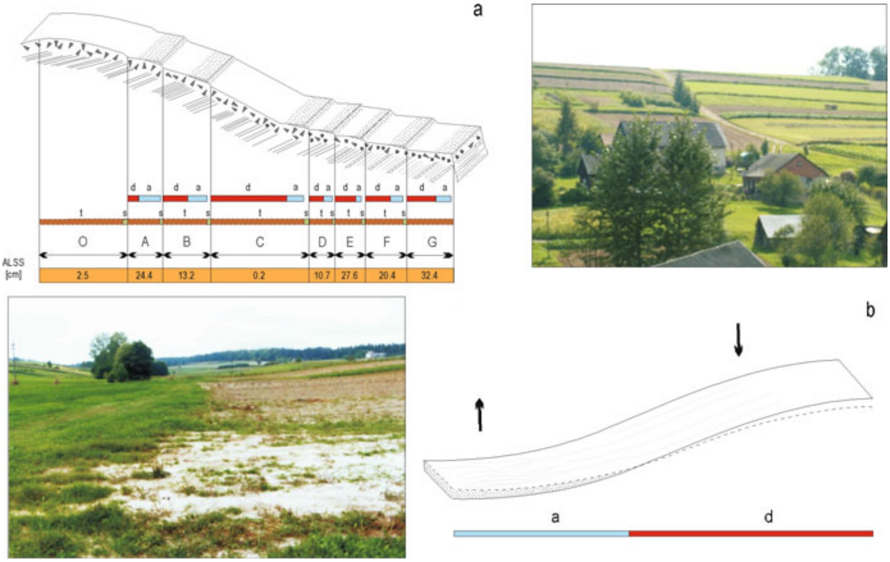


Fig. 5.10 (**a**) Terraced field pattern slope profile (Modified Gerlach 1966). *O* nonterraced slope, *A–G* agricultural slope, *t* terrace surface, *s* steep scarp (normally covered by grass or bush), *d* degradation surface, *a* aggradation surface, *ALSS* average loss of slope surface (cm), (**b**) Convex-concave slope profile: *d* degradation surface, *a* aggradation surface (Photo: J. Święchowicz)

Gil 1976; Święchowicz 2001, 2002c, 2008). Valley floors and river channels only receive material transported through (ephemeral) gullies during sudden downpours and continuous rainfall (Froehlich, Słupik 1986; Święchowicz 2008). The shape of slopes and their location with respect to local denudation zones influences material delivery to stream channels. Abundant material is transported down straight slopes and convex slopes located next to stream channels. Convex-concave slopes tend to deliver limited amounts of material. The longer the concave segment, the less material is delivered. When slopes deliver more material to valley floors, they themselves flatten out in downslope direction (downslope evolution). The opposite happens when more material is transported further in a stream channel than received (upslope evolution or stream-induced slope undercutting – Gerlach 1976b).

The transport and delivery of material from slopes to stream channels depend on the local *field pattern*. A mosaical field pattern characteristic of both the Beskidy Mountains and the Carpathian Foothills leads to different sheet and rill erosion rates on individual fields (Gil and Słupik 1972; Gerlach 1976b; Gil 1979; Święchowicz 2002a, c). In the Beskidy Mountains with high surface permeability and substantial variability in slope lengths and gradients, a field pattern parallel to contour lines prevails, and is stabilized by agricultural terraces (Gerlach 1976b; Gil 1979). In the marginal zone of the Carpathian Foothills, where permeability is lower and there is little diversity in terms of slope length and gradient, a longitudinal field pattern prevails with no elevated boundaries between fields (Święchowicz 2001). A permanent field boundary pattern and contour plowing significantly reduce soil erosion and lead to the formation of agricultural terraces. Water runoff and soil erosion take place on isolated fields that act as independent systems (Fig. 5.10) (Gerlach 1966; Słupik 1973; Gil 1976, 1979). Such fields are subject to short-distance transport and accumulation. Material is removed by a system of rills and field roads, which run downslope along field boundaries. It has been estimated that 40% less soil removal occurs on *terraced* agricultural *slopes* than on terrace-free slopes (Gerlach 1966).

Particle transfer systems are either linear-continuous or cascade-segmented. The former affects the entire length of the slope during snowmelt, most precipitation events, and is limited to roads, paths, and incisions such as rills, ephemeral gullies, and gullies. The latter affects terraced fields (Froehlich and Walling 1995). *Downslope plowing* intensifies erosion processes by orders of magnitude. If a slope is being used in a homogeneous manner from top to bottom, then sheet erosion intensity grows with the length of the slope and deluvial matter is accumulated at the foot of the slope (Fig. 5.10b) (Gil 1979; Święchowicz 2001, 2002c). Sheet and concentrated erosion are especially intense in large areas planted with the same type of crop (Święchowicz 2008). The absence of traditional erosion control measures such as elevated field boundaries and scarps intensify soil erosion. Deep rills or ephemeral gullies on slope surfaces, however, are only temporary phenomena and are usually destroyed by tillage operations (Święchowicz 2002c, 2004, 2008, 2009).

In Carpathian Foothill catchments, the delivery of material from slopes to stream channels happens locally and exclusively during extreme precipitation events. However, even then, most of the material is deposited on slope surfaces, deluvial flats at the foot of the slope, and on valley floors covered by permanent vegetation. The resulting relief is typical of the Carpathian Foothills: convex-concave, multi-segmented and stepped agricultural, and flat-bottom valleys with deluvial areas that constantly receive new slope material. The flats and flat valley floors constitute a zone that separates slopes from stream channels. It is also a zone where most of the material from slopes is deposited (Święchowicz 2002a, c).

In the Polish flysch Carpathians, the rate of slope transformation driven by sheet and rill erosion depends on vegetation cover and land use. Contour tillage, crop rotation, and all other tillage operations reduce soil erosion on arable land and lead to the formation of agricultural terraces and an irregular slope profile. On the other hand, on slopes with downslope plowing, the deposition of material at the foot of the slope or on the valley floor leads to the elongation of the concave part of the slope. This results in the elimination of clear morphological boundaries between slopes and valley floors (Fig. 5.10) (Święchowicz 2006).

5.5.2 Piping

In the Polish Carpathians the presence of debris, clayey-debris, and silt surfaces over impermeable bedrock favors piping (suffosion) (Czeppe 1960; Starkel 1960, 1965, 1972; Galarowski 1976). Piping occurs along fractures and channels produced mainly by small rodents in near-surface layers (40–60 cm). In deeper layers, living and dead tree roots serve as piping channels. Pipes also develop where slope clay comes into contact with the loam-debris layer or bedrock. Rainwater and meltwater flows under the compact turf layer along pipes and washes out fine particles. In most areas, piping is an episodic process during downpours and the melting of snow and ice (Starkel 1960). Piping features are created more frequently by meltwater than by rainwater (Czeppe 1960; Starkel 1960) and usually occur on permanent meadows and pasture lands. In tilled areas, plowing and related operations destroy existing channels and render impossible the development of typical piping landforms. Piping features are more common in the Eastern Polish Carpathians (Czeppe 1960; Starkel 1965; Galarowski 1976) than in the west (Starkel 1960). *Tunnels* are the only underground forms, as the development of shafts is limited by the shallow regolith. On the surface *piping pits*, foredeeps like cauldrons and funnels occur. Piping pits arise as a result of compaction and tend to be rather shallow (less than 1 m deep), covered with turf, and can be bowl-shaped, flat-bottom deeps, and ravines. The walls of foredeeps are usually steep while their bottoms tend to be flat and covered with turf. *Cauldrons* are of oval or rounded shape with steep sides, while their floors are parts of bottoms of piping channels. They are created when a ceiling caves in over a major tunnel. Cauldrons also develop due to landslides and the washing out of walls, eventually leading to the formation of a *piping fan* and dead-end valleys and gullies. Outlets of piping channels at the ends of valleys contribute to the upslope propagation of headwaters (Czeppe 1960; Starkel 1960; Galarowski 1976).

