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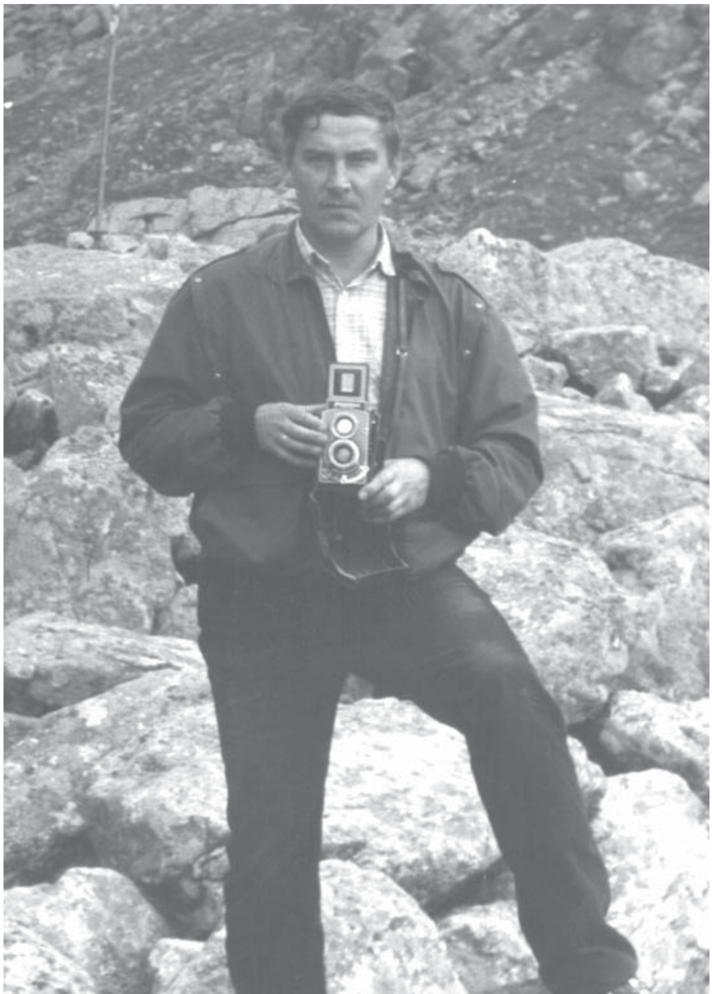

Professor Adam Kotarba during his field studies in the Tatra Mts. in the 1980’s.

PREFACE

The volume is dedicated to Professor Adam Kotarba on the occasion of his 70th birthday.

Prof. Kotarba was born at the foot of the Tatra Mts., and the highest range in the Polish Carpathians inevitably turned his scientific attention to the past and present-day evolution of mountains. This combination of directions—the parallel study of contemporaneous processes and their palaeogeographic roots—is a characteristic feature of the Cracow school of geomorphology. Prof. Kotarba's particular interest has centred on the mechanisms and rates of geomorphic processes transforming the slopes and valleys of the Tatra Mts. This area has been like a field laboratory when it comes to the understanding of the role played by a whole set of geomorphic processes, starting from chemical denudation, and proceeding via mechanical weathering, as well as gravitational, fluvial and eolian processes especially clearly expressed in the landscape during catastrophic events. In parallel to all this, by studying the ways in which the above find reflection in the transformation of talus slopes and in lake deposits, Prof. Kotarba identified several phases of frequent extreme events following the retreat of the valley glaciers. He carried out the comparative studies of similar phenomena in the Alps and other high European mountains, as well as lower belts of the Polish Carpathians.

The volume presented here, entitled "Modelling of high-mountain relief", includes papers from Prof. Kotarba's friends and co-workers relating to various of the phenomena and processes that have transformed the high-mountain landscapes over the whole postglacial period, as well as recently. Among the works are papers characterising geomorphic processes acting in the alpine belt of the Western Carpathians (B. Gądek, J. Hreško *et al.*, R. Midriak, Z. Rączkowska) and their complexity in the Sudetes (P. Migoń) and the Massif Central, France (K. Krzemień). Two papers then draw attention to the role through the whole postglacial period in deglaciated high mountains that has been played by chemical denudation (M.F. Andre), as well as to the rate at which rockfalls take place in the Canadian Rockies (B. Luckman). Other papers focus on the creative role of processes acting after deglaciation, i.e. periglacial processes in the Gaspesie Mts. (H. French and J. Bjornson), and gravitational transformation in the Italian Alps (F. Dramis and M. Guglielmin).

All the authors and guest editors are keen to wish Adam the best of good health and future scientific success.

Zofia Rączkowska
Leszek Starkel
December 2008

QUANTIFYING HOLOCENE SURFACE LOWERING OF LIMESTONE PAVEMENTS IN PREVIOUSLY GLACIATED ENVIRONMENTS

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Abstract: Valuable estimates of Holocene solutional erosion rates are offered by ice-scoured limestone pavements which provide adequate reference surfaces and reliable chronological control. From Spitsbergen to the equatorial mountains and hyperhumid Chilean Patagonia, increasing rates of surface lowering due to carbonate solution coincide with increasing precipitation amounts. Additional data from other environments confirm the primacy of this climatic control, but also point to the need for rock control (e.g. porosity and jointing) to be taken into account in comparative studies. As shown by Goldie (2005), when obtained in densely jointed and/or bedded limestones, the so-called 'solution rates' correspond to rates of mechanical weathering and cannot be compared to genuine solution rates obtained on compact and massive carbonate outcrops.

Key words: karst, limestone pavement, Polar and Alpine environments, postglacial solution rates, climatic control, rock control

INTRODUCTION

The evaluation of rates of weathering, slope retreat and debris accretion has been a major concern for the community of geomorphologists over the past half-century. In Europe, Adam Kotarba has played a major role in this research field through slope monitoring (e.g. Kotarba 1983), lichenometric dating (e.g. Kotarba 1989; 2001) and lake-sediment coring (e.g. Kotarba 1996), the special focus having been on the Polish Tatra Mountains.

In high mountains, as in polar areas, limestone pavements deglaciated at the end of the Würmian/Weichselian period offer conditions favourable to an evaluation of the amount of postglacial surface lowering due to solution. In the 1950s and 60s, pioneer investigations were conducted in European mountains by Corbel (1957) and Bögli (1961). Since then, numerous quantitative data have been collected from carbonate rocks of Alpine areas by European karst geomorphologists (Bauer 1964; Haserodt 1965; Julian *et al.* 1972; Nicod 1976; Kunaver 1979; Julian 1980; Maire 1990). More scattered data were also collected in Arctic and Subarctic areas (Åkerman 1983; Dionne and Michaud 1986; André 1996, 2002), including the Patagonian archipelagos (Maire *et al.* 1999; Hobléa *et al.* 2003).

The two objectives of this contribution are: (i) to provide an overview of surface lowering data obtained on Alpine and Polar deglaciated carbonated pavements; and (ii) to discuss the controlling factors behind the measured solution rates, partly based on a comparison with similar data from limestone pavements and marble tombstones in different environments.

1. LIMESTONE PAVEMENTS IN COLD ENVIRONMENTS: THEIR ATTRIBUTES AND VALUE IN THE ASSESSMENT OF RATES OF SURFACE LOWERING

1.1. SELECTED ALPINE AND POLAR SITES

All sites selected for the present comparative study are representative of the *Flachbarren* topography as defined by Bögli (1976). Within this karst landform family, two glaciokarstic subtypes are represented: the *Schichtreppenkarst* (presenting structural benches modified by ice) and the *Rundhockerkarst* (i.e. karst glacially scoured into *roches moutonnées*). Both types are free of vegetation, except for rare chasmophytes and rupicolous plants, and they usually display erratic blocks abandoned by the retreating ice. Though all pavements consist of carbonate rocks, genuine limestone is found mostly in the Alps, whereas dolomite and marble are more abundant in polar and subpolar sites. Last but not least, the annual rainfall of the selected cold-climate sites ranges from 400 mm (in the High Arctic) to 7300 mm (in the Patagonian islands).

1.2. REFERENCE SURFACES

The evaluation of the rate of surface lowering requires reliable reference surfaces (or 'zero-datum levels'), such as those offered by the glacially-scoured surfaces. Five minor-scale landforms are of special interest:

- The so-called *Encoches de banquette*, which "correspond to the dissolved layer from the upper bedding plane revealed by quarrying" (Maire 1990, p. 382; *cf.* Fig. 1A);
- the siliceous nodules made of flint or chert (Fig. 1B), used from early times on British tombstones (Goodchild 1875), and more recently by Dionne and Michaud (1986) in Subarctic Quebec, and by Maire (1990) in the French Alps and Pyrenees, and in northern Greece;
- the quartz and aplite veins polished by the Weichselian/Wisconsinian ice in the Scandinavian and Canadian Arctic (Åkerman 1983, Dionne and Michaud 1986, André 2002; *cf.* Fig. 1B);
- the glacially-scoured quartzite layers, within which dolomite layers are interbedded, that are commonly observed in North Scandinavia (e.g. André 1996, *cf.* Fig. 1C);
- the erratic pedestals (Fig. 1D), as observed some time ago by Corbel (1957) in

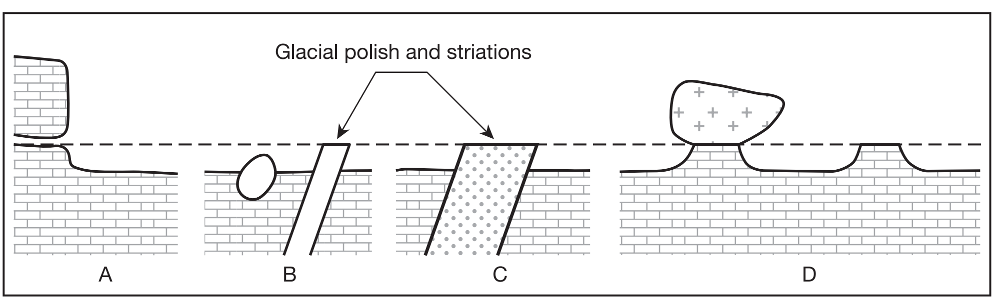

Figure 1. Minor-scale geomorphic features used to evaluate the amount of surface lowering due to solution since deglaciation on Polar and Alpine limestone pavements.
A. '*Encoche de banquette*'—B. Protruding siliceous nodules and quartz veins—C. Protruding quartzite layer—D. Erratic pedestal.

Lapland and systematically used in the Alps (e.g. Bögli 1961) as in Chilean Patagonia (Maire *et al.* 1999; Hobléa *et al.* 2001), where they are particularly spectacular (Fig. 2); the use of these pedestals by geomorphologists was inspired by the Alpine 'tables de glaciers', providing reference surfaces by which to assess the surface lowering of glaciers due to ice melting (Fig. 3).

Figure 2. Erratic pedestal in limestones of Monte Roberto, Madre de Dios, Chilean Patagonia. © Richard Maire (Ultima Patagonia 2006, Centre Terre). The boulder cap is made of volcanic rock.

Among these five minor-scale geomorphic features, erratic pedestals are by far the most commonly used in Alpine regions, whereas quartz veins offer the most widespread reference surfaces in the Arctic. As stressed by Maire (1990, p. 382), the most reliable results are obtained when a site yields similar heights for erratic pedestals and '*encoches de banquette*'. In contrast, when found alone, protruding siliceous nodules only provide minimal estimates of postglacial surface lowering.

1.3. CHRONOLOGICAL CONTROL

The dating of deglaciation is a prerequisite if rates of surface lowering are to be calculated. Radiocarbon dating is the main dating method at many sites. For instance, in Swedish Lapland, over fifty peat and gyttja samples have been dated in the Torneträsk Basin (Sonesson 1974), suggesting a deglaciation at c. 10,000 B.P., as confirmed by recent cosmogenic dating (Dixon *et al.* 2002). Radiocarbon dating is also available for many Alpine sites, and more locally for the equatorial mountains (e.g. Hope *et al.* 1976). In Chilean Patagonia, U/Th dating of the oldest stalagmites at 9600 B.P. provides a minimum age for the last deglaciation (Richard Maire, personal communication, 17 July 2008). At most sites, the reference period covers about 10,000 years. However, the deglaciation is more recent at some Arctic sites located close to the sea, which have been submitted to glacioisostatic rebound. Around the Hudson Bay, the emergence of dolomitic pavements dates back to a 6500 BP (Dionne and Michaud 1986), whereas on the West Spitsbergen strandflats, this emergence varies from 10,000 to 2000 BP (Åkerman 1983). Further north, lichenometry indicates an exposure time of c. 3500 yr for large marble blocks submitted to solution in rock glaciers (André 1993), though these blocks cannot be considered *in situ* outcrops.

solution rates, each receiving successively greater amounts of precipitation:

- the dry High Arctic (P= 400 mm and SR=2 mm ka^{-1})
- the rather dry continental Subarctic (P=700 mm and SR=5 mm ka^{-1})
- the humid temperate high mountains (P=2 000 mm and SR=13 mm ka^{-1})
- the very humid equatorial high mountains (P=4 000 mm and SR=30 mm ka^{-1})
- the hyperhumid subantarctic archipelagos (P=7 000 mm and SR=80 mm ka^{-1})

Figure 3. 'Glacier Table' on the Aar Glacier (Alps).
© G. Lory fils, Neuchâtel, 1822.
Such stones protecting the glaciers from ice melting were a source of inspiration for karst researchers who used erratic pedestals to quantify the surface lowering of limestone pavements due to solution.

2. SPATIAL VARIABILITY TO SOLUTIONAL EROSION RATES AND THE PRIMACY OF CLIMATIC CONTROL

The compiled average solution rates (SR) are expressed in mm ka^{-1} and appear in Table 1. They range from 2 to 80 mm ka^{-1} and are clearly controlled by the amount of annual precipitation. Five different regions can be distinguished on the basis of the obtained

The importance of pluviometric control of carbonate weathering rates was also demonstrated in other environments on the basis of detailed studies of marble tombstones. Meierding (1981) investigated tombstones made in Vermont marble across the United States from the semi-arid Nevada (P=200 mm) to the humid east coast (P=1000 mm), in areas unaffected by industrial and urban pollution. On the exposed upper portions of stone faces, he found rates ranging from 1 mm ka^{-1} in Nevada to 16 mm ka^{-1} on the East Coast. Similarly, Neil (1989) measured rates of weathering of Carrara marble tombstones in eastern Australia, based on the height of lead inscriptions. Results were obtained at sixteen sites from 17° to 35°S latitude, at which precipitation ranges from 500 to 2300 mm. Again, the measured rates of surface retreat—from 0.9 mm ka^{-1} to 6 mm ka^{-1}—are in direct proportion with mean annual rainfall.

However, closer examination of Table 1 reveals substantial differences in precipitation between the different temperate mountains, as well as apparent anomalies in the associated solution rates. Minimum values are found in the Brenta Dolomites (P=1500 mm) and maximums in the Dinaric Alps (P=3400 mm). The corresponding solution rates are respectively 15 mm ka^{-1} at the driest site and only 13.5 mm ka^{-1} at the more humid site (Nicod 1976; Kunaver 1979). More generally, temperate mountains are characterised by relatively low values (8–13 mm ka^{-1}) where sites have abundant precipitation (P=2500-3400 mm), whereas the highest solution rates (15–17.5 mm ka^{-1}) are found at sites with lower precipitation (P=1500-2000 mm). The significance of this apparent anomaly is three-fold: (i) methodological biases can exist (with uneven conditions of formation and/or conservation of the erratic pedestals); (ii) the validity of the chronological control can be questioned; (iii) differences in rock material (porosity, jointing) can be assumed to account for differences in rates of surface lowering. This last possibility deserves particular attention.

3. ROCK CONTROL AND IMPLICATIONS

A close look at Table 1 reveals that rock materials on the list are of various carbonate lithologies. Aside from genuine limestones, dolomites and marbles are also present. Moreover, no data on porosity and jointing are presented, which weakens the comparison between studies.

The importance of rock control should not be minimized, as the extremely high solution rates recently measured in Japan indicate. Matsukura *et al.* (2007) investigated pedestal rocks on Holocene raised coral-reef terraces on a subtropical Japanese island where caprock boulders are assumed to have been transported by large tsunami- or typhoon-generated waves. They obtained a mean lowering rate over 6000 years of 205 mm ka^{-1}, pulverizing the Patagonian record of 100 mm ka^{-1}. The main reason is the sensitivity to solution of the very high (up to 36%) porosity spongelike coral-reef limestone and the high organic acid and biogenic CO_2 contents of subtropical waters, which compensates for lower precipitation totals (only one-third as great as in Patagonia).

The importance of this last factor (i.e. water aggressivity) should not be overestimated, as demonstrated by Goldie (2005) in her re-evaluation of the solution rates obtained on the British limestone pavements. In the counties of Yorkshire (England) and Leitrim (Ireland), Sweeting (1966) and Williams (1966) had obtained high solution rates of up to 42 mm ka^{-1}, and later Trudgill (1986) accounted for the intensity of solution by invoking the acidity of waters, based on detailed MEM measurements at the surface of Yorkshire limestone pavements. To Goldie (2005), these high rates should not be considered as solution rates, as they were recorded in densely jointed and/or well laminated limestones. At such sites, mechanical weathering (including frost action) dominates over carbonate solution, as suggested by Ollier's drawing (Fig. 4) of the Craven pedestals. These are quite similar to the erratic pedestals found in the fissured basaltic outcrops of South Iceland (observations by Etienne and André, 16 July 1999). Finally, Goldie (2005) considers that the only relevant solution rates for the British pavements—at 3–13 mm ka^{-1}—are those measured on compact and massive limestones, such as those investigated in the Burren (Eire) by Williams (1966, 1970).