5.6 Aeolian Processes

Jolanta Święchowicz
Zofia Rączkowska

The rate of occurrence and the intensity of aeolian processes vary from year to year and with the individual landforms and also depend on elevation (Gerlach and Koszarski 1968; Janiga 1975; Gerlach 1976b, 1977; Welc 1977; Izmaiłow 1984a, b, 1995b; Drużkowski 1998).

Aeolian processes are induced by even moderately strong *winds* (>5 $m\,s^{-1}$), although they are the most effective with strong (>10 $m\,s^{-1}$) and very strong winds (>15 $m\,s^{-1}$). Under natural conditions, wind action is limited to vegetation-free surfaces (river banks, sandbars, levees), areas lying above the upper timberline, and in the case of larger forested areas, to the overturning of trees (Gerlach 1960, 1966, 1976a, b; Kotarba 1970). In the Tatra Mountains, as a result of a single wind event, 50,000 $m^3\,ha^{-1}$ of weathered and block-type material can be transferred on the roots of felled and uprooted trees. This produces a unique and relatively permanent slope micromorphology of mounds and depressions (Kotarba 1970).

Table 5.3 Rate of wind-driven soil particle deposition (t ha^{-1} year^{-1}) in different parts of the Polish Carpathians

Area	Author	Research period	Deposition (t ha^{-1} year^{-1})
Jasło-Sanok Depression	Gerlach T, Koszarski L (1969)	Winter 1964/1965	125.0–875.0
Bieszczady Mountains	Pękala K (1969)	Winters 1964/1965–1966/1967	1.1–9.0
Beskid Niski Mountains	Janiga S (1971)	Winters 1965/1966–1969/1970	70.0–241.0
Ciężkowice Foothills	Welc A (1977)	Winter 1968/1969	1.9–10.7
Tatra Mountains	Izmaiłow B (1984b)	1975–1980	0.01–2.65
Wieliczka Foothills	Izmaiłow B (1995a, b, c)	March 1994–February 1995	0.06–1.76

Based on Gerlach and Koszarski (1969), Pękala (1969), Janiga (1971), Welc (1977), Gerlach (1986), and Izmaiłow (1984b, 1995a, b, c)

Aeolian processes have been detected above the upper timberline in the Bieszczady Mountains (Pękala 1969) and the Tatra Mountains (Izmaiłow 1984a, b). The rate of aeolian *accumulation* on the leeward slopes of the Bieszczady Mountains ranges from 1.1 to 9.0 t ha^{-1} year^{-1}, while in the Tatra Mountains from 0.01 to 2.65 t ha^{-1} year^{-1} (Table 5.3). *Deflation niches* (Kotarba 1983) and *terracettes* develop on broad ridges in the Tatras, especially near mountain passes where wind speed can exceed 6.4 m s^{-1}, blowing away up to 1.63 t ha^{-1} year^{-1} of fine weathered particles (Izmaiłow 1984b). Aeolian processes are more decisive for relief evolution in environments affected by human impact. Arable land offers the most favorable conditions for aeolian processes. Deflation takes places mainly during winter (November–March) and is characterized by short episodes that do not recur year after year (Gerlach and Koszarski 1969; Janiga 1975; Welc 1977; Izmaiłow 1995a, b). The primary targets of the destructive power of the wind are flat surfaces and the upper parts of windward slopes, while leeward slopes are sites of accumulation. This reduces the heights of hills and slope gradients and causes footslope extension and accumulation in depressions (Gerlach 1986).

The most intensive deflation on windward slopes and the largest accumulation of aeolian sediment on leeward slopes, on the order of several hundred t ha^{-1} year^{-1}, has been detected in the Jasło-Sanok Depression (Gerlach and Koszarski 1968; 1969). Values recorded in the boundary zone between the Beskid Niski Mountains and the Carpathian Foothills do not exceed 10.7 t ha^{-1} year^{-1}, while at the edge of the Carpathian Foothills, despite the presence of silt formations, values on the order of several t ha^{-1} year^{-1} are found (Izmaiłow 1995a, c).

The intensity of deflation increases with increasing elevation above sea level (Janiga 1971, 1975). The average quantity of material deposited in the shallow valleys of the Jasło-Sanok Depression (320–350 m) separated by small hills reaches 70 t ha^{-1} year^{-1}. In the lower parts of the Beskid Niski Mountains (350–500 m), characterized by foothill relief, average deposition reached slightly over

128 t ha^{-1} year^{-1}, while in forests of higher elevation (550–600 m), it is slightly more than 241 t ha^{-1} year^{-1} (Table 5.3) (Janiga 1975).

In some parts of the Polish Carpathians, aeolian processes play a more significant part in geomorphic evolution than sheet erosion (Gerlach and Koszarski 1968). On the windward slopes of the Jasło-Sanok Depression, deflation was responsible for 60% of the soil removed over the course of the past 600 years, while sheet erosion was responsible for 40%. On the other hand, soil deposition on leeward slopes was far more common than sheet erosion (Gerlach and Koszarski 1968, 1969; Gerlach 1976b). The heights of rounded hilltops and the middle and upper parts of windward slopes have become reduced while their lower parts built up and extended. The reduction in height during the era of human activity has been about 60 cm while the amount of accumulation has reached 105 cm and extension about 50 m. Leeward slopes, except for their upper parts, have become longer and higher during the same period of time. Leeward slopes have become about 33 m longer and 15–115 cm higher. Where aeolian accumulation has been the greatest, an average of 3.5 m of new deposits have been recorded only in the last 100 years (Gerlach and Koszarski 1968, 1969; Gerlach 1976b, 1986). With the recent tendency to reduce the amount of cultivated land in the Polish Carpathians, the land area susceptible to deflation is steadily shrinking.

5.7 Periglacial Processes and Landforms

Zofia Rączkowska

In modern times, periglacial processes are significant only in high mountains such as the Tatras and Babia Góra Massif above the upper timberline. In the Tatra Mountains, they are predominant (Jahn 1958, 1975; Rączkowska 2007), although most of the active periglacial landforms can be found here at elevations ranging from 1,800 to 2,050 m (Kotarba 1976; Karrasch 1977; Rączkowska 2007). Only in the Tatras exist isolated patches of *permafrost* – above 2,300 m on southern-facing slopes and above 1,930 m on northern-facing slopes (Dobiński 1997, 2004; Mościcki and Kędzia 2001; Kędzia et al. 1998; Gądek et al. 2009; Gądek and Kędzia 2008).

Frost weathering is the principal type of weathering here. However, judging from the low rates of rockwall retreat, its intensity is rather low (Table 5.1). The intensity of weathering and rockfall was most likely the highest during the nineteenth century (Kotarba and Pech 2002), as a result of climatic conditions and earthquakes (Kotarba 2004). The lowering of ridges and slopes and the retreat of rockwalls in the Tatra Mountains as well as the cryoplanation terraces in the Babia Góra Massif are all effects of weathering.

Frost processes, such as *needle ice activity* and *solifluction*, participate in the transfer of weathered material on rock slopes with regolith mantle or on slopes of weathered material only. The rates of cryogenic processes are greater in the alpine belt than in the subalpine belt (Table 5.1). Both solifluction and creep show substantial spatial and temporal variability (Rączkowska 1999; Baranowski et al. 2004). Frost action and solifluction are primarily associated with daily and seasonal freeze-thaw cycles of the ground, mainly in spring and autumn (Baranowski et al. 2004; Rączkowska 2007). The resulting slope microtopography takes the form of *terracettes* and *solifluction garlands*. A much less common feature is the bound solifluction lobe. The processes of frost heaving and sorting produce *patterned grounds* including polygons and circles with a maximum diameter of 1 m but mostly a few dozen centimeters. In addition to patterned grounds, in the Tatras and Beskids, frost processes also create *thufurs*.

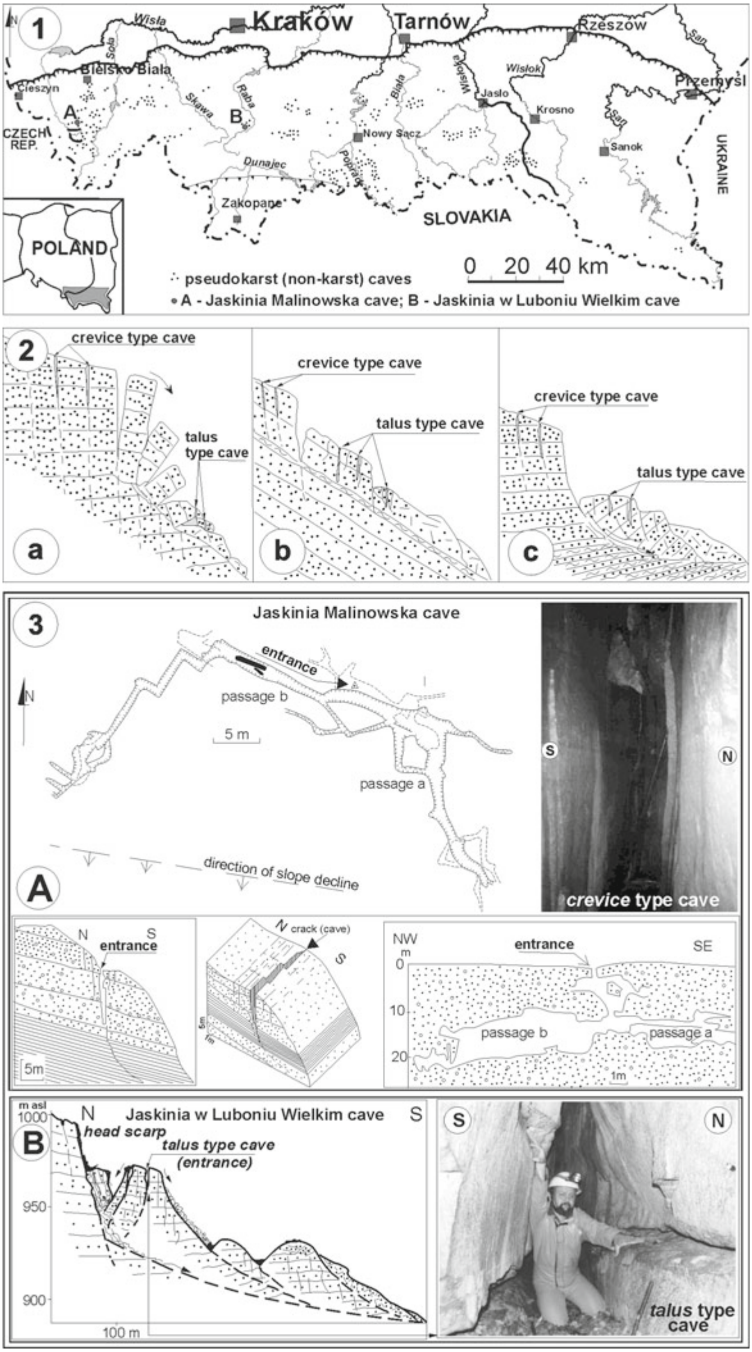
The role of *nivation* processes varies from case to case. In the Tatras, erosion processes associated with patches of snow lead to the fragmentation of slopes. *Nival niches* develop in areas occupied by snowpatches. The edges of nival niches retreat at a rate of 1–5 cm year^{-1}, while meltwater creates 20 cm deep gullies at the niche bottom or on the slope surface below. Accumulation nival niches develop on talus slopes (Rączkowska 1993, 1995, 2007). Protalus ramparts are rarely found and the height of active landforms does not exceed 1.5 m. The morphological role of nivation is limited to the Tatra Mountains and Babia Góra Massif.

5.8 Karst and Pseudokarst

Zofia Rączkowska
Włodzimierz Margielewski

Karst surfaces in the Polish Carpathians have a quite limited extension, as only fragments of the Tatras and the Pieniny Mountains are built of carbonate rocks. Given the low intensity of chemical denudation, their morphogenetic role is small under current climate conditions (Table 5.1). The highest rate of chemical denudation has been observed on carbonates in the forest belt of the Tatra Mountains (Kotarba 1972). All types of *karren* can be found on the northern-facing slopes of the Tatra Mountains, most commonly at 1,400–1,900 m elevation (Wrzosek 1933; Głazek and Wójcik 1963; Kotarba 1967, 1972). In addition, *dolines* can often become reflected in postglacial deposits in forested zones (Głazek 1964). Among the underground karst features in the Tatra massifs, foredeep-type *caves* and vertical caves with few speleothems can be mentioned (Kotarba 1972; Głazek and Grodzicki 1996). In the Pieniny Mountains, there are smaller caves less in number (Zarzycki 1982).

In the rock massifs of the Flysch Carpathians, non-karstifying rock cavities of any size are often called *pseudokarst* landforms (Vitek 1983). The largest of them can be over 20 m high and can exceed 1,500 m in length (Pulina 1997; Klassek and Mleczek 2008). They are quite common in the Polish Carpathian region. Until now, over 1,000 features of this type have been identified (Fig. 5.11: 1) (Klassek and Mleczek 2008). In most cases, they are partly of mass movement origin and can be called crevice-type (fissure caves) or talus-type caves (Fig. 5.11: 2a–c) (Vitek 1983). *Crevice-type caves* form as a result of the gradual release of shearing stress arising as a result of disturbances in slope equilibrium within rock massifs.