Though one cannot discard Goldie's analysis, it must be stressed that it is not

Table 1. Postglacial solution rates on polar and high-alpine limestone pavements (after Maire 1990 and André 1996, updated).

CLIMATE ZONE Location (lat/alt)	Lithology	Rainfall (mm yr^{-1})	Geomorphological indicators of Postglacial surface lowering	Period (years)	Average solution rate (mm ka^{-1})	Source
ARCTIC						
West Spitsbergen (Cape Linné, 78°N)	Dolomitic limestone	430	Protruding veins	10,000 (max.)	2.50	Åkerman 1980
Northwest Spitsbergen (Blomstrand, 79°N)	Fractured marble	380	Protruding veins	3,500	2.80	André 1993
All sites (mean)	—	400	—	—	2.65	—
DRY SUBARCTIC						
Swedish Lapland (Abisko Mts., 68°N)	Dolomite	800	Protruding quartzite layers	10,000	5.30	André 1996
Quebec, Canada (Hudsonie, 56°N)	Dolomite	600	Protruding veins and nodules	6,500	6.00	Dionne and Michaud 1986
All sites (mean)	—	700	—	—	5.60	—
TEMPERATE MOUNTAINS						
Swiss Alps (Plaine morte, 2 600 m)	Limestone	2,500	Erratic pedestals	10,000	8.00	Maire 1990
French Pyrenees (Pierre St Martin, 2 000 m)	Limestone	2,500	Erratic pedestals, protruding nodules, 'encoches de banquette'	10,000	9.00	Maire 1990
French Alps (Désert de Platé, 2 300 m)	Limestone	2,900	Erratic pedestals, 'encoches de banquette'	10,000	12.00	Maire 1990
Kanin Mt., Yugoslavia (2 100 m)	Limestone	3,400	Erratic pedestals	10,000	13.50	Kunaver 1979
Swiss Alps (Märenberg, 2 200 m)	Limestone	> 2,000	Erratic pedestals	10,000	14.50	Bögli 1961
Southern French Alps (2 000 m)	Limestone	1,500	Erratic pedestals	10,000	15.00	Julian 1980
Brenta Dolomites, Italy (2 500 m)	Dolomitic limestone	1,500	Erratic pedestals	10,300	15.00	Nicod 1976
Sennes Dolomites (2 400 m)	Dolomitic limestone	1,500	Erratic pedestals	> 10,000	10–15	Sauro *et al.* 1995
Austrian Alps (2 000 m)	Limestone	2,000	Erratic pedestals	10,000	17.50	Bauer 1964; Haserodt 1965
All sites (mean)	—	2,300	—	—	13.00	—
EQUATORIAL MOUNTAINS						
Irian Jaya, Papouasia – New Guinea (4 300 m)	Limestone	4,000	Erratic pedestals	9,500	32.00	Peterson 1982
HYPERHUMID SUBARCTIC						
Diego de Almagro, Chilean Patagonia (51°S)	Marbles Limestones	7,330	Protruding dykes, erratic pedestals	c. 10,000	60.00	Maire *et al.* 1999
Madre de Dios, Chilean Patagonia (50°S)	Marbles Limestones	7,330	Protruding dykes, erratic pedestals	c. 10,000	100.00	Hobléa *et al.* 2001
All sites (mean)	—	7,330	—	—	80.00	—

always possible to discriminate between mechanical and chemical weathering. For instance, where the marbles of the U.S. cemeteries are concerned, Meierding (1981) has revealed the prominent role of 'chemically-induced granular disintegration'. Indeed, weathering often operates as a combination and/or suite of processes, both physical and chemical (and often also biological).

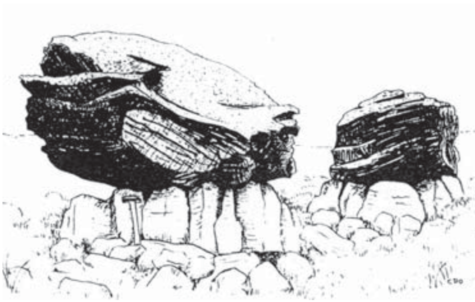

Figure 4. Erratics of Silurian sandstone on Carboniferous limestone pedestals. Norber Grags, Craven, England. Sketch by C.D. Ollier (*source*: Jennings 1987, Fig. 28 p. 85).
Note the jointing of the bedrock, which accounts for the prevalence of mechanical weathering over chemical weathering.

CONCLUSION

Whatever the limitations of the method, the use of the height of pedestals is of special interest when it comes to the assessment of rates of Holocene surface lowering, and should be extended to additional study sites in various climatic contexts. Pedestal rocks prove to be an almost ubiquitous geomorphic feature, whose use can provide for comparisons of datasets collected on limestone pavements and terraces from glacial environments to subtropical islands. More numerous, systematic and detailed studies based on this method should help to improve the assessment of the relative importance of the environmental and geological controls on weathering rates. Though environmental controls are fashionable and addressed more frequently, the role of rock properties should not be overlooked. The divorce between climatic and structural geomorphology is far behind us. The time for a more balanced and integrative view of erosional landform evolution has now come.

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PERMAFROST DEGRADATION AND SLOPE INSTABILITY IN THE ITALIAN ALPS

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Abstract: The occurrence of permafrost in the Italian Alps is an important predisposing factor for landslides. Debris flows are frequent on slopes overlying permafrost and are likely the result of active-layer supersaturation. The role of permafrost degradation in triggering landslides from hard rock slopes as a response to climatic warming has been pointed out only recently. Geotechnical investigations demonstrate that ice-filled fractured rock slopes at temperatures between –2° and 0° C are less stable than when in an unfrozen state. The Val Pola rockslide, whose accumulation mass included ice-cemented blocks, could be explained therefore as the result of strength reduction along pre-existing potential sliding surfaces, filled with warming ground-ice.

Key words: permafrost degradation, landslides, global warming, Italian Alps.

INTRODUCTION

The stability of permafrost terrain is sensitive to thermal disturbance, expecially where the ice content is relatively high. The effects of thermal changes operate on different time scales and at different depths, and include: (1) thickening of the active layer with thaw settlement in supersaturated materials (*the immediate response*), (2) disturbance of the temperature distribution at depth (*the intermediate response*), and (3) basal melting of permafrost ice with thaw settlement of supersaturated materials (*the final response*). The first effect, which commonly occurs seasonally, results in high pore-water pressures (Morgenstern and Nixon 1971; Harris *et al.* 2001) and is associated with mass movements such as skin flows, debris flows or detachment failures (e.g. Lewkowicz 1988). Thermal changes in permafrost are responsible for modifications of the unfrozen water within the ground over the temperature range 0 to –3°C (Williams and Smith 1989); it also reduces ice strength as temperatures approach 0°C (Patterson 1994). As a consequence, the ice/rock "adherence" decreases, thereby reducing the total strength of the rock mass. At the same time, the increasing amounts of meltwater can induce high pore-water pressure (Davies *et al.* 2001). The melting of ground ice, which may occur as a delayed response to thermal warming, depends on the original surface temperature, the distribution of ground ice, the thermal properties of the rock mass, and the intensity and duration of variations in ground surface temperature (Lachenbruch 1986). These conditions can generate potential sliding surfaces, especially in supersaturated deposits.

Where the rock mass is characterised by geometricaly-favourable discontinuities, the conditions of filling ground ice drive the possible triggering of landslides. Therefore, either melting of ice and the consequent built up of pore-water pressure or changes in the strength of ice are the main triggering factors behind instability below the permafrost table.

As a consequence of the recent climate warming (Hulme *et al.* 1999), some factors controlling permafrost thickness and persistence, such as air temperature, precipitation, and snow cover distribution, are predicted to change. This may induce widespread permafrost degradation (Haeberli and Beniston 1998), with negative effects on slope stability. These effects may be extremely dangerous in the mid-latitude high mountain valleys, where human settlements, railways and roads are often present. Despite these scenarios, awareness of the role of permafrost degradation as a potential geological hazard is still relatively limited.

This short paper stresses the role of permafrost degradation in inducing slope instability and mass movements in the Italian Alps, with special reference to debris flows.

PERMAFROST DISTRIBUTION

The distribution of permafrost requires indirect investigation methods, such as geophysical soundings (e.g. Guglielmin *et al.* 1994; Hauck *et al.* 2001), remote sensing techniques and automatic modelling (Hoelzle 1992; Keller 1992; Antoninetti *et al.* 1993). On the regional scale, some sectors of the Alps have already been mapped using automatic modelling (Keller *et al.* 1998; Guglielmin and Siletto 2000). For the Italian Alps, additional information is provided by rock glacier inventories (Guglielmin and Smiraglia 1997). Equally, on the medium scale (1: 10 000—1: 20 000), the knowledge of permafrost distribution is poor, since only a few sectors of the Central Italian Alps have been investigated sufficiently (Guglielmin 2007; Guglielmin *et al.* 2003). On the local scale, there are some sites in the Central and Western Italian Alps at which permafrost thickness and thermal conditions are relatively well known (Guglielmin *et al.* 2001;

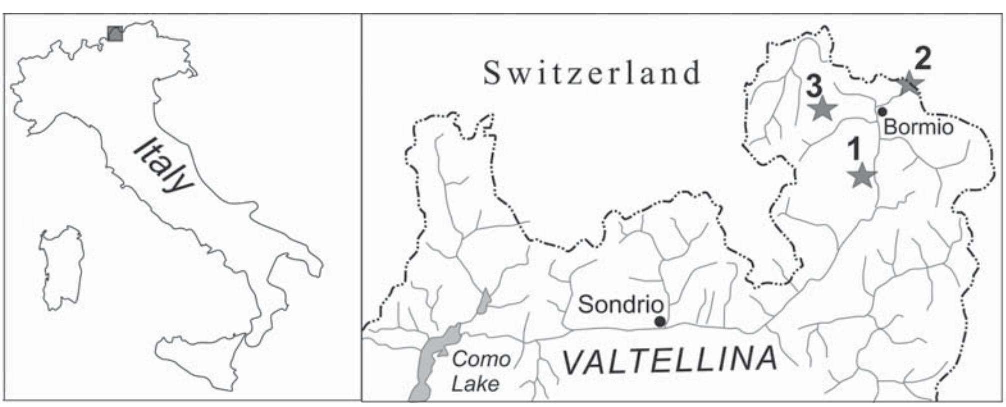

Figure 1. Permafrost distribution in some sectors of the Italian Alps and locations of the areas cited in the text: 1 Val Pola Landslide; 2 Stelvio Pass, 3 Foscagno area.

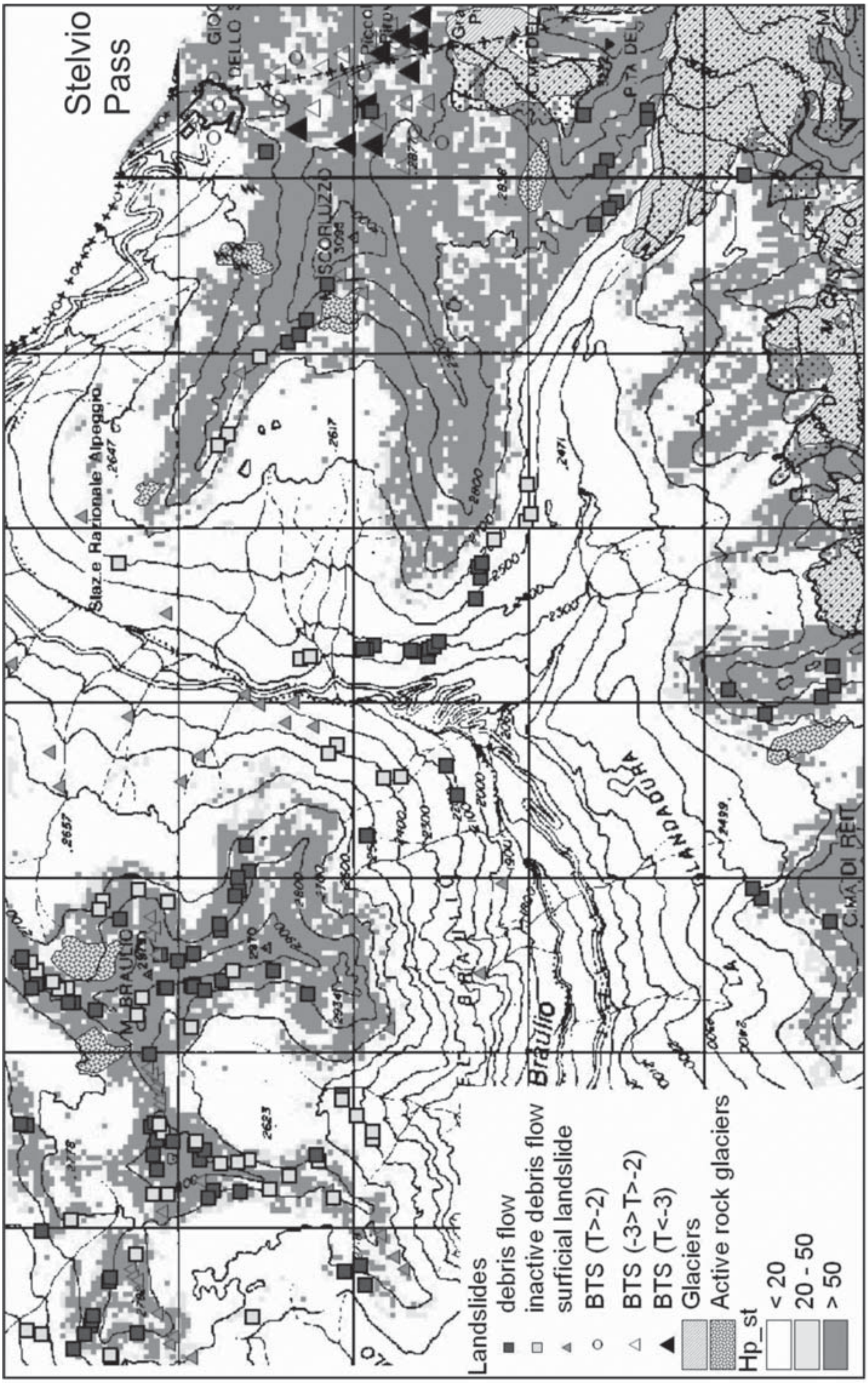

Figure 2. Permafrost map and landslides in the Stelvio Area (Silletto and Guglielmin, unpublished data).

Guglielmin 2004; Cannone *et al.* 2003; Ribolini and Fabre 2006; Ribolini *et al.* 2007).

PERMAFROST OCCURRENCE AND LANDSLIDE DISTRIBUTION

Zimmermann and Haeberli (1989) reported that more than 60% of the debris flows triggered during the exceptionally rainy summer of 1987 occurred in the periglacial belt of the Swiss Alps. More recently, catastrophic rock avalanches occurred in 1988 and 1991 on high mountain slopes with permafrost conditions (Schindler *et al.* 1993; Haeberli 1992). Landslides triggered within the permafrost zone are also reported from the Western Italian Alps (e.g. Brenva Glacier, Dutto and Mortara 1991). For this part of the Alps, an empirical model of permafrost distribution, based on the rule of thumb (Keller 1992), has been compared with a GIS landslide inventory. This comparison indicated that 21 landslides occurred above 2420 m a.s.l., that is the lowest altitude predicted for Alpine permafrost occurence. Moreover, Dramis *et al.* (1995) stressed the possible role of permafrost degradation in triggering the major catastrophic event of the last 20 years in Italy: the Val Pola landslide.

In this paper, we present three cases that support this interpretation. They are all located in Upper Valtellina, in the Central Italian Alps (Fig. 1).

DEBRIS FLOWS

A GIS inventory of landslides from the Stelvio Pass area, some kilometres north of Val Pola, has been compared with a permafrost distribution map of the same area (Siletto and Guglielmin, unpublished data). This reveals that more than 60% of events occurred within permafrost areas (Fig. 2). Where considerations are limited to recent/active debris flows, the percentage increases to more than 70%.