1
Kraków
Tarnów
Rzeszów
Wisła
Soła
Bielsko Biała
Cieszyn
A
CZECH REP.
Skawa
Raba
B
Biała
Wisłoka
Jasło
Wisłok
Krosno
Przemyśl
San
Sanok
UKRAINE
Nowy Sącz
Dunajec
Poprad
Zakopane
SLOVAKIA
POLAND
0 20 40 km
pseudokarst (non-karst) caves
A - Jaskinia Malinowska cave; B - Jaskinia w Luboniu Wielkim cave
2
crevice type cave
talus type cave
a
b
c
3
Jaskinia Malinowska cave
entrance
passage b
5 m
passage a
direction of slope decline
A
S
N
crevice type cave
N
S
entrance
5m
crack (cave)
NW
SE
m
0
10
20
1m
B
Jaskinia w Luboniu Wielkim cave
m asl
1000
950
900
head scarp
talus type cave
(entrance)
100 m
talus type cave

The relaxation of such stress can also take place along the surface of tectonic discontinuities such as fractures and faults, which slowly widen to cracks. This process continues until the massif becomes destroyed and gravity displaces the fragments of rock separated from the main rock mass by a system of cracks (Margielewski and Urban 2003, 2005). Crevice-type caves are an initial step in the triggering of mass movements on slopes that have not been subject to gravitational displacement up to this point (Fig. 5.11: 3A). They can also form unloading joints above landslide scarps (Fig. 5.11: 2). A unique kind of crevice-type landform are fissure caves featuring fractures along the slipping zones of landslides, called fissure macrodilatancy (Margielewski et al. 2007). *Talus-type caves* form within landslides (Fig. 5.11: 2) and can be found in gravitationally displaced rock packets (forming along fractures and faults) (Fig. 5.11: 3B), amidst chaotically arranged blocks of rock (Pulina 1997; Margielewski and Urban 2003). Speleothems are rarely found in fissure caves in flysch (Urban et al. 2007a, b). In addition, the flysch Carpathians also abound in *weathering-type caves* (fissure-type, bedding-type), partially of human and complex origin (Waga 1993; Urban and Otęska-Budzyn 1998).

5.9 Fluvial Processes

Adam Łajczak

The Polish part of the Carpathian Mountains is normally drained to the north by mountain and foothill tributaries of the Vistula River. At least 10% of this area is made up of valleys and basins. It contains Pleistocene and Holocene dry terraces that have remained intact to some extent, floodplains, river channels, old alluvial fans, and younger accumulating alluvial fans (Klimaszewski and Starkel 1972; Klimek 1979, 1991; Starkel 1991). The size of each fluvial landform depends on valley width and the sediment transport by the river flowing through the valley. The largest of fluvial landforms are found in mid-mountain basins and wider sections of large river valleys.

The rate of transformation of river channels, floodplains, and active alluvial fans depends on the frequency and magnitude of floods, the quantity of material transported

Fig. 5.11 Pseudokarst forms in the Polish Carpathians. 1, distribution of non-karst caves (pseudokarst caves) in the Polish flysch Carpathians (After Pulina 1997; compiled by Margielewski and Urban 2003, supplemented); 2, location of typical non-karst (pseudokarst) forms: crevice and talus type caves within mass movement forms, as: *a* topple, *b* translational slide, *c* rotational slide; 3, examples of non carst caves: *A* typical crevice type cave developed on slope not transformed by mass movements: Jaskinia Malinowska cave, Beskid Śląski Mountains (Map after Rachwaniec and Holek 1997; cross section after Ganszer and Pukowski 1997), *B* talus type cave in block landslide body of landslide (compound type) on Luboń Wielki Mountain, Beskid Wyspowy Mountains (After Margielewski 2006b). Position of both caves on the map above (Photo: W. Margielewski) by rivers, and on human activity over the past few centuries. Older fluvial landforms found beyond the reach of peak floods are inactive. The construction of dams results in rapidly developing deltas, even large ones. These include channels, islands, levees, inter-levee basins, crevasses, and crevasse splays (Klimek et al. 1990; Łajczak 2006). Beaver activity is an additional factor that helps shape valley floor relief.

River channels are the most dynamic fluvial landforms. Their morphology is linked to variable lithological, orographic, and climatic conditions, vegetation cover, and past and present land use (Klimek 1979). Suspended sediment transport increases towards the east in large rivers, while bedload transport is reducing (Łajczak 1999). For this reason, west of the Dunajec River, most rivers have gravel beds, while rivers to the east tend to have long, wide sections of rock bed. Fluvial transport and fluvial landform dynamics are the most intense during summer flood events in the western part of the region. In the eastern part, on the other hand, this occurs during snowmelt. The delivery of material to channels is uneven along the lengths of major rivers depending on geological structure, relief, and land use in basins. Other factors of influence include channelization efforts and the construction of dams. The Polish part of the Carpathians constitutes 30% of the drainage basin of the upper Vistula but delivers 70% of the water and as much as 90% of the suspended sediment (Łajczak 1999). Spatially variable material delivery leads to the formation of eroded sections, transit-only sections, and accumulation sections that often alternate along the course of a river (Kaszowski et al. 1976; Froehlich et al. 1977; Kaszowski and Krzemień 1977; Klimek 1987, 1991).

Two periods, different in terms of the intensity and direction of fluvial landform development, have been identified for the period of human economic activity in the Polish Carpathians. The first period (from the seventeenth to the nineteenth century) is associated with agricultural colonization and logging activity, which have led to increased water flow dynamics in basins and increased material delivery to river channels. The second period, lasting since the late nineteenth century, has included river channelization measures and the construction of dams. The original purpose of human intervention was to limit the extent and intensity of floods as well as to affect fluvial landform dynamics.

Flood dynamics increased in the western part of the Polish Carpathians in the nineteenth century and involved a growing concentration of flood waves and increased discharge, which further enhanced sediment transport (Niemirowski 1974; Wyżga 1993a, b). Historical documents and hydrological research begun in the nineteenth century indicate that floods on the Soła River floodplain became ever more frequent from the fifteenth century to the end of the nineteenth century. The Soła is a typical gravel-bed river of the Polish Carpathians (Łajczak 2007a). The dominant tendency of floodplain accumulation and channel widening continued in the western Polish Carpathians (including the Dunajec River) until the end of the nineteenth century (Fig. 5.12: B). At times, the rivers in the region inundate the floodplain and form wider braided channels (maps by Mieg 1779–1782, Kummerer 1855, and “Die Spezialkarte…” 1894).