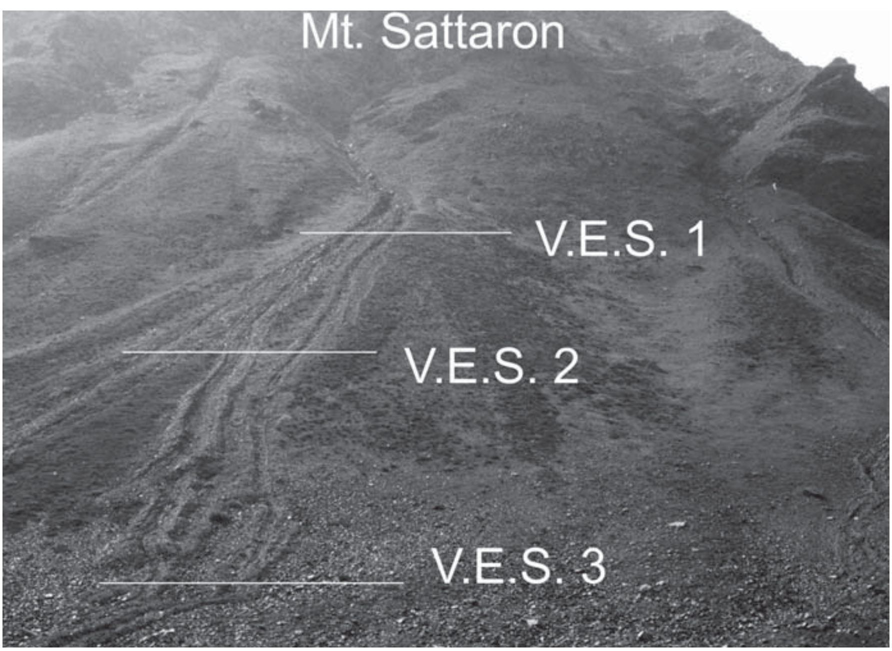

Figure 3. A view of the July 1987 debris flows on the north-eastern slope of Mt. Sattaron. The white lines indicate the locations of vertical electrical soundings shown in Fig. 4. (by Guglielmin M.).

The Foscagno area shows approximately the same results. In this area, along the north-eastern slope of Mt. Sattaron, four debris flows occurred in July 1987 (Fig. 3), starting between 2300 and 2400 m a.s.l., within the permafrost zone modelled by Guglielmin *et al.* (2003). Some geophysical soundings (vertical electrical soundings, Fig. 4), carried out in summer 2000, confirmed the occurence of "warm permafrost" in line with the definition of Haeberli and Vonder Mühll (1996). At almost the same altitude, but within the Foscagno Rock Glacier, a borehole was drilled in 1998 within the framework of the European PACE Project (Harris *et al.* 2003). The thermal profile recorded on August 29, 2002 (Fig. 5) showed a characteristic isothermal trend that was very close to 0 °C below the active layer (2.6 m).

ROCK SLIDES/AVALANCHES AND ROCK FALLS

The Val Pola landslide, northern Valtellina (Fig. 6), occurred on 28 July 1987. Its location was on the eastern slope of Zandila Peak, that consisted of densely-fractured and folded gneiss intruded by gabbro and diorite and overlain by thin glacial and colluvial deposits (Costa 1991). Some days earlier

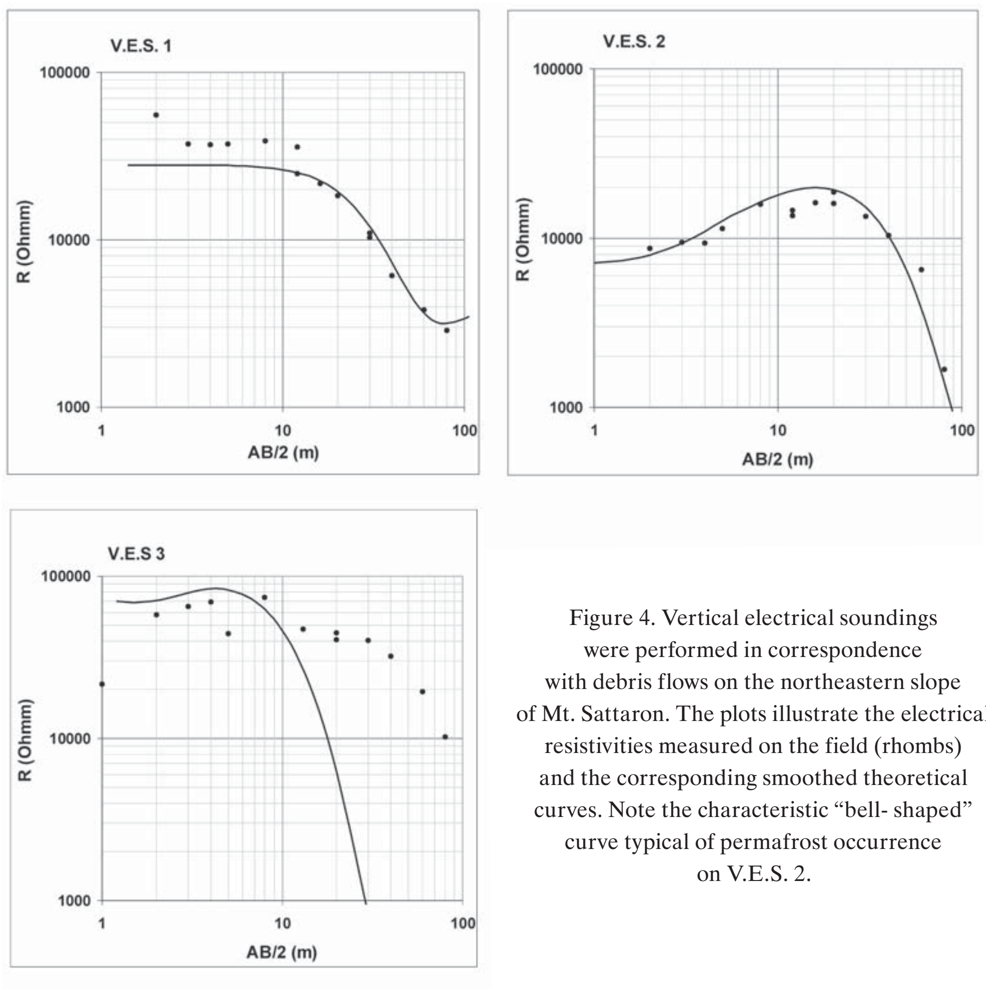

Figure 4. Vertical electrical soundings were performed in correspondence with debris flows on the northeastern slope of Mt. Sattaron. The plots illustrate the electrical resistivities measured on the field (rhombs) and the corresponding smoothed theoretical curves. Note the characteristic "bell- shaped" curve typical of permafrost occurrence on V.E.S. 2.

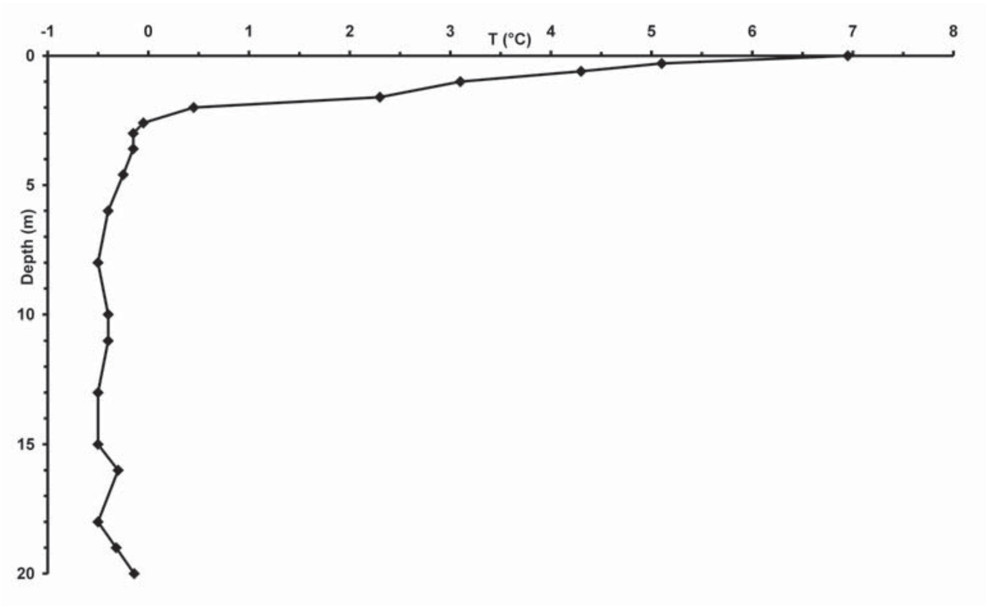

Figure 5. Thermal profile recorded on 29 August 2002 within the borehole drilled in the frontal part of the Foscagno Rock Glacier (2500 m a.s.l.).

(18–19 July) debris flows had also occurred, damming the Adda River. The landslide started at 2200 m a.s.l., on a slope affected by deep-seated gravitational deformation, located at the intersection of two major joint sets dipping 45° and 80° to the valley floor over an average hillside slope of 32°. It generated a rock avalanche of 33 million m^3 that fell to the valley floor (ca. 1000 m a.s.l.) and climbed to more than 300 m on the opposite valley side.

According to Costa (1991) and Crosta *et al.* (2004) the Val Pola landslide was essentially triggered by intense rainfall which induced high pore-water pressures in the fractured rock mass. However, the occurrence of ice-cemented blocks (from 0.5 to 2 m^3 in volume) within the landslide scree suggests a possible connection with permafrost degradation in the landslide detachment area (Dramis *et al.* 1995). Moreover, permafrost modelling by Guglielmin and Siletto (2000) indicates that the Val Pola slopes above 2200 m a.s.l. are *probable* and *possible* zones of permafrost occurrence. In addition, geoelectrical soundings confirm the presence of ground ice above 2300–2400 m a.s.l. on the western side of Zandila Peak (Dramis *et al.* 1995), as well as in the Zandila Cirque, just 1 km north of the landslide (Guglielmin, unpublished data). In this latter place, a 31.5 m deep borehole, drilled at 2490 m a.s.l. on 29 July 2004 found permafrost down to 23.5 m below an active layer of 2.2 m. The temperature was isothermal and very close to 0°C (Fig. 7).

DISCUSSION

In permafrost regions, the role of permafrost degradation in triggering landslides is well known, as regards the occurrence of skin flows and debris flows within the active layer during the early thawing season. Pore water pressure induced by thaw consolidation in fine-grained sediments is one of the main factors underpinning slope instability (e.g. Harris *et al.* 2001).

In coarse-grained sediments, such as the scree or till mantles that overly the Upper Valtellina slopes, the role of permafrost deg-

Figure 6. View of the Val Pola landslide The black arrow shows the location of the borehole of Zandila Cirque (by Guglielmin).

radation in triggering landslides is less clear. Nevertheless, permafrost occurrence seems linked to debris flow activity, as shown in Fig. 2. The results of a multivariate analysis that compares topographic factors (such as slope angle, aspect and altitude), with outcropping lithology, and permafrost occurrence, indicate that recent/active debris flows are best correlated with permafrost occurence. It is significant that most debris flows occurred in the early summer (June-July) period, when thawing was just beginning, and that this generally corresponds with episodes of heavy rainstorms. It is likely, therefore, that permafrost works as an impervious barrier reducing the water retention capacity of soil, increasing pore-water pressures and decreasing the shear strength of soil material.

In two sectors of the Upper Valtellina, one near Stelvio Pass and one near Foscagno Pass, the incidence of coarse-grained debris flows correlates well with permafrost occurrence. This is explained by the rapid oversaturation of an active layer above an impervious permafrost table at a time of stormwater occurrence.

It is more difficult to evaluate the effects of permafrost degradation in triggering mass movements (such as rock slides and rock falls) in massive bedrock slopes. As geotechnical investigations have made clear (Davies *et al.* 2001), ice-filled fractured bedrock at temperatures between –2°C and 0°C is potentially more unstable than similar rock in an unfrozen state, because the presence of liquid water within the ground ice causes a reduction in rock mass strength.

Deep-reaching rock failures, such as the Val Pola landslide, could therefore be explained, at least in part, as the result of the strength reduction along ice-filled potential sliding/detachment surfaces (e.g. joints, faults or gravity-induced deep fractures), as a consequence of permafrost warming. The deep location of the main sliding surfaces, more than 20 m below ground level, excludes any possible direct seasonal effect, pointing to the role of deep penetration of the last century's increase in surface temperature, which, according to Harris *et al.* (2003), has been responsible for a 0.5–0.8°C warming of the top layers of permafrost in the European Alps.

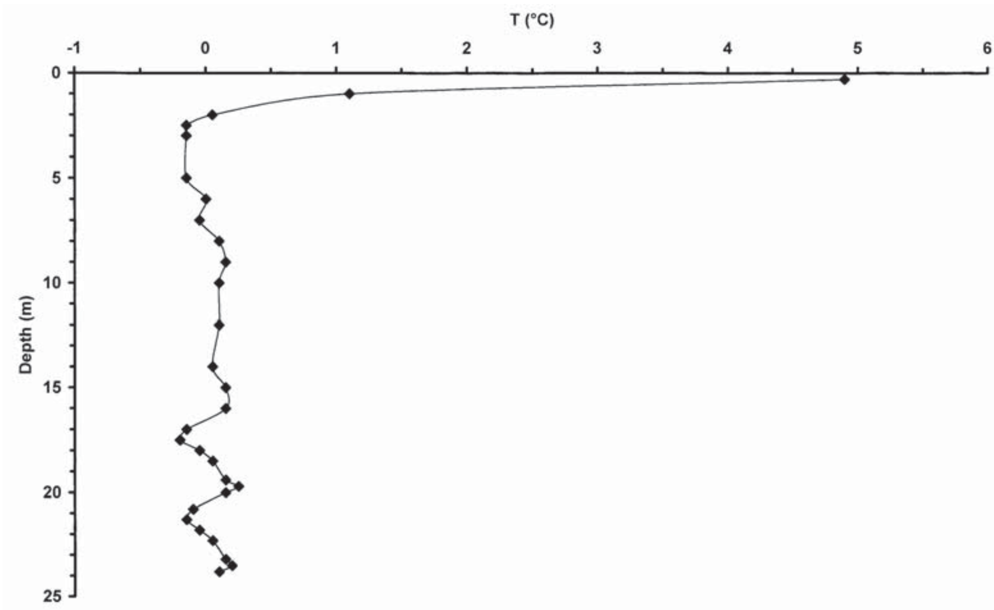

Figure 7. Thermal profile recorded within the borehole drilled in the Zandila Cirque on 29 July 2004.

The seasonal response causing the thickening of the active layer is delayed by only a few months, and is possibly responsible for an increasing incidence of rock-falls and topplings, as were observed in the hot summer of 2003 (e.g. Gruber *et al.* 2004).

CONCLUSION

There is an urgent need for further investigations on the role of permafrost degradation as a triggering factor for landslides, together with studies documenting permafrost distribution and its thermal state. The described case studies confirm that the stability of slopes may be markedly influenced by permafrost degradation. Despite the lack of *in situ* monitoring and laboratory simulation of coarse-grained soils in permafrost condition, it is reasonable to hypothesise that the triggering of debris flows can be explained by the rapid oversaturation that occurs in the active layer due to the presence of an impervious permafrost table. Taking into account the fact that the present-day permafrost temperature in the Italian Alps is generally higher than –2°C—as an effect of climate warming, and considering that future scenarios of climate warming indicate a progressive increase in air temperature, it appears clear that an increasing number of landslides are to be expected in the the near future.

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MOUNTAIN-TOP DETRITUS AND PATTERNED GROUND IN THE GASPÉSIE MOUNTAINS, QUÉBEC, CANADA

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Abstract: Mountain-top detritus characterizes the two high summits of the Gaspésie Mountains, eastern Canada. It is suggested that these angular rock-rubble accumulations developed from the disintegration of coarse-grained igneous bedrock exposed to thermal stress and ice segregation during prolonged episodes of permafrost formation in the cold periods of the Pleistocene. Frost wedging and frost heaving ('jacking') were the primary mechanisms. Today, climatic conditions on the summits permit only thin and marginal permafrost bodies. Stone nets and stripes are developed where a residual bedrock-derived debris mantle is present. They reflect frost-induced movements within the active layer. The latest of these movements probably occurred during the cold period following the LGM and persisted into the mid-Holocene. The transition from nets to stripes relates to slope angle.

Key words: mechanical weathering, permafrost, blockfields, patterned ground, Gaspésie Mountains, Québec, Canada.

INTRODUCTION

Blockfields ('felsenmeer'), and associated stone nets and stone stripes, are well known cold-climate phenomena (e.g. see Rea, 2007). These angular rock-rubble deposits are more generally referred to as 'mountain-top detritus' (Ballantyne and Harris 1993; Ballantyne 1998), and this term is therefore used here. Most is regarded as autochthonous (i.e. formed *in-situ*). While some is forming actively, as in Arctic Canada, the majority is thought to be of either Pleistocene or pre-Pleistocene age. On the basis of studies in the Highlands of Scotland, Ballantyne (1998) identifies three types of mountain-top detritus: clast-supported, matrix-supported, and matrix-supported with evidence for frost sorting.

Angular rock-rubble accumulations occur in many upland regions of the world. They have been described previously from Europe (e.g. Rea *et al.* 1996; Ballantyne 1998), North America (e.g. Dyke 1984; Clark 1992), and southern Russia and central Asia (e.g. Romanovskii *et al.* 1989). At mid-latitudes they appear especially well developed in areas that were either ice-free or marginal to the maximum extent of Pleistocene ice. One such area comprises the Carpathian Mountains and associated uplands of central Europe and southern Poland (e.g., Klatka 1962; Żurawek and Migoń 1999; Traczyk and Żurawek 1999). Hence, one of the early descriptions of these rock-rubble accumulations was from the Gorgany Range, by the Polish geologist Walery Łoziński (1912), who referred to them as 'periglacial facies' (Table 1).