Large-scale, river-based transportation of logs led to the exposure of bedrock along upstream sections of rivers in the Beskidy Mountains in the nineteenth and the early twentieth centuries. Material removed along these sections of river would

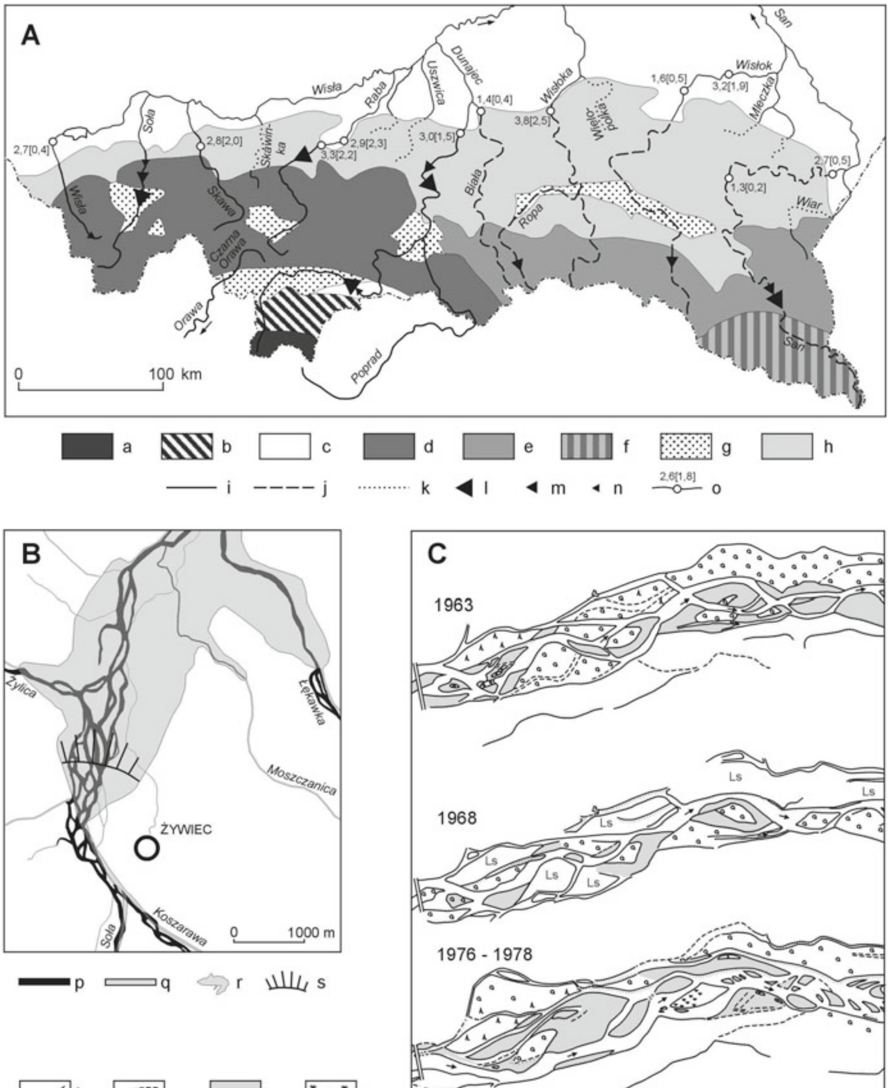


Fig. 5.12 Distribution of different types of river channels in Polish Carpathians (A), transformations of the Soła river channel in Żywiecka Basin since the 1850s (B) (After Starkel and Łajczak 2008, supplemented); transformations of the Białka river channel in Podhale, 1963–1978 (C) (After Baumgart-Kotarba 1983). Types of channels: a, Tatra; b, sub-Tatra; c, Pieniny; d, western Beskidy; e, eastern Beskidy; f, Bieszczady; g, intermontane basin; h, foothills; i, Vistula western Carpathian transit tributaries; j, Vistula eastern Carpathian transit tributaries; k, local foothill rivers. Dam reservoirs of the following water-storage capacity: l, >100 million m^3; m, 10–100 million m^3; n, <10 million m^3; o, deepening [m] of river channels in the selected water gauging stations during the twentieth century, a value in *brackets* concerns the second half of the twentieth century (After Wyżga and Lach 2002); p, a pattern of the Soła river channel according to the map by C. Kummerer (1855); q, present pattern; r, limit of backwater of the dam at Tresna on the Soła river (Żywiecki Reservoir) at the medium river stage; s, present limit of the delta front of the Soła river in the reservoir; t, clear escarpment; u, escarpments with gentle profile; v, fresh gravel bar; w, grass-covered bars; x, bars vegetated by willow and young forest; y, forest; z, channels with flow direction

build up alluvial fans and braided channels in the valleys of major rivers. The process of the *alluviation of valley floors* was associated with human activity and happened to coincide with climate change during the Little Ice Age. The gravel beds of rivers in the Polish Carpathians became shallower in the nineteenth century as a result of the introduction of potato cultivation (Klimek and Trafas 1972).

River engineering efforts commenced at the turn of the twentieth century started a trend of unprecedented river incision and this trend continues to this day. *Debris check dams* made of stone along the upstream sections of rivers and their tributaries halt bedload, allowing for the formation of small gravel-type alluvial fans. There is a tendency for these types of fluvial landforms to have coarser load towards the surface of sediment deposits. *Channelized streams* acquired a step-like longitudinal profile. Embankments made of stone or fascine stabilize the channel system. In the case of wide braided river channels, they create a narrow flow path and limit the amount of material being flushed out from the bases of riverbanks (Wyżga 1993a, b). The key reasons for Polish Carpathian rivers becoming deeper are the channelization of their downstream sections a century ago and of the Vistula River in the Polish Carpathian foreland (Klimek 1987; Wyżga 1993a; Łajczak 1999; Wyżga and Lach 2002). Water level measurements have shown that the downstream and middle sections of Carpathian tributaries of the Vistula River have become 2–3 m deeper during the twentieth century and often more than half of this change dates to the second half of the century. Land use change (that reduces suspended load transfer) has favorably affected upstream incision along eastern Polish Carpathian rivers since the 1940s. The same has been true of rivers across the entire Polish Carpathian region after 1989 (Łajczak 1999). The removal of coarse alluvia from river channels can lead to the exposure of bedrock (Soja 1977; Wyżga and Lach 2002; Krzemień 2003; Zawiejska and Krzemień 2004). The replacement of braided channels with single-thread channels, increasing channel depth, and riverbank gradient force a river to become more “efficient.” Channelized rivers can transport material farther than ever before (Starkel and Łajczak 2008) with the exception of sections located behind dams with reverse flow. This does not favor the accumulation of *floodplains* in Polish Carpathian valleys, especially basins, and the part of the Carpathian Foothills where artificial levees raise riverbanks. Higher rates of accumulation have been observed across the full width of valley floors but only upstream from dams (Łajczak 1994, 1996), where large *deltas* continue to build of sand and silt. Deltas in far upstream sections of rivers could also build of coarse material. Deltas in reservoirs show three distinct zones: the normally exposed topset zone, the partially exposed foreset zone, and the usually submerged bottomset zone. In light of large fluctuations in the water levels of reservoirs up to 12 m, the longitudinal profiles of deltas are not consistent with the Gilbert delta model (Klimek et al. 1990; Łajczak 2006). Deltas extended towards the dam with sediment (mainly loam) accumulation at the bottom of the reservoir. An underwater wall, several meters high, forms upstream the dam and parallel to it, at the spot with a rising current. Downstream of dams, river channels incise as their water is not loaded with sediment. Fine particles are flushed out from coarse alluvium (Malarz 2002).