Table 1. Some mid-latitude localities at which mountain-top detritus is attributed to mechanical weathering during Pleistocene cold-climate conditions.

A. HISTORIC:

1. '*Stone runs*', Falkland Islands, latitude 52°S, (Andersson 1906):
   Lithology: Quartzite
2. '*Periglacial facies*', Gorgany Range, Poland/Ukraine, latitude 49°N (Łoziński 1912):
   Lithology: Sandstone

B. MORE RECENT DESCRIPTIONS:

1. '*Blockfields*'; '*Felsenmeer*'; Carpathian Mountains, Europe; latitude 49–51°N:
   (Klatka 1962; Żurawek 1999)
   Quartzite—Łysogóry Mountains, Carpathian Foreland, Poland Quartzite—Hruby Jesenik Mountains, Czech Republic
   Basalt—Hessian Highlands, Germany Granite—Karkonosze Mountains, Poland
   Gabbro—Ślęża Massif, Sudeten Foreland, Poland
2. '*Boulderfields*'; Appalachian Mountains, Eastern USA; latitude 39–41°N:
   (Clark and Ciolkosz 1988)
   Orthoquartzites—Appalachian Plateau and Ridge and Valley
   Metaquartzites and greenstone meta-basalts—Northern Blue Ridge
3. '*Kurums*'; Southern Yakutia and northern Transbaikalia, Russia; latitude ~55°N:
   (Romanovskii *et al.* 1989)
   Proterozoic and Archaean metamorphic rocks, quartzite-like sandstones
4. '*Mountain-top detritus*'; British Isles; latitude 58°N:
   (Ballantyne 1998; Ballantyne and Harris 1993)
   Quartzite, siliceous schist and mica-schist, granulite, sandstone—NW Scotland
   Gneiss—Outer Hebrides
4. '*Mountain-top detritus*', Gaspésie Mountains, Québec; latitude 49°N:
   (French and Bjornson this paper)
   Syenite, granodiorite and monzogranite—Mont Jacques-Cartier
   Serpentized peridotite and pyroxene—Mont Albert

This paper describes the mountain-top detritus and associated polygons and stripes that exist on the higher summits of the Gaspésie Mountains of eastern Québec, Canada (49°N, 66°W). The summits under discussion include Mont Jacques-Cartier, the second highest elevation in eastern Canada (1270 m a.s.l.), and Mont Albert, a plateau-like surface rising to 1050 m a.s.l. (Fig. 1). Both summits lie within the boundary of Le Parc de la Gaspésie, a Québec Provincial Park.

REGIONAL CONTEXT

Mont Jacques-Cartier is a Devonian-age igneous intrusion, or batholith, consisting of acidic hybrid rocks such as syenite, granodiorite and monzogranite. These resistant rocks have been exposed by removal of enclosing Cambro-Ordovician metasediments. Mont Albert is an older, Cambrian-age, igneous intrusion consisting of serpentinized peridotite and pyroxene surrounded by an amphibolite belt. The rest of the Gaspésie Peninsula comprises folded meta-sediments of Cambrian and Palaeozoic age, in the northern part of the Appalachian (Hercynian) mountain chain.

The summits of both studied mountains rise above the treeline, and form broad and gently-sloping tundra surfaces that cover areas of approximately 4 km² and 12 km² respectively. The treeline in the Gaspésie Mountains is a climatically-controlled krummholz transition that separates subalpine white-spruce-dominated forest from alpine tundra. It occurs at an elevation of approximately 1050–1080 m a.s.l. on Mont Jacques-Cartier, but is slightly lower—due to the serpentine nature of the rock—on Mont Albert. Both summits exhibit extensive rock-rubble surfaces (Fig. 2), these being everywhere colonized by lichens. In general, the summit terrain of both mountains is indicative of limited active landscape modification.

Many of the upland surfaces of the Gaspésie Peninsula are covered with a residual mantle of locally-derived angular bedrock clasts contained within a matrix of fine silty-clay and sand (Veillette and Cloutier 1993). This mantle is ~1.0 to 1.5 m thick. Where glacial material is incorporated, this rests near the surface. The general absence from the Gaspésie Peninsula of macro-scale glacial lineaments and ice-sculpted terrain suggest that the Late Wisconsinan (Laurentide) ice was cold-based and largely non-erosive in the area. However, glacial striae and erratic dispersal trains within several U-shaped valleys suggest radial movement outwards from the higher elevations. Regional deglaciation is thought to have occurred approximately 13–11 ka BP (Richard *et al.* 1997), though active glacier-ice bodies may have persisted within certain valleys until at least 10 ka BP.

The patterned ground on the Mont Jacques-Cartier summit was first described by Gaumond and Hamelin (1960). Subsequent investigations indicate that permafrost on Mont Jacques-Cartier is approximately 45–60 m thick, the depth of zero annual amplitude is ~12 m, the temperature at that depth is ~-1°C and the bedrock active layer is ~5.75 m thick (Gray and Brown 1979; Gray *et al.* 1987). There is no information on ground ice conditions. Small solifluction lobes occur in the alpine tundra zone; C-14

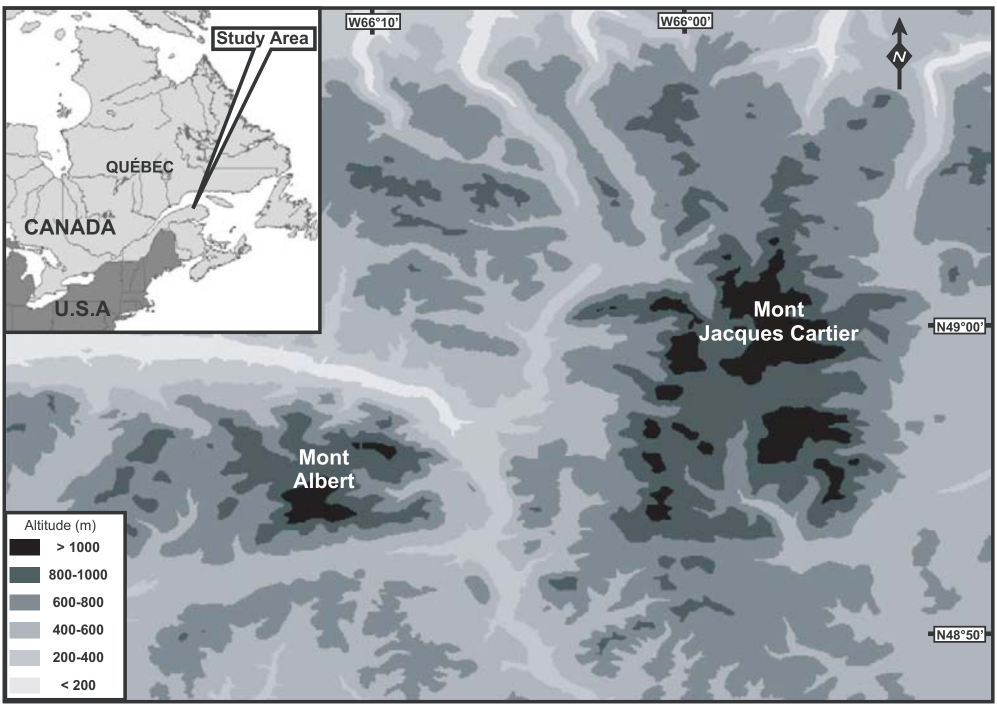

Figure 1. Location of the study area, Gaspésie Peninsula, Québec, eastern Canada.

Figures 2. The mountain-top detritus of the Gaspésie Mountains:
(A) On Mont Jacques-Cartier, it is developed upon Devonian-age syenite and granodiorite;
(B) On Mont Albert, it is developed upon Cambrian-age serpentized peridotite;
(C) A typical bedrock outcrop of serpentized peridotite showing weathered joints and fractures on Mont Albert;
(D) A typical frost-jacked block of lichen-encrusted granodiorite on Mont Jacques-Cartier.

dating indicates these are related to the cool period following the mid-Holocene hypsithermal (Payette and Boudreau, 1984).

MOUNTAIN-TOP DETRITUS

DESCRIPTION

Large areas of the Mont Jacques-Cartier and Mont Albert summits have accumulations of sub-angular bedrock blocks (Figs. 2 A, B). At the surface, the blocks are typically between 0.3 and 1.3 m in dimensions, often with rounded edges and, on Mont Jacques-Cartier, with lichens on exposed surfaces. Adopting Ballantyne's classification, the blockfields of the Gaspésie summits are predominantly of Type 1 (clast-supported, openwork surface) but, on sloping terrain and where the weathered mantle is present, they are best described as Type 3 (matrix-supported with evidence of frost sorting). On both summits, small (~1–2 m high) bedrock outcrops occur; these reflect the upslope edges of 'cryoplanation' terraces, structurally-controlled bedrock surfaces (Fig. 2C), or diabase dykes (see Gray *et al.* 1987). In places, frost-heaved bedrock blocks occur (Fig. 2D).

Large exposures of *in-situ* bedrock on the summits of both Mont Jacques-Cartier and Mont Albert are limited to the small diabase dykes and bedrock surfaces mentioned above. For environmental reasons, few excavations have been permitted by Park authorities in recent years. Those that have been allowed on the Jacques-Cartier summit showed a 1–3 m thick weathered mantle that grades into unweathered syenite bedrock (Gray *et al.* 1987). In the only deep excavation, the surficial materials at depth were composed of alternating beds of oxidized and non-oxidized silty sand, containing increasingly weathered bedrock blocks. Our own excavations, undertaken in September 2000, 2001 and 2002 to depths of approximately 1.5 m, confirm that blocks at the surface grade quickly at depth into *in-situ* fractured bedrock. There is no downward-fining of the coarse debris. X-ray diffraction analyses of the clay and silt fraction present 0.75 m down also confirm an advanced degree of chemical weathering with a dominance of gibbsite and kaolin minerals.

ORIGIN

It is generally accepted that most examples of mountain-top detritus (blockfields) at mid-latitudes are the result of Pleistocene frost action (Ballantyne and Harris 1993). This is inferred from the tendency for coarse debris to decrease in size with depth, suggesting vertical frost sorting. The frost-action explanation, first invoked by Andersson (1906) in his discussion of the quartzite 'stone runs' of the Falkland Islands, has been widely invoked to explain the mechanical weathering of resistant bedrock in many parts of the world. Andersson envisaged production of coarse debris by the expansion of water as it underwent freezing in joints and cracks (i.e. frost wedging or hydro-fracturing) followed by removal of weathered material by solifluction. Ultimately, stabilization of the angular rock-rubble was achieved as fines were progressively washed out by rain and snowmelt.

This elegant explanation is not without criticism. For example, on Mont Jacques-Cartier, the absence of near-surface fines may merely reflect eluviation; in the spring and early summer months in the Gaspésie Mountains, and also following rain events, water can be heard moving through the blocks in the near-surface. More general criticisms reflect increasing doubt as to the efficacy of frost-wedging in producing angular blocky rubble, the more so where relatively resistant low-porosity crystalline bedrock is concerned. This scepticism is largely attributed to the paucity of freeze-thaw cycles at depth within bedrock, as well as an increased appreciation of the role of ice segregation and frost heaving in the disintegration of rock (e.g. Murton *et al.* 2006; Murton 2007). For example, in the permafrost terrain of Arctic Canada, frost heaving leads not only to the shattering of sedimentary rocks such as carbonates and sandstones, but also to the rupturing and dislocation of massive quartzitic, granitic and gneissic bedrock (e.g. L.D. Dyke 1984; A.S. Dyke 1978).

The mountain-top detritus on the Gaspésie summits is especially problematic because (1) the Gaspésie, in common with most of eastern Canada, experienced heavy Late Wisconsinan glaciation, and (2) regional mapping by Veillette and Cloutier (1993) indicates that blockfields are only preserved on the summits associated with the igneous intrusions of Mont Jacques-Cartier and Mont Albert. In the Gaspésie, the lack of glacially-sculpted lineaments (Veillette and Cloutier 1993) suggests the Late-Wisconsinan (Laurentide) ice sheet was non-erosive and cold-based. Although deglaciation occurred between 13–11 ka BP (Richard *et al.* 1997), active rock glaciers and glacial ice continued to exist in some of the deeper valleys as late as 9 ka BP (Hetu *et al.* 2003).

A significant characteristic of mountain-top detritus is that, in most instances, it appears associated with predominantly coarse-grained crystalline bedrock. The latter include granite, quartzite, gneiss, mica-schist, gabbro, and sandstone (see Table 1). Thus, the mountain-top detritus of the Gaspésie Mountains is not unusual. A second characteristic is that the majority of these angular rock-rubble surfaces appear inactive, or relict. This appears to be the case for the Gaspésie Mountains, in which the short period of post-glacial time available for the mechanical disintegration of these highly resistant batholiths is a strong argument for their being relict in nature.

Unfortunately, no direct dating of the bedrock surfaces in the Gaspésie Mountains is yet available. However, recent cosmogenic dating from elsewhere in eastern North America is beginning to clarify the possible time(s) at which the Gaspésie mountain-top debris formed. For example, cosmogenic nuclide exposure ages for bedrock outcrops and tors in adjacent Labrador and Québec indicate these features existed prior to the Sangamonian interglacial (i.e.OIS-6 or OIS-8), and survived the cold period of the Late-Wisconsinan (OIS-2 or LGM) in Eastern Canada (Marquette *et al.*, 2004).

Relatively little is also known about the speed at which massive igneous bedrock fractures and disintegrates under cold-climate conditions. One of the few qualitative estimates for the rupturing and dislocation of granitic gneiss is from Baffin Island, where '…5000 to 10,000 years are required to produce minor disruption… of bedrock terrain, and 20,000 years or more are required to produce felsenmeer' (Dyke 1978). This conclusion mirrors a general consensus that mountain-top detritus takes many tens of thousands of years to form.

The most probable age for the angular rock-rubble on the Gaspésie summits is that it developed during either the Late Tertiary or during the prolonged cold periods of the Early and Middle Pleistocene, when the Gaspésie Peninsula was either ice-free or had higher elevations existing as nunataks above the continental ice sheet. If either of these scenarios were the case, permafrost would have penetrated deep into the exposed and deeply weathered bedrock. To judge from the mechanical disintegration and frost jacking of bedrock in the Canadian Arctic today, ground temperatures of -5 to -15°C would have been typical. If permafrost were involved, attention would also have to focus upon our understanding of the response of coarse-grained crystalline rocks when subject to sub-zero temperatures. For example, silica-rich rocks experience thermal stresses under cryogenic conditions that enhance mechanical disintegration. The coefficients of linear and volumetric thermal expansion and contraction are highest in rocks and minerals that have a low energy of the crystalline lattice. In other words, the greater the SiO_2 content of a rock, the larger these coefficients become. Experimental studies have also demonstrated that the coefficient of thermal expansion of acidic rock is several times greater than that of ultrabasic rock (Yershov 1990). These considerations raise the possibility that permafrost-induced stresses not only enlarge existing joints and fracture planes in igneous rocks but also initiate small cracks and fractures, thereby enhancing the efficacy of frost wedging and frost heaving ('jacking').

STONE NETS AND STONE STRIPES

In places, the Jacques-Cartier summit possesses a mantle of bedrock-derived weathered debris. This angular rock-rubble has been sorted into stone nets (polygons) and stripes (Fig. 3). The stripes are very similar to the quartzite 'stone runs' first described by Andersson (1906).

DESCRIPTION

The stone nets are located on flat to sloping terrain, usually less than 2–3° in angle, generally occurring at lower elevations than the blockfields and best developed on the broad gentle summit slopes. Typically, their maximum dimensions are 2–4 m. In most places, the pattern constitutes a net, rather than a polygon. The polygon centres, often vegetated, consist of a silty-sand matrix containing angular bedrock fragments and boulders. By contrast, the edges are marked by an openwork accumulation of larger boulders, many on edge, that sometimes form shallow linear depressions or, where polygons intersect, stone pits.

As slope angle progressively increases away from the Jacques-Cartier summit, the nets change to openwork stone stripes, typically ~1–1.5 m wide, separated from each other by vegetated soil stripes ~1.5–3.0 m wide (Fig. 3A). The stripes extend downslope, sometimes for several hundred metres, in approximately parallel lines. Not all are continuous, since some merge with adjacent

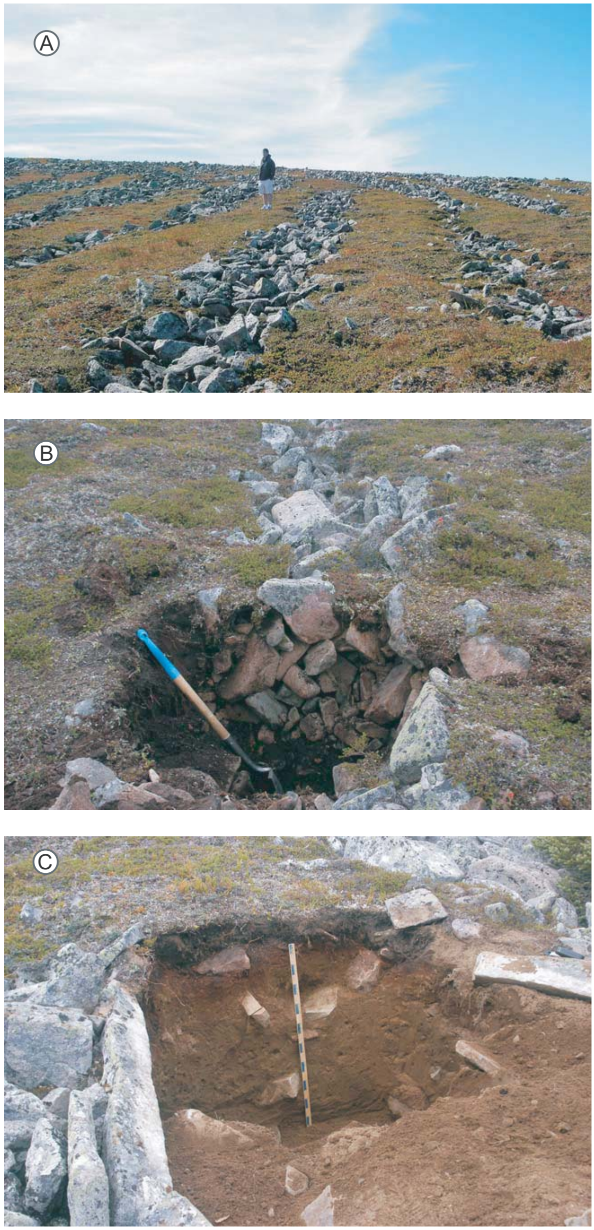

Figure 3. Patterned ground on the Mont Jacques-Cartier summit:
(A) Stone stripes on low-angled (3–7°) slopes are separated from each other by vegetated stripes of finer material;
(B) An exposure through a stone stripe reveals an openwork matrix of angular rock fragments that extend downwards to a depth of less than 1.5 m;
(C) The vegetated stripes are developed upon a bedrock-derived diamict that consists of angular bedrock clasts within a silty-sand matrix.

stripes, some disappear beneath the alpine tundra vegetation on the lower slopes, and others end without any apparent cause.

ORIGIN

Like the mountain-top-detritus described earlier, neither the polygons nor stripes appear active today. For example, the clasts that form the stone stripes are rounded or sub-angular, covered with lichen, and are stable. Their current geomorphic role appears to be that of concentrating surface run-off from the summit following spring snowmelt and summer rain events. The intervening stripes are vegetated, and although they show little sign of movement, no quantitative process studies are available to indicate current rates of soil movement. In all probability, mass wasting is ongoing, given the cold climate experienced on the summit and the presence of marginal permafrost today. However, as Payette and Boudreau (1984) concluded, the most recent episode of enhanced solifluction occurred during the mid-Holocene.

Sections were excavated in September each year between 2000 and 2005 across a stone stripe and the adjacent soil (vegetated) stripes on an east-facing 5-degree slope on Mont Jacques Cartier. Each excavation was at the same spot each year, but progressed 0.5 m upslope on each occasion in order to minimize the extent of terrain disturbance. The section was backfilled each year. The stone stripes (Fig. 3B) consist of sub-angular bedrock blocks at the surface, many of which are aligned towards the vertical, resting in an open matrix. At ~0.75 m depth, the stone stripe ceased to exist as it became increasingly incorporated within the silty sandy and blocky matrix making up adjacent vegetated stripes (Fig. 3C).

Almost certainly, the sorted nets and stripes are the result of differential frost heave within the heterogeneous soil matrix. They are best interpreted in terms of soil circulation within the active layer. Recent field, experimental and modelling studies indicate that ice lenses displace soil towards soil-rich domains, and stones towards stone-rich domains (Kessler and Werner 2003; Matsuoka *et al.* 2003). The stone accumulations are then squeezed and confined as the freezing domains expand, and angular clasts tend towards the vertical. According to Kessler and Werner (2003), polygonal networks form when stone domain squeezing and confinement dominate, and stripes form as slope gradient increases.

CONCLUSIONS

Our interpretation of the origins of the mountain-top detritus (blockfields) and patterned ground on the Gaspésie summits is illustrated schematically in Fig. 4. The mountain-top detritus is certainly relict and probably formed when the summits, previously subject to deep weathering in the Tertiary, experienced long periods of cold permafrost conditions during the Pleistocene. During these latter periods, thermal stresses within the bedrock enhanced the efficacy of ice segregation and bedrock heave. The stone nets and stripes formed in the active layer wherever a surficial weathered mantle of unconsolidated sediments was present. Differential frost-heave led to the outward movement of angular clasts, combined with a net inward movement of fines.

We are unsure as to the actual time at which the mountain-top detritus formed. To judge by conditions necessary for bedrock heave in crystalline rocks in Arctic Canada today, mean annual ground temperatures of between -5 and -15°C must be assumed. For this to occur, the summits must have been ice-free and exposed to low air temperatures for several tens of thousands of years. Therefore, a pre-Late-Wisconsinan age is implied. Also, in order for the shattered bedrock to be preserved *in situ*, the Late-Wisconsinan (Laurentide) ice that covered the Gaspésie Peninsula must have been non-erosive. The nets and stripes are of lesser antiquity; in all probability, the latest period for their formation was the cold period following deglaciation from the LGM. This continued until the mid- Holocene, when they became largely inactive.

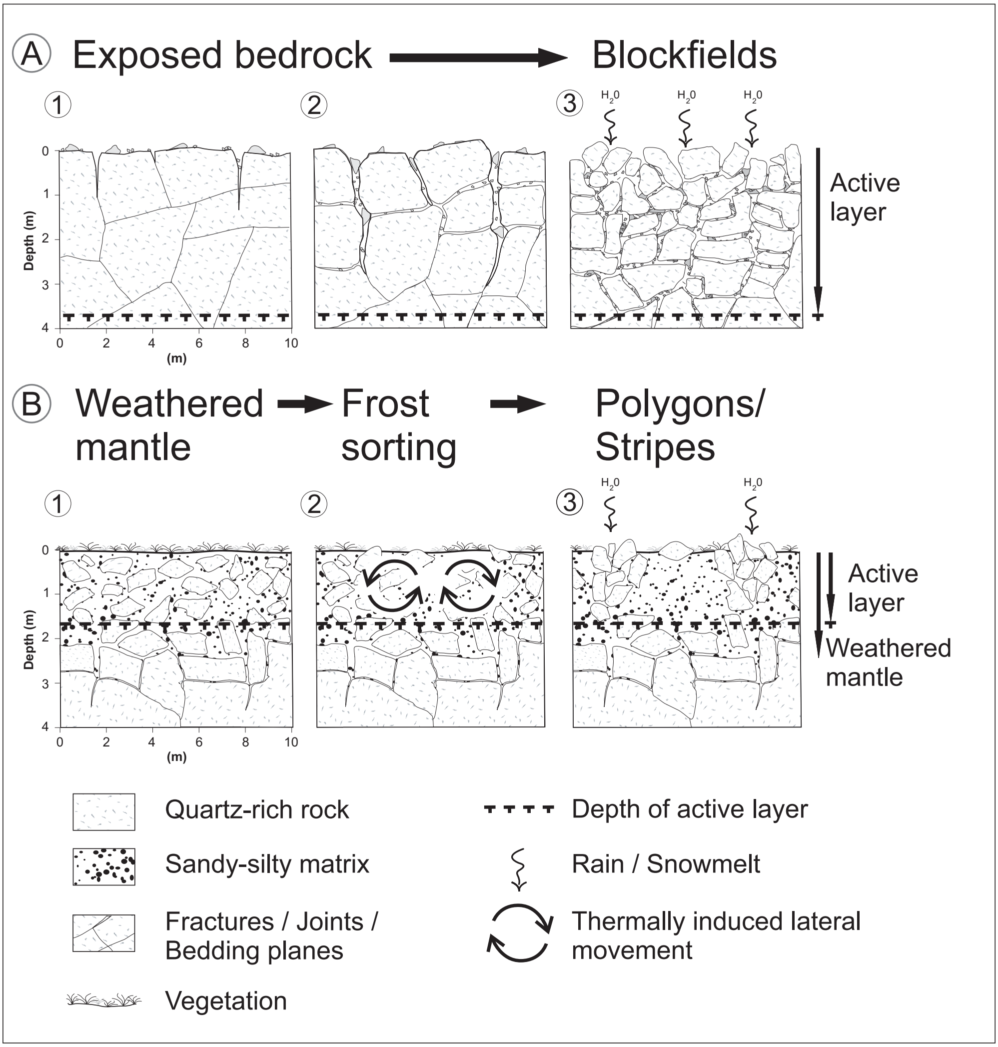

Figure 4. Schematic diagram showing suggested evolution of the mountain-top detritus and patterned ground on Mont Jacques-Cartier and Mont Albert, Gaspésie Mountains.

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This paper pays tribute to the seminal studies undertaken by Professor Adam Kotarba in understanding the mountain environments of Central Europe. HMF also wishes to thank Dr Roman Żurawek for

sharing his knowledge of the blockfields and related phenomena of southern Poland.

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THE PROBLEM OF FIRN-ICE PATCHES IN THE POLISH TATRAS AS AN INDICATOR OF CLIMATIC FLUCTUATIONS

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Abstract: This paper attempts to determine the relationship between multi-annual variability of air temperature, precipitation and wind velocity, and changes in the front limits of the perennial firn-ice forms (glacierets) developing under different topographic conditions. Problems with the palaeoclimatic interpretation of their internal structure are also discussed. The obtained results attest to the fact that fluctuations in firn-ice patches in the Tatras are probably most connected with the weather regimes in winter seasons. Similar changes of individual forms depend on their similarity in terms of type of snow accumulation and the altitude at which they are located.

Key words: glacierets, cryospheric indicators of climate change, Tatra Mts.

INTRODUCTION

The global warming that terminated the Little Ice Age in the 1850s is at present most probably being intensified by human impact due to the so-called greenhouse gases (IPCC 2007). The present trend towards increases in air temperature is clearly reflected in changes of the natural environment in glacial areas. The results of international monitoring of the cryosphere reveal the mass loss of the larger glaciers studied (IAHS/UNESCO 1993; 1998; 2003; 2007), and, in periglacial regions, a warming of permafrost is unmistakable (eg. Harris *et al.* 2003, Osterkamp 2007). In addition, high-mountain areas that are not glacierized at present may also show marked sensitivity to change because, above the timberline, the natural environment is mainly affected by processes associated with frost and snow. It seems that firn-ice patches, also called "perennial snow patches" or "glacierets" (IAHS/UNESCO 1970), are especially sensitive to climate fluctuations (Higuchi 1975; Watanabe 1988; Gams 1994; Jania 1997a; Grunewald *et al.* 2006; Hoffman *et al.* 2007).

References to the presence of multi-annual snow in the Tatras, referred to as "Montes Nives" in the late Middle Ages, have been published since the beginning of the 17th century (Szaflarski 1972). In 1839, Zejszner recognized some firn-ice bodies as glacierets (Olędzki 1965). Over eighty years later, Gadomski (1925) produced the first descriptions and measurements of these. In the next decades of the 20th century, work to document the firn-ice patches in the Polish Tatras was carried out more frequently (eg. Milata 1949; Wdowiak 1961; Olędzki 1965; Kłapa 1980; Wiśliński 1985;

1996; 2002; Rączkowska 1995; Jania 1997a; Gądek 2002). Data on annual limit changes of the Mięguszowiecki, Pod Bulą and Pod Cubryną glacierets have been monitored since 1980 by Wiśliński (2002), and published in successive volumes of *Fluctuations of Glaciers* (IAHS/UNESCO 1993; 1998; 2003). The results of investigations carried out by Wiśliński may be summarised in the following terms: a) the development cycles of the observed firn-ice patches are differentiated, b) changes in their dimensions are very much connected with the destruction of subglacial tunnels, c) these changes cannot be explained by changes in air temperature (Wiśliński 2002; Ciupak *et al.* 2005). However, on the basis of analysis of all the data from the period 1923–1994 (including the results of his own 6-year photogrammetric monitoring of the Mięguszowiecki and Medeny glacierets), Jania (1997a) was able to demonstrate a reduction in area and thickness that was recognised as a reflection of general climatic warming. However, in the Tatra Mountains at that time there were no changes in air temperature or precipitation that might be regarded as the effect of warming (Niedźwiedź 1996, 2005). On the other hand, it was determined the whole vertical profile of the Tatras was witnessing negative trends as regards snow cover stability and seasonal maximum thickness (Falarz 2001).

In the Western Carpathians, the climatic snow line (CSL), as the so-called theoretical lower limit of the annual preservation of snow cover on a horizontal and non-shaded surface (Jania 1997b), runs at an altitude where mean annual air temperature is -8°C (i.e. at the non-existent ~ 3400 m a.s.l.) (Hess 1965). Firn-ice patches in the Tatras occur 1000–1800 m below that line, and are nourished mainly by avalanches and blown snow. Their occurrence is therefore more connected with topographic conditions than climatic ones. Differentiation of the dimensions of these forms results from variability in the dimensions to snow/ice accumulation and ablation in places where they occur. The first of the mentioned mass-balance elements of the firn-ice patches is connected with snow precipitation and redistribution, and the second mainly with heat balance of their surfaces, which correlates well with totals for positive degree-days (PDD) (Gądek 2002; 2003). Variability to snow precipitation and PDD also "steer" the location of CSL. The question to be asked thus concerns the extent to which the local redistribution of snow (via avalanches and blowing), as conditioned by relief, can limit statistical relationships between regional climatic fluctuations and the changes in dimension characterising firn-ice patches? In that context, the role of rain and the concentrated flow of meteoric water along avalanche tracks is intriguing. It seems that good indicators of the mass-balance variability to small ice forms are changes in their length (Bahr *et al.* 1997) and the number and thickness of their annual layers (Wdowiak 1961; Jaworowski 1966; Gądek and Kotyrba 2003). This work therefore shows the results for the correlation noted for changes in glacieret fronts monitored by Wiśliński (op. cit) in the Rybi Potok Valley as set against snow and rain precipitation, air temperature and wind velocity in the alpine zone of the Tatras. Issues relating to the palaeoclimatic interpretation of the internal structure of firn-ice patches are discussed by reference to the results of the 4-year observation of the way in which annual layers develop in the Mięguszowiecki glacieret (the largest in the Polish Tatras), as well as their georadar documentation (Gądek and Kotyrba 2003).

STUDY AREA

The Rybi Potok Valley is situated in Poland's High Tatra Mountains (Fig. 1). The highest summit of the main ridge bordering this area from the S and SW reaches 2499 m a.s.l. (at Rysy), while the mean altitude along the ridge is 2300 m a.s.l. Pleistocene glacial and periglacial forms predominate in the relief. According to Wiśliński (1996), as of the early 1980s, about 60% of all firn-ice patches in the Polish Tatras were present here. In the period 1981–1985, the number grew from 27 to 58, while their total area increased

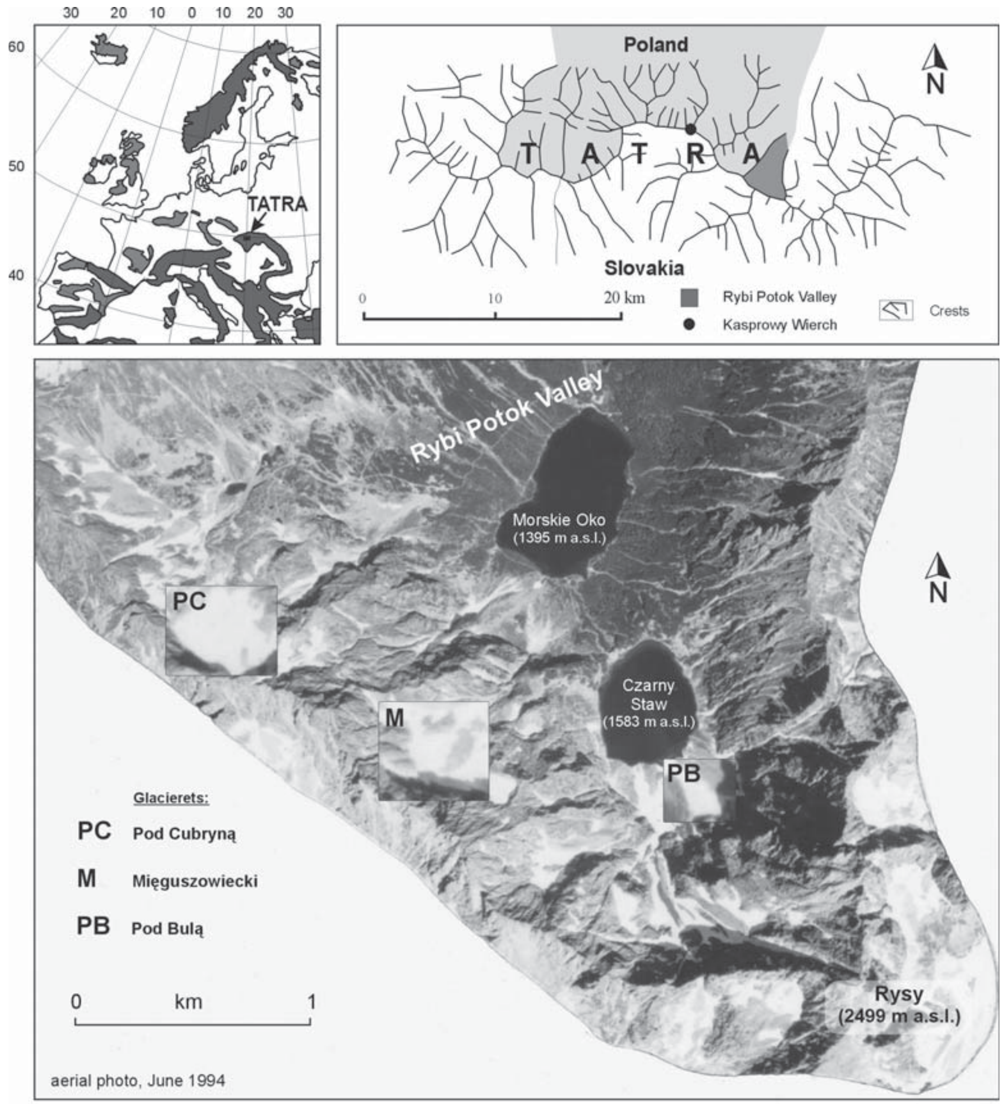

Figure 1. The location of the research area.