Despite intense human activity, there are still sections of river exhibiting natural flow dynamics in the Polish Carpathian region. Streams of I–IV order were usually dredged in the 1960s, while V–VII order streams tend to experience aggradation that has been moving upstream in non-forested areas (Kaszowski et al. 1976). In the eastern part of the Beskidy Mountains, an alternating pattern of erosion and accumulation sections can be observed. The following *types of river channel* have been identified in the Polish Carpathians (Fig. 5.12: A):

(a) Stream channels in the Tatras that cut into moraines, glacio-fluvial formations, and solid rock (carbonate and crystalline), with appreciable ion transport; most stable in middle sections of valleys (Kaszowski and Krzemień 1979; Krzemień 1985, 1991);
(b) River channels in the Tatra foreland flowing atop Podhale flysch, incising into glacio-fluvial and alluvial formations of the braided type. Channelization and removal of pebbles in some sections has exposed bedrock (Baumgart-Kotarba 1983; Krzemień 2003; Kościelniak 2004; Zawiejska and Krzemień 2004);
(c) The river channel of the Dunajec, unique for the Polish Carpathian region, flows through a deep gorge in the Pieniny Mountains, transporting less material downstream below the cascade;
(d) Stream and river channels in deep valleys in the western part of the Beskidy Mountains that tend to be rocky upstream and alluvial downstream (also in alluvial fans), continue as wide alluvial channels, and still braided in many places (Ziętara 1968; Krzemień 1976, 1984b);
(e) Slowly widening channels in the eastern Beskidy Mountains, which tend to alternate between gravel bars and rocky channels (Starkel 1965; Kaszowski, Kotarba 1967; Klimaszewski and Starkel 1972; Izmaiłow et al. 2006);
(f) Channels of rivers in the High Bieszczady Mountains that fit somewhere between the (d) and (e) channel types with alternating rocky channels and channels with thick layers of bottom gravel;
(g) Gravel-bed river channels in middle mountain basins, which exhibit aggradation caused by neotectonics; gravel extraction modifying their evolution; neighboring meandering streams and rivers that drain peat bogs (Baumgart-Kotarba 1983; Krzemień 2003; Kościelniak 2004; Zawiejska and Krzemień 2004; Łajczak 2007b); levee-bound floodplains in large basins with only local accumulation (Starkel and Łajczak 2008);
(h) Stream and river channels in the Carpathian Foothills which receive primarily fine-grained material from undercut banks (Krzemień and Sobiecki 1998; Święchowicz 2002c); floodplains of rivers with no artificial levees built up of suspended load;
(i) Channels of large rivers in valleys of varying width, which used to exhibit aggradation prior to channelization and now serve as transport routes for large amounts of suspended and bedload (still increasing downstream in the wake of channelization); local gravel bar building (Klimek and Trafas 1972; Wyżga 1993a; Malarz 2002) and floodplain alluviation.

The deepening of major river channels in the eastern Polish Carpathians in the twentieth century has altered their former floodplains near the edge of the Carpathians into (up to 9 m high) terraces. The narrow *floodplains* are quickly accumulating. Despite a larger gradient, the channels of gravel-bed rivers in the western Polish Carpathians have not incised considerably and their narrow floodplains below the flood terrace are not aggrading so rapidly (Starkel 2001).

5.10 Biogenic Processes

Adam Łajczak

Biogenic landforms in the Polish Carpathians include raised bogs, accumulations of large woody debris in channels and beaver habitats. The largest of peat-covered areas can be found in the Orawsko-Nowotarska Basin, in the Western Bieszczady Mountains in the upstream San and Wołosatka valleys, locally in the Tatra Mountain valleys, and on the slopes of the Beskidy Mountains (Żurek 1983, 1987). Human impact is transforming the landscape in this part of the region.

In the Orawsko-Nowotarska Basin more than 50% of the area was covered by *peat bogs* prior to its agricultural colonization (Łajczak 2007b). Areas of low peat occupied three times more space than the 18 large peat domes, two of which reached lengths of 4 and 6 km. As peat has been extracted, drained, and burned for agricultural purposes on a larger scale since the eighteenth century, total peat area in this basin dropped drastically. As a result of peat removal and fragmentation, remnants of peat domes, with numerous hollows and drainage canals, are bordered by scarps as high as 6 m. Large post-peat areas became exposed with their mineral parent material and the *paleo-channels* where peat formation had begun (Wójcikiewicz 1979; Obidowicz 1990). The removal of peat was restricted after 1990 and a recovery of peat domes and former peat bogs ensued. Currently, there are 30 peat bogs in the Polish part of the Orawsko-Nowotarska Basin with a total area of 1,312 ha. The existing peat bogs, low peat and former peat bogs included, occupy 70,000 ha or 14% of the basin. There are 18 raised bogs in the Polish part of the Bieszczady Mountains and 17 are found in a valley (Ralska-Jasiewiczowa 1972, 1980, 1989; Łajczak 2011). The bogs had been subject to human impact until the 1980s. Given the advanced state of regrowth in these areas, the geomorphological effects of this process are impossible to assess. Small peat bogs can be found in the Sucha Woda Valley and the Pańszczyca Valley in the Tatras. There are some minor peat bogs on slopes in headwater areas and in the landslide zones of the Polish Beskidy Mountains, mainly concentrated on Mount Barania Góra (1220), Pilsko Massif (1557), at the foot of the Babia Góra Massif (1725), and in the Beskid Niski Mountains (Gil et al. 1974; Obidowicz 2003; Margielewski 2006a).

Permanent accumulations of *large woody debris* in the stream and river channels of the Polish Carpathians have been noted in nature reserves and national parks (Kaczka 2003, 2008; Wyżga et al. 2003). *Beaver habitats* are expanding in Polish Carpathian valleys and, along with accumulations of woody debris, tend to cause fluvial deposition.

5.11 Human Impact

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Jolanta Święchowicz

Different forms of human impact that cause changes in relief and the intensity of geomorphological processes on a local and a regional scale have been intensifying in the Carpathian Foothills since the Neolithic (Pietrzak 2002). In the mountain interior, human impact has been a factor since the end of the Middle Ages and reached its peak intensity in the nineteenth and twentieth centuries. Various forms of human impact have been identified, in chronological order: settlement, construction, agriculture, logging, pastoral activity, mining, metal manufacturing, timber rafting, transportation, summer and winter tourism, and river engineering. Each of them developed and reached its peak at different dates.

Human impact has affected different elevations in different ways and to different degrees. High mountain areas have experienced minor human impact with the western part of the Polish Carpathians affected by more intense impact than the eastern part. Certain types of human activity still affect relief and geomorphological processes (logging, construction, river engineering, tourism) while others have been largely abandoned, however, their effects are still noticeable (mining, pastoral activity).