from 1.5 to 3.7 ha. The firn-ice patches were restricted to the altitudinal zone between 1520 and 2430 m a.s.l., usually on north-facing slopes. About 85% of these were restricted to the alpine zone (at 1800–2200 m a.s.l.), in shaded and concave topographical forms. The Rybi Potok Valley also has the longest monitored glacierets in the Tatra Mountains (Wiśliński 2002), i.e. the Mięguszowiecki (1978–2035 m a.s.l., facing N), the Pod Bulą (1670–1710 m a.s.l., facing NW) and the Pod Cubryną (2110–2140 m a.s.l., facing N). The Mięguszowiecki and Pod Bulą glacierets are mainly nourished by avalanche snow, while the existence of the Pod Cubryną glacieret is connected with snow blowing. The process of metamorphosis of snow into white ice takes from one to eight years (Wiśliński 1985) and occurs with the involvement of melting and meteoric water. Patches of firn-ice are usually composed of several to a dozen or so annual layers, though they disappear periodically (Kłapa, 1966; Wiśliński 2002). The Mięguszowiecki glacieret is a unique form, as it has probably existed throughout the subatlantic period (Jania 1997a). It is a very stable avalanche cone built mainly from firn and white ice. Its area usually covers 0.5 ha, while its maximum thickness reaches 22 m (Wdowiak 1961; Wiśliński 1985; Gądek 2002). Its annual layers move rotationally towards the front and melt gradually. As total mass exchange probably takes 100–150 years (Wdowiak 1961; Gądek 2002), the presently existing form is one exclusively connected with climatic conditions occurring in the Tatra Mountains after the Little Ice Age.

The only meteorological station located in the alpine zone is on Kasprowy Wierch, and this is also the most-elevated station in the Polish Tatras (at 1991 m a.s.l.). This summit is located on the main ridge near the eastern limit of the Western Tatras (Fig. 1).

DATA AND METHODS

METEOROLOLOGICAL DATA

The study made use of daily precipitation data (including kind of precipitation), air temperature and wind velocity recorded at the synoptic station on the top of Kasprowy Wierch (WMO 12 650) in the years 1980–2001. These data provided the basis for calculations of totals for snow precipitation, totals for rain precipitation, maximum daily totals for rain precipitation and PDD totals (totals of mean daily air temperature at a height of 2 m) in successive balance years—consistent with the dates of glacieret observations (showed below). In the calculations of snow precipitation totals and rain precipitation totals, mixed precipitation was included, assuming that the participation of rain and snow is 50% each. Moreover values for mean wind velocity in individual seasons with dry snow cover (snow blowing possibilities) were calculated, the latter usually developing from December through to end of March in the zone above the upper timberline (Kłapowa 1980).

FLUCTUATIONS OF FIRN-ICE PATCHES

Data on annual changes in the limits of firn-ice patches over the period 1980–1999 were taken from volumes 6–8 of *Fluctuations of Glaciers* (IAHS/UNESCO 1993; 1998; 2003). They were collected by Wiśliński for the Pod Bulą, Mięguszowiecki and Pod Cubryną glacierets. The measurements were made using measurement tape and a compass with clinometers at the end of the following ablation seasons: 3–18 September 1980, the last week of September or the first ten days of October 1981–2000. They consisted in the delimitation of the distance of the glacieret limits from benchmarks (Wiśliński 1996; 2002).

Changes affecting the Mięguszowiecki glacieret in the period 1998–2001 were also determined. To this end, use was made of stereopairs of ground photographs taken from the same sites (marked by survey pins) on: 11 September 1998, 24 August 1999, 16 September 2000 and 29 September 2001. The increase or decrease in glacieret' limits in the vertical and horizontal planes was interpreted as a result of positive or negative mass balances.

The dependence of glacieret fluctuations upon changes in totals for snow and rain precipitation and PDD in successive balance years, as well as upon mean wind velocity in the months December-March, was checked using the linear regression method and the preparation of tables for Pearson correlation coefficients and values for their statistical significance. Statistica v.8 software (from *StatSoft Inc.)* was used in the calculations.

INTERNAL STRUCTURE OF THE MIĘGUSZOWIECKI GLACIERET

The recording of the internal structure of the glacieret was carried out on 29 September 2001 using SIR2 radar gear (*Geophysical Survey Systems*) and the reflexive method (Gądek and Kotyrba 2003). An antenna converter (transceiver) at a frequency of 500 MHz was applied. The time window for registration was 50 and 200 ns. The converter was moved manually across the ice surface. The sounding was carried out along the central line of the glacieret of length 70 m and slope equal to 35°. The velocity of propagation of electromagnetic impulses was determined using the relative dielectric constant with the application of Looyeng's model (Glen and Paren 1975; Ulriksen 1982). The velocity was 1.62*105 km s^{-1}, which corresponds with a dielectric constant $\varepsilon_i = 3.4$.

RESULTS

FLUCTUATIONS IN METEOROLOGICAL CONDITIONS

The analysed meteorological data are shown in Fig. 2. In the period 1980–2001, the average annual snow precipitation total on Kasprowy Wierch was 954 mm of water equivalent (w.e.). The highest values were recorded in balance years 1999/2000, 1997/1998 and 1988/1989. These recorded 1205, 1179 and 1143 mm w.e. respectively. The lowest totals for snow precipitation were recorded in balance years 1998/1999 and 1983/1984, with 726 and 805 mm w.e. respectively.

The average total for rain precipitation was of 842 mm. The highest values were recorded in balance years 2000/2001, 1990/1991 and 1996/1997, when respective totals of 1712, 1202 and 1113 mm were recorded. The smallest amount of rain fell in balance years 1999/2000 (503 mm) and 1989/1990 (580 mm). The highest values for daily precipitation totals were the 166, 146 and 119 mm recorded on 8 July 1997, 14 July 1983 and 16 August 1988 respectively.

Mean annual totals for positive degree-days on Kasprowy Wierch were of 1020.4°C d. The highest values were obtained in the years 1982/1983, 1993/1994 and 1999/2000 and the lowest in the years 1983/1984, 1989/1990 and 1987/1988. In the first cases they accounted to 1233.1, 1200.6 and 1182.9°C d respectively, and in the second to 740.7, 867.4 and 886°C d.

Average wind velocity in the months December-March was of 8.1 ms^{-1}. This oscillated from 5.8 ms^{-1} in the season 1995/1996 to 9.7 ms^{-1} in the season 1987/1988.

The variation to the analysed climatic features across the study period did not show any statistically significant trends.

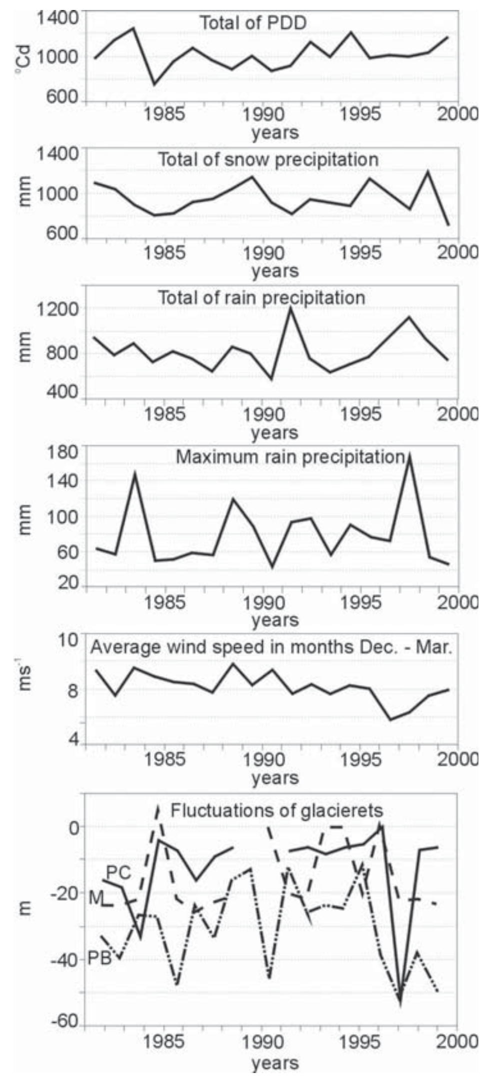

Figure 2. Variability of air temperature, precipitation and wind speed at Kasprowy Wierch (Institute of Meteorology and Water Management data) versus cumulated changes in the lengths of the Pod Bulą (PB), Mięguszowiecki (M) and Pod Cubryną (PC) glacierets in the balance years 1980–2000 (based on data by Wiśliński: IAHS/UNESCO 1993; 1998; 2003).

CHANGES OF FIRN-ICE PATCH FRONTS

Changes of front limits of the Mięguszowiecki, Pod Cubryną and Pod Bulą glacierets in the period 1980–2000 are as shown in Figure 2. These glacierets were changing from year to year, decreasing or increasing in their dimensions. The fluctuations were usually not synchronous (i.e. there was a lack of any statistically significant correlation) and did not show any trends. Similar changes for the glacierets studied were recorded only 5 times, in that in the balance years 1980/1981, 1984/1985 and 1996/1997 all the glacierets were in recession, while in the balance years 1987/1988 and 1994/1995 all of them advanced.

In 1980/1981 the glacierets decreased considerably despite large snow precipitation and an average PDD total. Their recession was connected with melting of their firn edges, which had been preserved in the previous cool summer season. A similar situation applied in 1984/1985. At that time very low snow precipitation totals were recorded. On the other hand, the recession of the glacierets in 1996/1997 seems to have been connected mainly with heavy rain precipitation—especially with the record high maximum daily precipitation. At that time, the Pod Cubryną glacieret showed the most marked recession, and the Mięguszowiecki glacieret the most limited recession. The simultaneous advance of both glacierets studied was recorded in 1987/1988 and 1994/1995, when both relatively large totals for snow precipitation and low PDD totals occurred.

In other years one or each of the monitored glacierets showed different development. In general, fluctuations of the glacierets did not reflect unambiguously the variation to meteorological conditions on Kasprowy Wierch. Statistical significance ($p<0.05$) was only achieved for the correlations between fluctuations of the Mięguszowiecki glacieret and PDD totals, and between fluctuations of the Pod Cubryną glacieret and maximum totals for annual rain precipitation. However, the values of the correlation coefficients (r) were not high even in this case, accounting for -0.67 and -0.73 respectively.

VARIABILITY TO THE INTERNAL STRUCTURE OF THE MIĘGUSZOWIECKI GLACIERET

The obtained results from georadar measurements and photogrammetric monitoring of the Mięguszowiecki glacieret combine with meteorological data from Kasprowy Wierch to show a complex process by which the internal structure of the glacieret developed in the period 1998–2001 (Fig. 3).

The balance year 1998 stood out from others during the 1980s and 1990s in that it had one of the snowiest accumulation seasons. Despite a warm summer, snow from the previous winter survived on the whole surface of the Mięguszowiecki Glacieret. In the balance year 1998/1999 the snow precipitation was lower by about 30%, while the ablation season was very warm. The glacieret net mass blance was negative. A sand-debris cover developed in its highest part. The accumulation period which occurred later was very cool, but there was no significant snow precipitation. However, avalanche accumulation in 2000 was so great that, despite being followed by the warmest ablation seasons of the last 50 years (Niedźwiedź 2005), the mass balance for the glacieret was positive. In the summer of 2001, there was melting of both snow cover from the previous winter and, in part, the firn layer from 2000. This resulted, not only from a winter which had featured only a small amount of snow and a very warm ablation season, but also from a very rainy summer. The annual total for rain precipitation on Kasprowy Wierch was of 1712 mm, while the maximum daily precipitation reached 100.5 mm. The maximum thickness of the glacieret decreased from c. 22 m to only c. 15 m. On its surface, outcrops of annual layers and shear planes were uncovered. In this period two annual layers developed (Fig. 3). These structures did not survive the ablation season of 2003, when the upper part of the glacieret—also the roof of an ice chamber—collapsed (Fig. 4).

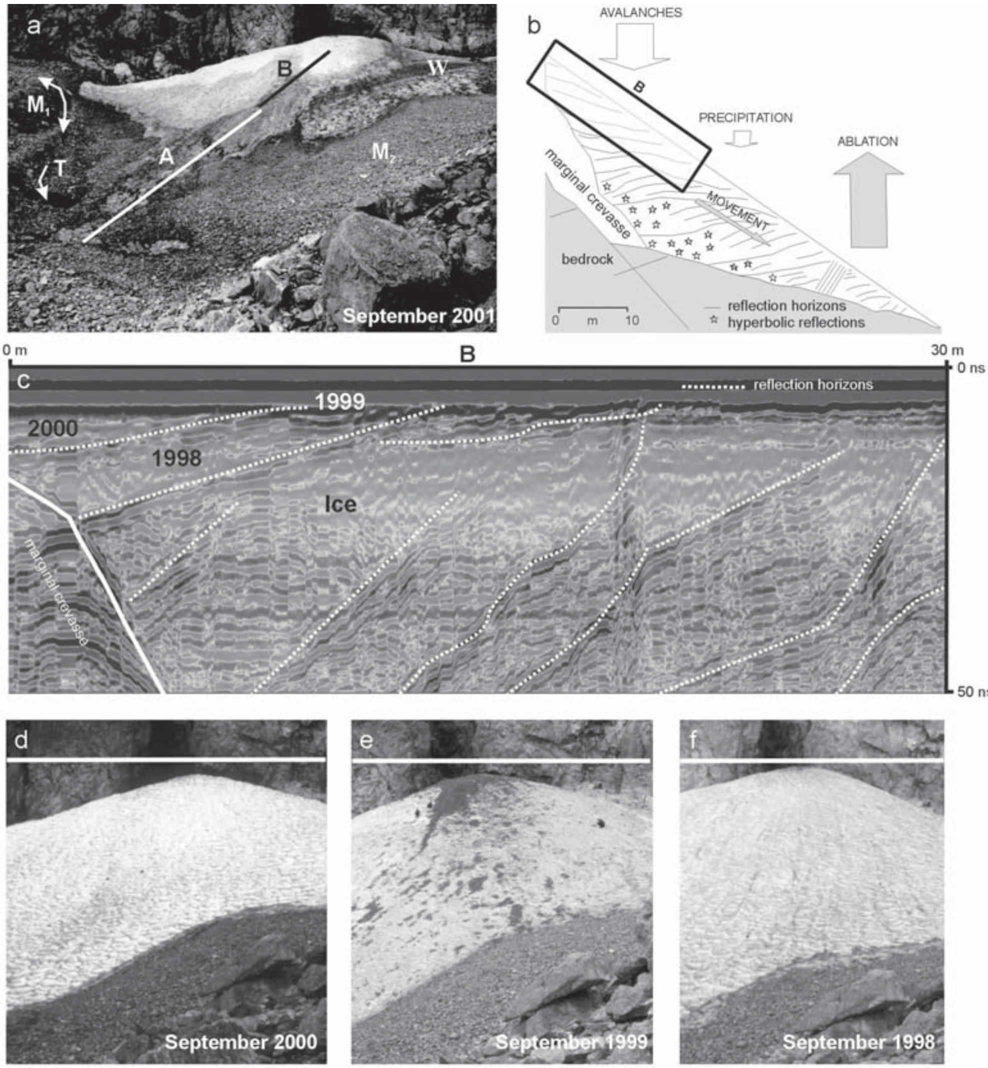

Figure 3. Mass circulation and internal structure of the Mięguszowiecki Glacieret;
a) location of the line of Ground-Penetrating Radar surveys (A, B) of the glacieret
(M1- protulus ramparts, M2—surface moraine, T—subglacial channel opening,
W—outcrops of annual ice layers and shear planes,
b) longitudinal profile and scheme for mass circulation of the glacieret,
c) GPR section of the upper part of the Mięguszowiecki Glacieret in a time window of 50 ns (based on Gądek and Kotyrba 2003),
d)-f) the central part of the glacieret in the period 1998–2000

DISCUSSION

Asynchronous changes of glacieret length as monitored by Wiśliński (2002) attest to a considerable role for topography in the “steering” of their development. The different topography of places where firn-ice forms occur is the reason for differentiation

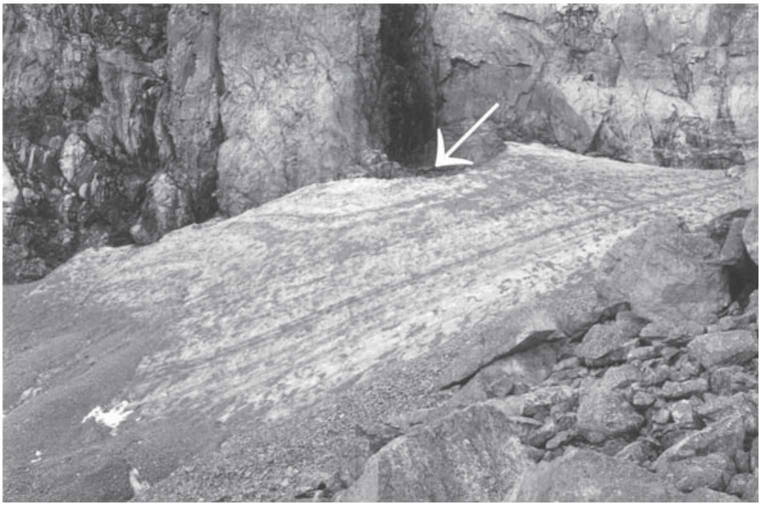

Figure 4. The Mięguszowiecki glacieret in September 2003. The arrow shows the collapsed part.

in the size and structure of both snow accumulation (Lehning *et al.* 2000) and ablation (Hock 2003). The analysed data do not confirm a strong relationship between glacierets and snow patch development on the one hand and air temperature (Hoffman e*t al.* 2007) or wind velocity (Nyberg and Lindh 1990; Daimaru *et al.* 2002) on the other. However, the fluctuations in dimensions affecting these forms have to be connected with variation in meteorological conditions. The lack of a strong relationship between glacieret dimensions and PDD totals, snow or rain precipitation totals and mean wind velocity indicates that the weather regimes of the winter season have a crucial influence. Both avalanche snow accumulation (Lehning *et al.* 1999) and its aeolian version (Greene *et al.* 1999) are connected with that regime.