Human impact in high levels of the Tatras has been a fact of life for at least 200–300 years (Libelt 1988; Kaszowski et al. 1988; Mirek 1996), but intensified in the late nineteenth and the first half of the twentieth centuries (Jahn 1979; Libelt 1988; Hreško et al. 2005; Kotarba 2004, 2005), manifested in *mining* and *metal smelting,* which have left behind numerous traces: mine entrances, roads, and quarries. Furthermore, intense *sheep grazing*, which lasted until the 1960s, caused the destruction of subalpine shrubs and compact alpine meadows. The result has been intensified erosion as well as the formation of erosion niches and cattle terracettes on slopes. It took about 20 years for erosion to slow down and biological aggradation to take place following the end of sheep grazing (Jahn 1979; Kozłowska and Rączkowska 1999; Rączkowska and Kozłowska 2002). On the other hand, the homogeneous forests planted to replace natural woodland are still susceptible to the uprooting of trees by foehn winds (Kotarba 1970; Koreň 2005). Today, human impact primarily results from summer *tourism* (Kozłowska and Rączkowska 1999; Gorczyca and Krzemień 2002) and is limited to the 0.5–3.0 m wide zone along trails, where geomorphological processes are more active (Gerlach 1959; Kotarba 1976; Kłapa 1980; Krzemień 1991; Krusiec 1996; Czochański 2000; Gorczyca and Krzemień 2002; Balon 2002). These processes start in areas characterized by destroyed vegetation cover and stone pavement on trails. In addition to summer tourism, mountain climbing and skiing also impact the landscape to some extent. *Logging* in the forest zone also produces rills that result from the dragging of logs downslope. Once logging routes are abandoned, they are gradually obliterated by natural processes (Dudziak 1974). The geomorphological effects of skiing and other forms of human impact have been investigated in the Pilsko Massif area (Beskid Żywiecki Mountains).

Changes in land use in the Polish Flysch Carpathians have altered drainage patterns as well as reduced water retention depth, which favors accelerated runoff. As a result, extreme precipitation events often lead to the activation of shallow *landslides* (Jakubowski 1965, 1968; Gil 1997; Poprawa and Rączkowski 1998; Gorczyca 2004). In addition, poorly constructed hillside homes as well as road and rail construction that undercuts slopes induce pressure and landslides. The end results can be catastrophic in economic terms and as threats to human life (Fig. 5.7: 4) (Ziętara 1968; Poprawa and Rączkowski 2003; Oszczypko et al. 2002; Bajgier-Kowalska 2004). Intensive stone and gravel extraction in river valleys also upsets the equilibrium of slopes and valley sides, which can lead to mass movements (Pietrzyk-Sokulska 2005). Finally, *reservoirs* built along Polish Carpathian rivers help induce mass movements along riverbanks (Ziętara 1974; Bajgier 1992).

Intense deforestation in the eighteenth and nineteenth centuries led to an expansion of arable land. The beginning of the twentieth century saw a rapid division of farmland into smaller plots, which produced a dense network of *field roads* and *plot boundaries*. This plot mosaic survives to this day and affects the course and intensity of geomorphic processes on cultivated slopes. Consequently, weathering processes have changed significantly. *Sheet and rill erosion* is an important factor in slope evolution, as a result of which, mineral material is transported down-slope and accumulate at the base. Soil erosion is also a key process responsible for the delivery of suspended load to stream and river channels (Gerlach 1966, 1976a, b; Gil 1976, 1999; Froehlich and Słupik 1980a, 1986; Froehlich 1982; Lach 1984; Święchowicz 2002c). Intense *gully erosion* occurs on forested mountain and foothill slopes where logging is a key human activity. Agricultural slopes are characterized by terraces, plot boundaries, and wooded scarps that limit the amount of material transported downslope (cascade-fragmented system) and change their longitudinal profiles from convex or convex-concave to step-like. On shorter foothill slopes with no plot boundaries, eroded soil deposited at the slopefoot forms distinct deluvial flats, which act as a barrier to newer eroded material heading for the channel. The transfer of eroded material takes place primarily via a system of downslope ruts and field roads (linear-continuous system). In the light of the recent economic tendency to decrease the amount of arable land in the Polish Carpathians, the hazard of sheet and rill erosion and aeolian processes reduces remarkably.

Material transport and fluvial form dynamics in the river channels of the Polish Carpathians are undergoing changes as a result of human interference in the hydrology of drainage basins, a trend started as early as the seventeenth century (Klimek and Trafas 1972; Froehlich 1982; Wyżga 1993a, b; Łajczak 1999, 2007a; Krzemień 2003). The greatest *impact on fluvial processes* and landforms can be attributed to:

(a) Deforestation, agricultural colonization, the introduction of root crop cultivation on slopes (even at high elevations) with fields being plowed in downslope direction, as well as the creation of a dense network of roads separating fields, which has resulted in substantially accelerated runoff and increased sediment delivery from slopes to stream and river channels (at least since the seventeenth century);

(b) Replacement of mixed forests in the lower forest belt with homogeneous spruce forests, another anthropogenic contribution to higher flood risk in the valleys of the Polish Carpathians and in the Vistula River valley of the Carpathian foreland as well as a key factor behind increased fluvial dynamics (nineteenth century);

(c) Creation of a dense network of forest roads, which leads to accelerated runoff and increased sediment delivery to channels (nineteenth and twentieth centuries);

(d) Channelization of streams and rivers designed to stabilize their banks; construction of debris control dams designed to halt the movement of bedload; and the construction of levees designed to decrease the size of the flooded zone in river valleys (twentieth century);

(e) Construction of dams, which permanently stop all bedload transport and up to 99% of suspended load (since 1932).

Changes in river channel morphology were rather gradual in the Polish Carpathians and have reflected the type of human impact predominant in the given period (Starkel and Łajczak 2008).

5.12 Regional Overview of Recent Landform Evolution

5.12.1 Landform Evolution in High Mountains

Zofia Rączkowska

Trends in relief transformation of high mountain slopes are dependent on slope type and elevation. *Rockwalls* and *rocky slopes* retreat by rockfalls and frost weathering and become fragmented and shortened by chutes and lowering (Kotarba et al. 1987). Chutes become deeper and wider through corrasion, avalanches, and debris flows (Kotarba 2002). *Debris slopes*, on the other hand, even at lower elevations with a denser plant cover, continue to accumulate. At the same time, they are fragmented by debris flow gullies and nival niches (Kotarba et al. 1987; Rączkowska 1995, 1999). Regolith slopes are shaped by cryonival processes, erosion, and mass movements, which lead to uniform transfer of slope deposits across the entire slope surface and finally to its degradation. Fresh debris flow gullies and erosion niches are rather rare.

Changes in the topography of the *valley floors* are usually limited, vary along the longitudinal profile, and are primarily restricted to stream channels (Kaszowski 1973; Baumgart-Kotarba and Kotarba 1979; Krzemień 1985, 1991; Kotarba 2002). Pleistocene glacial cirques are being filled with material delivered by debris flows and avalanches (Krzemień 1988; Kotarba 1992, 1995; Rączkowska 1999). The most significant changes occurring in valley floors, glaciated during the Pleistocene, affect footslopes through accumulation from debris flows and avalanches in the shape of poorly defined cones and tongues (Krzemień 1985, 1991; Kotarba 1992). Minor changes in stream *channel* morphology primarily arise from lateral erosion during flood events (Krzemień 1985, 1991). Major permanent changes along limited stretches of streams occur following extreme floods (e.g. in 1997) (Kotarba 1999). Despite relatively intense fluvial action, non-glacial valley floors are stable during event-free periods (Kaszowski 1973; Krzemień 1991). Bedrock channels exhibit a smoothing trend along their longitudinal profile (Kaszowski 1973). Major changes in channels are caused by extreme hydrometeorological events, which may completely flush out deposits (Kaszowski 1973; Kotarba 1999).

In general, debris flows, dirty avalanches, and extreme floods are crucial in the transformation of the relief. For this reason, the stabilizing trend after the Little Ice Age and with reduced human impact may reverse as the frequency of extreme events is increasing.