Owing to the asynchronous fluctuations of glacierets, investigations of their relationships with multi-annual climate variability need a distinction to be drawn between those particular forms nourished mainly by avalanche snow, blown snow, meteoric snow or mixed. Similar classifications of firn-ice patches were proposed by Kłapa (1966) and Higuchi (1968). The altitude at which they occur is also very important. Fluctuations of the avalanche-supplied Mięguszowiecki glacieret did not reflect changes in either the Pod Bulą glacieret (also nourished by avalanche snow but located 300 m lower down), or the Pod Cubryną glacieret (which occurs 100 m higher and is nourished mainly by blown snow). The Pod Cubryną glacieret also showed limited inclination, thereby increasing susceptibility to degradation by heavy rain falls. On the other hand, Jania (1997a) did not record differences in the development of the Mięguszowiecki and Medeny glacierets (Slovak Tatras), both of which represent the same type, occur at similar altitude, have the same exposure as regards the sun and are of similar inclination.

The increase in air temperature recorded in the 1990s in the whole Tatras, which according to Niedźwiedź (2000) does not exceed the standard range to its fluctuations, is not reflected in the development of firn-ice patches. The area and maximum thickness of the Mięguszowiecki glacieret—which is the most sensitive of any of the monitored glaciers to air temperature changes—were changing from year to year, but in 1959, 1982 and 1999 they were similar (Wdowiak 1961; Wiśliński 1985; Gądek 2002). On the basis of the air temperature from Kasprowy Wierch, and assuming that snow ablation expressed in mm w.e. is about three times as great as a PDD total (Braithwaite 1995), it seems that to remain mass balance of the Międuszowiecki glacieret in the period studied, sufficient snow accumulation probably amounted to 2.2–3.7 m w.e.

The results of observations of how the internal structure of the Międuszowiecki glacieret develops indicate that, despite a clear layered structure and evidence of rotational movement (Wdowiak 1961; Gądek and Kotyrba 2003), a correlation of yearly layers with meteorological data is not possible. The scheme for mass circulation of the glacieret is complicated by: a) multi-annual variability in the relation of the accumulation area to the area of the whole glacieret in the range from 0 to 1, b) the possibility of the occurrence of several separate zones of positive mass balance—both in the upper and marginal (moraine-adjacent) parts of the glacieret, c) participation of the marginal crevasse and collapsed forms (associated with subglacial ablation) in mass balance creation. The complexity of the process of mass exchange of permanent firn-ice forms is also evidenced by fossil ice dated at 1000–1700 years, which was found inside the Kuranosuke Glacieret in the Northern Japanese Alps (Yoshida *et al.* 1990), and by buried ice in the substratum of the Medeny glacieret in the Slovak Tatras (Gądek and Kotyrba 2007).

CONCLUSIONS

Fluctuations of the Tatras' firn-ice patches depend most of all on weather regime of winter seasons and local topographic conditions.

Correlation of the fluctuations of firn-ice patches depends on a similar type of snow accumulation (avalanche, aeolian, precipitation, mixed) and the altitude at which they occur.

Variability of the factors influencing the internal structure of firn-ice forms may make dating of these forms impossible. Neither is a reconstruction of their mass balance based on the number and thickness of annual layers possible—even at a scale of one decade and in comparison with meteorological data.

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THE DYNAMICS OF RECENT GEOMORPHIC PROCESSES IN THE ALPINE ZONE OF THE TATRA MOUNTAINS

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Abstract: The energy of high-mountain georelief is evidently transformed into intensity and range of impact of gravitational, water- or snow-induced processes, cryogenic processes, solifluction and deflation. The altitude and climatic conditions of the alpine zone allow for the conservation of some relict or development of some recent processes in the periglacial environment. This paper presents the results of direct measures of some geomorphic processes acting, or said to be active, in the alpine area of the Tatra Mountains. Measurements of debris flows and debris slides, as well as of eolian-nivational, solifluction and ploughing boulder processes, have been conducted at 25 sites distributed across: the Jalovecká Valley in the Western Tatra Mts. of Slovakia, as well as the source area of the Predné Meďodoly and Zadné Meďodoly Valleys in the Belianske Tatry. The results have been compared with those from previous observations.

Key words: debris flows; debris slides; deflation; nivation; ploughing boulder; alpine zone; the Tatra Mountains; Slovakia

INTRODUCTION

A geomorphic process is one entailing a change of state that immediately causes changes in georelief (Minár 1995). The term includes particular material flows and changes in the matter-energetic balance, as well as complex changes of georelief (Urbánek 1974). Recent geomorphic processes are those being observed, or at least considered probable, in a specific region, within the last few decades. The processes usually act together within temporal and/or spatial complexes, thus creating morpho-dynamic systems (Hreško 1994, 1997).

In relation to the intensity and effect of the aforementioned process on humans and the environment geomorphic hazards may also be identified. These are understood to be phenomena entailing the rapid triggering and transport of a great amount of material over a relatively long distance. These most often cause distinct changes in relief, as

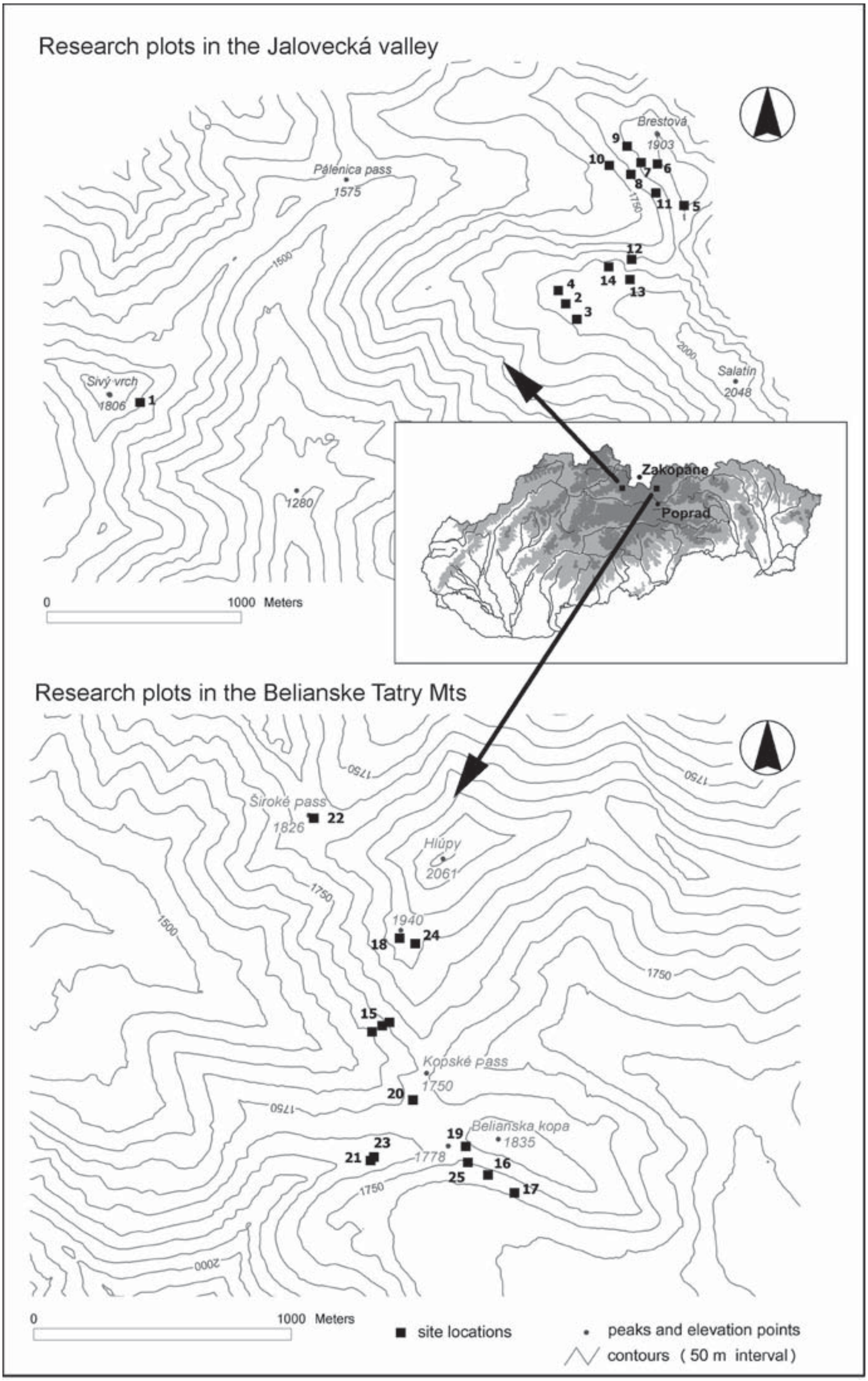

Figure 1. Locations of research plots.

well as damage to vegetation (Rączkowska 2006). Within the alpine area of the Tatra Mountains of Slovakia and Poland they form a group of natural hazards including soil erosion (mediated by water or wind), avalanches, rock falls, debris flows and various forms of slope modelling (Midriak 1974, 1993, 1995; Drdoš 1992; Hrnčiarová 1996; Kotarba 1996; Hreško 1998, 1999, 2000; Hreško, Bugár 1999, 2000; Rączkowska 2006; Boltižiar 2007).

The aim of the paper is to present the current state of research as regards recent geomorphic processes, on the basis of direct field measures of the dynamics to these processes.

STUDY AREA

The research sites are located within the alpine zone of the Slovakian Tatra Mts. being distributed across two separate areas distant from one another, i.e. the Jalovecká Valley (in the Western Tatras) and the divide area between the Predné Meďodoly and Zadné Meďodoly Valleys, which is partly in the contact zone between the eastern part of the High Tatras and the Belianske Tatry (Fig. 1).

Within the Jalovecká Valley, measurements were made at 14 sites of altitudes 1750–1930 m a.s.l. Geologically, the area falls within the region of crystalline rocks (schist, gneiss and migmatite) and of granitoids (granodiorite). One site is located in the marginal Mesozoic part of the territory in which limestones, dolomites and various slates are dominant (Nemčok *et al.* 1993, 1994). Depending on the elevation and topography, climates here oscillate between the moderately cold and the cold (Konček 1974). Total annual precipitation varies from 950 to 2000 mm. The long-term means for annual temperature range between +3°C and 0°C. Shallow alpine soil types, podzols and podzolic leptosols are dominant here. Vegetation cover consists of natural alpine grasslands with a prevalence of the *Juncion trifidi* association. Small clumps of dwarf pine (in *Pinion mugi*) diversify this relatively homogeneous vegetation structure at lower elevations.

The measurements in the Belianske Tatry were conducted at 10 sites of altitudes 1600–1850 m a.s.l. Mesozoic rocks (limestones, dolomites and Werfenian slates) dominate in this part of the mountains. Climatically, the territory comprises moderately cold or cold mountainous areas with mean annual temperatures ranging from +2°C to +4°C (Konček 1974). Mean annual precipitation is in turn of 900–1200 mm. Soils reflect the natural conditions close to a periglacial environment. The typical vertical zonation of soils is not developed here. The structure of soil cover is represented by lithic leptosols, leptosols, rendzinas, cambic rendzinas, podzolic cambisols and some other azonal subtypes (Bedrna *et al.* 2000). Vegetation reflects more fertile soil conditions on calcareous substrates through the presence of varied grassland communities of Class *Elyno-Seslerietea*, while communities of the *Juncetea trifidi* association occupy the less-fertile areas. Small clumps, or rarely larger areas, of dwarf pine stands complete the vegetational mosaic.

METHODS

While fieldwork to measure the intensity of recent geomorphic processes has been in progress in both areas since 2001, the Jalovecká Valley research follows work conducted in the 1990s (Hreško 1997). Frequency of measurement has depended on the type of process but has not taken place more rarely than once a year. A chronological database of photographs has been created for each site. A set of direct measuring methods has been used in the research: a) the method of one moving and one fixed point; b) the method of one moving between two fixed points; c) the method of coloured lines and squares.

The first method is based on the measurement of distances between one fixed and one "moving" point. The fixed point (a steel rod in the earth, or a mark on a stable boulder) has to be set in a position in which much more limited geodynamic activity is assumed (even no activity at all). The "moving" point is located in a position in which activity is expected again taking the form of a steel or wooden stick, or else a mark on an unstable boulder). This method is used in the measurement of solifluction-gravitational processes, as well as the activity of ploughing boulders.

The method of a moving point between two fixed ones is used in measuring the edges of eolian-nivational niches and changes therein. The method is based on the measuring of the distance from a fixed point located outside the niche to the niche edge (the moving point) in the direction of the next fixed point inside the niche area.

The method of coloured lines and squares is used to measure the activity of the accumulated material within debris flows and debris slides, gravitational creep, or eolian deflation of debris particles. By means of the coloured straight lines sprayed (between two fixed points) on debris perpendicularly to the flow direction shifts of debris fragments can be measured over a specific time period. The coloured sampling squares (50 x 50 cm) are of 3 size categories according to the size of debris particles, with diameter limits up to 0.5 cm, 2 cm, and 5 cm, sprayed on the debris surface. They have been used to measure both shift distances and the amount of moved material (recalculated to square metres or as percentages). Where wind activity is concerned, only the shifts aside of the gravitationally conditioned flow direction can be taken into account. In addition to the distance measures, prevailing wind directions and wind intensity can also be estimated.

DEBRIS FLOWS

Debris flows can transport a considerable amount of saturated fragments of weathered

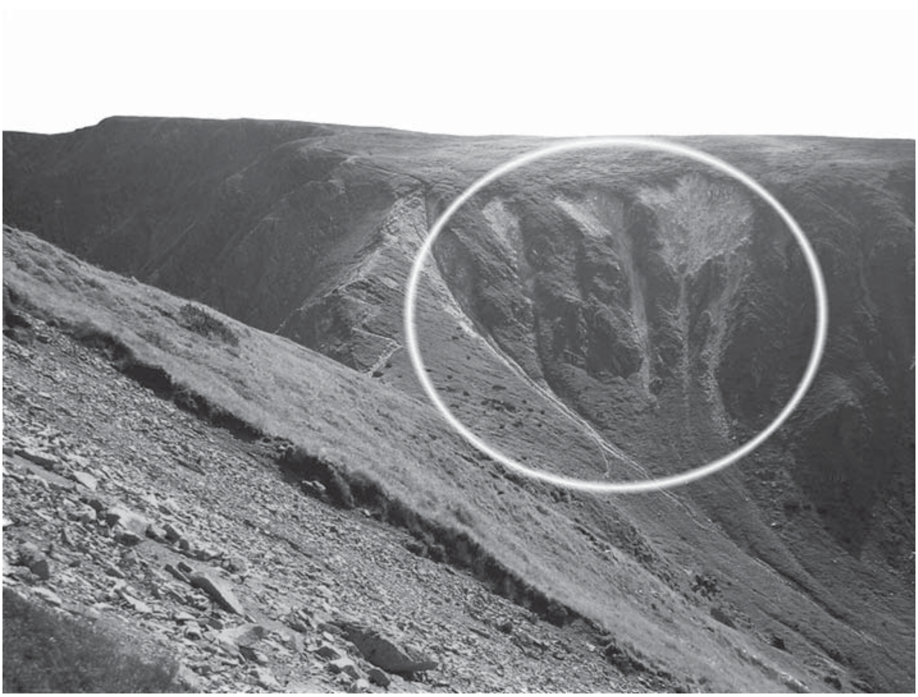

Figure 2. Debris flows (at 1950–1750 m a.s.l.) with eolian-nivational niches at starting zones; N slopes of the Grapy side branch of Mt. Salatín (Jalovecká Valley).