5.12.2 Landform Evolution in Middle Mountains

Włodzimierz Margielewski
Jolanta Święchowicz

In middle and low mountains, *linear (gully and stream) erosion* is central in the transformation of slopes during catastrophic precipitation events, leading to the incision of valleys, field roads, and logging roads, as well as to the initiation of surface mass movements (Ziętara 1968). Leaching, piping, and the uprooting of trees by wind are also very important, while sheet erosion is negligible on wooded and grassy slopes in the Beskids.

Numerous shallow landslides affect weathered bedrock (Fig. 5.7: 4 and 5. 8). Valley deepening and widening during floods induces mass movements, which transform the sides of valleys (Ziętara 1968). In some cases, landslides can shape large slope surfaces and their development is often stimulated by human activity (Ziętara 1968; Oszczypko et al. 2002; Bajgier-Kowalska 2004). However, given their discrete occurrence, their contribution to general slope evolution is limited and local. Furthermore, rock slides, deep-seated and extensive landslides, are rather rare in modern times (Fig. 5.7: 5). What happens most of the time can be termed a *reactivation of old landslides* (as observed in 1958–1960 and 1997–2002) involving only minor changes (local widening of valley sides and negligible retreats of valley heads – Ziętara 1968). Significant changes in the relief of the Beskidy Mountains only occur over the longer term (in thousands of years) and include the retreat of valley heads, the lengthening of river valleys, and the formation of tributary valleys and concave landslide slope profiles (Starkel 1960; Ziętara 1968; Kotarba 1986). In general, contemporary landslides are combined with mudflows, soil creep, sheet erosion, piping in unconsolidated material, erosion, and weathering (Fig. 5.8 upper left photo) (Ziętara 1968).

Terraced cultivated slopes are far less affected by sheet and rill erosion. Material is transported from slope surfaces directly to river channels only during heavy and lasting rainfalls. In such cases, natural rills, gullies, and field roads serve as main transport routes (Froehlich and Słupik 1980a, 1986; Froehlich 1982).

5.12.3 Landform Evolution in Foothills

Włodzimierz Margielewski
Jolanta Święchowicz

The Carpathian Foothills are used for agricultural purposes and shaped by soil erosion and mass movements, while in valley bottoms sediments transported downslope are deposited. In contrast to the Beskids, the Carpathian Foothills tend to be characterized by long fields perpendicular or diagonal to contour lines with no stabilized plot boundaries (Święchowicz 2001). Downslope plowing of the slope intensifies erosion processes quite substantially. *Sheet, rill,* and *ephemeral gully erosion* after heavy rainfalls particularly affects commercial farms with large acreage of crops (Święchowicz 2008, 2009). In foothill catchments, slope material is delivered to stream channels only in small amounts and during extreme precipitation events. Even then, most of the transported material is deposited close to footslopes, on deluvial plains, and across valley floors. This produces landforms typical of the Carpathian Foothills: *convex-concave, multi-segmented and step-like agricultural slopes* as well as *flat valley floors* with occasional deluvial surfaces (Święchowicz 2001, 2002a, b, 2008, 2009). Aeolian activity is only important in certain regions (e.g., in the Jasło-Sanok Depression), reducing slope inclination and transforming slope profiles (Gerlach and Koszarski 1968; Gerlach 1986).

Mass movements are important and sometimes predominant geomorphic processes. Although many *landslides* induced by extreme precipitation may be small (Fig. 5.7: 1–2), they are numerous. They make slopes steeper and smoother as well as help form wider river valleys. In general, however, they smooth out all morphological irregularities on slopes, creating *slopes of undulating profile* (Starkel 1960; Gorczyca 2004). Landslides in the heads of minor tributary valleys make them extend upslope, leading to crest fragmentation, overall crest lowering, and narrower ridges over the long term in the Beskids (Kotarba 1986).

Contemporary changes in land use (less arable land and more woods and meadows) reduce sheet erosion, but increase gully erosion and channel incision and also favor shallow landslides (Jakubowski 1964).

5.12.4 Valley Floor Evolution

Adam Łajczak

The longitudinal profiles of rivers in the Polish Carpathians and the Carpathian foreland are undergoing changes caused by *river engineering*, which include dam construction and other modifications of river channels and floodplains, which have reached unprecedented extents locally (Starkel and Łajczak 2008). Prior to river engineering, the gravel-bed braided rivers in the Polish Carpathians used to incise their channels upstream, aggrade, and widen downstream, allowing the inundation of floodplains. The erosional sections of rivers draining the eastern part of the Polish Carpathians used to be longer than those of rivers in the west. It was at that time that channels tended to become deeper downstream. Lower floodplains were starting to form, although aggradation also happened along the meandering sections of rivers. River engineering efforts and changes in land use led to the lengthening of erosional sections of river channels. Today, most rivers are eroding virtually along their entire length. Large-scale *deposition* of (mainly suspended) material occurs only in the backwaters of dams, especially *in long and deep reservoirs*. The suspended load accumulates in the flooded channel and across the former floodplain. *Channel incision* intensifies *below dams* and former floodplains are turned into dry terraces. Three dynamic types of longitudinal river channel and floodplain profiles have been identified in the Polish Carpathians in recent years, reflecting current land use trends (Starkel and Łajczak 2008):

(a) The gravel-bed type with an incised downstream channel and large growing deltas in reservoirs;
(b) Eastern Carpathian type with mainly suspended matter transports and currently incision virtually along the entire river length;
(c) Foothill river type with a floodplain undergoing rapid deposition.

5.13 Conclusions

Adam Łajczak
Włodzimierz Margielewski
Zofia Rączkowska
Jolanta Święchowicz

Deforestation and cultivation, particularly in the Carpathian Foothills, have made *sheet erosion* the predominant geomorphic process during the past 200 years. Material washed off fields builds up across deluvial flats at the base of slopes, which encroach onto flat valley floors. The steeper slopes of the Beskids show agricultural terraces that stabilize slope surfaces with fields that run perpendicular to contour lines, which results in limited sheet erosion. Gullies and field roads play key roles in the transport of surface material from slopes. In certain parts of the Polish Carpathians, intense aeolian processes lower crests and windward slopes, while leeward slopes build up and become longer.

Over the last 20 years the amount of *agricultural land* tends to *decrease* along with the rate of sheet erosion and deflation, while gully erosion is intensifying on slopes, river channels are incising, and shallow landslides are reactivating. The reduced delivery of fine-grained material from slopes to valley bottoms as well as the channelization of rivers since the early twentieth century has resulted in deeper river channels and reduced rates of sediment deposition on floodplains. The opposite is true in valley floors upstream from dams. On the other hand, rivers in the Carpathian Foothills that drain areas still dominated by agriculture are characterized by wide floodplains that continue to collect sediment during flood events.

Only high-energy precipitation can normally lead to simultaneous transformations of slopes and valley floors in high mountain, middle mountain, and foothill areas. Despite the creation of terraces and the construction of all types of dams on valley floors, localized *debris flows* overrun barriers and transport material along the full lengths of slopes and river channels, which carry suspended sediment beyond the Polish Carpathian region.

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