Figure 3. The activity of some debris flows (at site no. 15 in the Belianske Tatry Mts.) is evident. A pair of photographs shows the same location from the Belianske Tatry Mts in 2005 (left) and 2006 (right).

rock being accumulated below crags, furrows and gullies, often without any noticeable motion for several years or decades. The causes of debris flows can vary, but a frequently postulated triggering mechanism is the influence of water and the resulting increase of pore-water pressure (Hürlimann *et al.* 2003). In the Western Carpathians, the intensity of these processes have been investigated and evaluated by several studies (Lukniš 1973; Mahr 1973; Nemčok 1982; Midriak 1983, 1993; Kotarba 1996; Hreško 1994, 1996 a, 1996 b, 1997; Barka 2005; Rączkowska 2006).

In the area of the Jalovecká Valley it has proved possible to identify 47 debris flows, mostly located at altitudes of 1700–2100 m a.s.l., on slopes of 20–50° both north and south facing (Fig. 2). The methods used do not entail measurement of the frequency of occurrence, size and dynamics of the debris flows in the strict definition of this phenomenon. Direct measures have focused on the dynamics of accumulated material after the real process of debris flow has become relatively stabilised. In line with the relationship between the transport dynamics and sizes of rock fragments, the material of debris flows has been categorised into (A) a group in intensive motion (mean size of fragments less than 2 cm); and (B) a group without or with less intensive motion (mean size of fragments greater than 5 cm). Mean annual movements of some centimetres (group B) up to 4 metres (group A) have been measured.

In the area of the Belianske Tatry Mts., the observed debris flows are located at altitudes of 1650–1750 m a.s.l., on SW-facing slopes with inclinations of 25–45°. The measured values of the shift (for mean size of debris fragments = 5 cm) in the source zone are 0.5–1.5 m/year; cf. 1.5–10 m/year in the transport zone and 0.3–1 m/year in the accumulation zone. The most intensive transport activity is observed each year after the spring melt, as well as at various times during heavy rainstorms, when it is mainly the finer debris that moves down—prevailingly along the centreline of the debris flow (Fig. 3).

DEBRIS SLIDES

It is displacement of a soil profile including weathering mantle and, possibly, periglacial debris that creates debris slides. Unlike the debris flows, these gravitational forms accumulate coarser material in bottom parts (Mazúr 1955). The generation of debris slides is associated with smooth hillsides with slope angles over 30°, once a waterlogged debris layer becomes separated from the subsoil and begins to move down (Plesník 1971; Lukniš 1973). This process nevertheless resembles that involvd with debris flows in being conditioned by water supply and gravitation. Debris slides mainly affect shallow layers of fine debris.

In the Jalovecká Valley, the occurrence of the latter has been registered on slopes with angles in the range 20–50°, at altitudes of 1600–2100 m a.s.l. generally in south-facing areas (Fig. 4). The measurements of debris dynamics have been made by means of coloured lines (distance measures) and squares (distance and area measures). Differences in dynamics in relation to the size of fragments have been determined as well. While the fragments of mean size greater than 5 cm have been transported up to 1 metre, rarely 2 metres per year, those smaller than 2 cm moved 2–4 metres per year. Results of measurements from the 1990s (Hreško 1997) reveal shifts of 0.4–1.2 metres per year.

EOLIAN-NIVATIONAL PROCESSES

Wind activity represents a significant phenomenon in an alpine environment (Midriak 1983). The effect of wind erosion is expressed mainly in deflation that causes the transport of fine debris particles on (or above) the surface, with larger residual frag-

Figure 4. Extensive areas of debris slides on the SW slope of Mt. Brestová (1931 m a.s.l.).

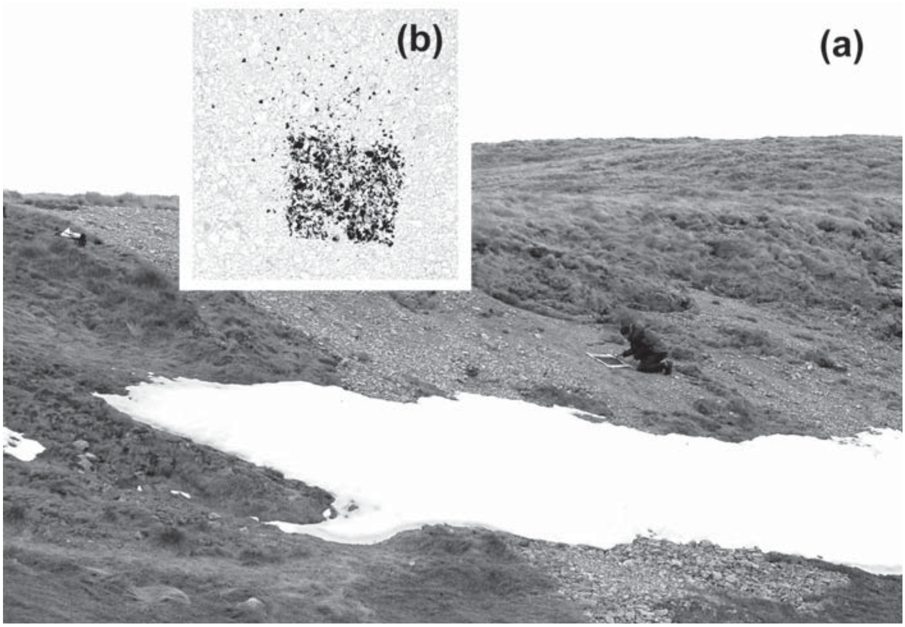

Figure 5. Installation of coloured squares
(a) on the surface of the eolian-nivational niches (at 1925 m a.s.l.) below the plateau of the western branch of Mt. Salatín;
(b) displacement of debris fragments after 5 months (in the May-October period).

ments creating patches of coarser material. The intensity of the process can increase when the ice crystals drift in the wind. Indirect eolian activity is connected with snow transfer from windward to leeward sides of crests, thus forming snowdrifts as sources for either avalanche triggering or the nivation processes characteristic of long-term snow patches.

The erosive effect of snow accumulations (nivation niches) on the surface is significant in areas with discontinuous vegetation cover, though it differs in relation to slope type, in that there is lithologically-controlled fragment-size composition of cover. The role of melt water is thus different in various lithological types of niches (Rączkowska 1995).

Observations on snow dynamics in relation to snow redistribution by eolian drift and its role in water balance have also been made in the Jalovecká Valley (Holko, Kostka and Parajka 2003).

Patches affected by eolian deflation and nivation have been identified at 114 sites in the Jalovecká Valley, these occupying a total area of 1.8 ha. They are usually located on crest positions, close to the edges of summit plateaus or saddles at altitudes of 1800–2100 m a.s.l., on gentle slopes of various aspects, if with a prevalence of N-NW and S-SW sectors (Fig. 5).

Measurements were made at 7 sites (see Table 1 and Fig. 1). The results from observations based on coloured squares (Fig. 5) reveal losses of fine (less than 0.5 cm) particles equal to 50–70% of the square area. This compares with 0–30% a year in the case of coarse debris of up to 5 cm. The distance measurements as regards debris transport by nivation have detected shifts of 10–200 cm/year depending on slope inclination and the size of debris fragments. Drifts of fine particles by means of wind action have reached 10–110 cm/year, while drift angles indicate the prevalence of the S-SE wind directions.

Table 1. Characteristics of the sampling sites.

Site No.	Process/Form	Altitude [m a.s.l.]	Mean slope [deg.]	Aspect	Soil and/or Lithology	Land Cover (in site / vicinity)
The Jalovecká Valley						
1	ploughing boulder	1725	35	SE	dolomitic block	grassland / dwarf pine
2	eolian-nivational	1920	5	SW	fine granitoid debris	debris / grassland
3	eolian-nivational	1920	15	SW	granitoid debris	debris / grassland
4	eolian-nivational	1920	5	SW	granitoid debris	debris / grassland
5	eolian-nivational	1925	3	SW	granitoid debris	debris / grassland
6	eolian-nivational	1890	7	SW	fine granitoid debris	debris / grassland
7	solifluction lobes	1880	30	SW	Podzolic Leptosols on coarse granitoid debris	grassland
8	ploughing boulder	1875	30	SW	granitoid block	grassland
9	debris shift	1875	30	SW	granitoid debris	debris / grassland
10	debris shift	1850	32	SW	granitoid debris	debris / grassland
11	debris shift	1880	35	SW	granitoid debris	debris / grassland
12	debris flow	1930, 1800	35	N	granitoid debris	debris / grassland
13	eolian-nivational	1935	7	N	fine granitoid debris	debris / grassland
14	eolian-nivational	1935	15	N	fine granitoid debris	debris / grassland
The Belianske Tatry Mts.						
15	debris flow	1675, 1700, 1725	35	SW	dolomitic debris	debris, tufts of grass / grassland
16	debris flow	1740	35	S	coarse dolomitic debris	debris / grassland, dwarf pine
17	debris flow	1720	35	S	coarse dolomitic debris	debris / grassland
18	debris creep	1916	40	SW	dolomitic debris	debris, tufts of grass / grassland
19	debris creep	1780	35	SW	coarse dolomitic debris	debris / grassland
20	eolian niche	1763	3	N	fine slate debris	debris / grassland
21	eolian niche	1830	15	N	fine slate debris	debris / grassland
23	eolian niche	1825	3	NW	fine slate debris	debris, tufts of grass / grassland
24	solifluction lobes	1920	30	E	Podzolic Cambisols on coarse debris	grassland
25	solifluction lobes	1760	30	S	Cambic Rendzinas on coarse calcareous debris	grassland

Measurements of the degree of widening of niches by means of abrasion of edges have been made at 2 sites. Average shifts of 0.5 up to 2.5 cm per year have been reported. Previous research (Hreško 1997) reported values of 0.5–1.5 cm/year, or 5–10 cm/year in sporadic cases.

Average values from measurements made at sites in the Belianske Tatry Mts. vary from 0.5 to 2 cm/year. Kotarba (1976) gives values of 3–16 cm/year for limestone area at sites at similar altitudes. Our results also indicate seasonal retrograde fluctuations, in that, for example, the direction of the shift between May and August 2001 was counter to the total annual shift between May 2001 and June 2002. It is assumed that this is caused by soil cohesion and volume changes during periods of freezing and thawing.

SOLIFLUCTION

Solifluction, broadly defined as slow mass wasting resulting from freeze–thaw action in fine-textured soils, involves several components: needle ice creep and diurnal frost creep originating from diurnal freeze–thaw action; annual frost creep, gelifluction and plug-like flow originating from annual

Figure 6. The Jalovecká valley. Sequence of solifluction-gravitational lobes (highlighted by white dashed lines) on the SW slope of the Mt. Brestová side branch (at 1875 m a.s.l.). Black arrow indicates position of “moving point” on garland vertex.

freeze–thaw action; and retrograde movement caused by soil cohesion (Matsuoka 2001). Lukniš (1973) described solifluction lobes as elongated steps on slopes inclined by more than 22° on which turf soil is pushed up into arched ramparts at the front. The lower limit of solifluction activity in the Western Carpathians is to be found at altitudes of about 1700–1800 m a.s.l. (Kotarba 1976).

Relevant measurements in this case were made at 3 sites: one in the Jalovecká Valley (at an altitude of 1875 m a.s.l.), and two in a region of the Belianske Tatry Mts. (at 1920 m and 1760 m a.s.l.). The heights of lobes vary from 0.5 m to 1.5 m, while lengths are of 3 up to 5 metres (Fig. 6). Values measured in the Belianske Tatry Mts. indicate displacements of 1–2.5 cm/year; in the Jalovecká valley the average values are 0.5–2 cm/year. Previous observations (Hreško 1997) have not been recorded any displacements.

PLOUGHING BOULDERS

Ploughing boulders represent a relatively common phenomenon in the periglacial environment. They occur in areas of active solifluction, on frost-susceptible soils with low plastic and liquid limits allowing for the frost heaving and creeping of blocks. During movement these rotate to adopt an alignment of least resistance (Berthling *et al.* 2001; Ballantyne 2001). The co-occurrence of ploughing boulders and stony lobes has been confirmed (Lukniš 1973; Garcia-Ruiz *et al.* 1990).

A complex form consists of three elements—the boulder itself (usually of 1–1.5 m in diameter), and its frontal mound and upslope furrow as evidence of displacement. Mean annual shifts are generally of less than 1 cm, only sporadically exceeding 2 cm/year (Kotarba 1976; Hreško 1997). Measurements made in the period 2001–2007 revealed shifts of 0.5–1.5 cm/year on average (Fig. 7).

Figure 7. Measuring ploughing boulder (site no. 10) displacement on the SW slope of Mt. Brestová.

CONCLUSION

Research on periglacial processes assumes particular importance in light of the need to identify climate change. The activity or passivity of the process, or else signs of changing intensity can point to both long-term climatic shifts and short-term fluctuations. Detailed monitoring-like observation supports the evaluation of such environmental problems as erosional processes, changes in the structure of vegetation associations and the possible invasion of allochtonous species into disturbed patches.

The results presented here confirm that processes have been maintained at relatively constant intensity. However, since 2000, the alpine environment has experienced a moderate increase in fluviation and more frequent activity as regards processes linked to nivation. However, final confirmation of this trend will require testing of the correlations with climatic data.

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CONTEMPORARY LANDFORM DEVELOPMENT IN THE MONTS DORE MASSIF, FRANCE

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Abstract: The Monts Dore massif is a typical high-mountain area with nival relief dominant above the timberline. The massif is situated at the centre of the Massif Central and constitutes its highest part. Nivation, cryogenic processes and deflation are the key morphogenetic processes, but runoff action, especially on slopes with deteriorated vegetation cover, plays an equally important role due to the dominant maritime climate.

Key words: Massif Central, cryo-nival processes, nival relief, morphogenetic altitudinal zones

1. INTRODUCTION

The morphogenetic structure of high mountains generally includes perennial frost, glacial, cryo-nival and denudational or denudational-fluvial altitudinal zones (Kaszowski 1984). Such zonation is found in all high mountain systems, regardless of latitude, though some ranges may feature less than complete vertical zoning sequences, depending on the surrounding climatic zone. The general surrounding climate most influences the lowermost altitudinal zone, i.e. the denudational or denudational-fluvial zone. As the Monts Dore massif, the subject of this study, is located in the temperate zone, high mountain relief is typically shaped by episodic extreme processes. During such events, increased volumes of debris material are mobilised, transported away from the mountains and mostly deposited in the foreland. The bulk of the material washed away from the mountains comes from their lower parts, while the highest parts typically only contribute granularities up to the size of gravel.

This study aims to identify the role of contemporary morphogenetic processes in the shaping of the Monts Dore massif, a glacier-free mountain range of the temperate zone featuring a typical maritime climate.

2. STUDY AREA

The volcanic Monts Dore massif, a typical high-mountain geomorphological system, forms the highest part of the Massif Central, France (Fig. 1). The Monts Dore massif meets C. Troll's criteria for an Alpine mountain area (1973), i.e. it rises above the timberline (peaking at 1886 m a.s.l.), as well as above the Pleistocene line of permanent snow, and features active peri-nival processes.

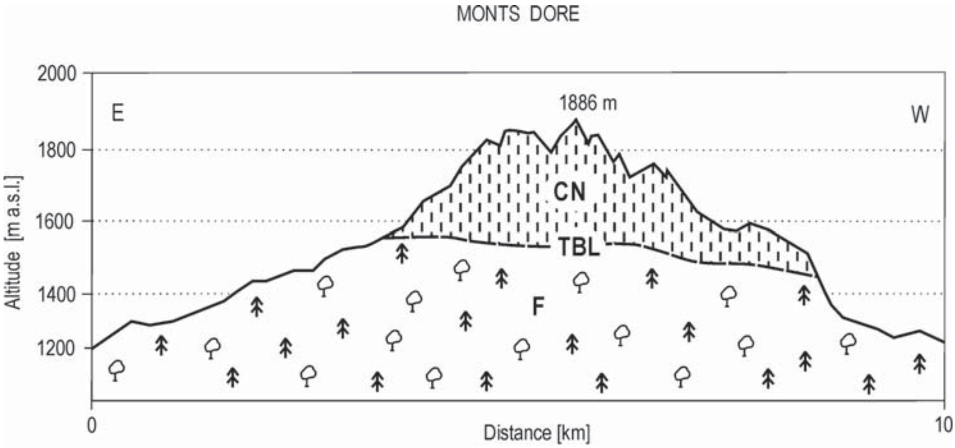

Figure 1. A cross-section of the Monts Dore massif:
CN—cryo-nival zone; F—forest zone, TBL—timberline.

Monts Dore is a Tertiary strato-volcano built of various types of lava (basalts, sancyites, doreites, phonolites, trachytes) and pyroclastic formations (Peterlongo 1978). During the Pleistocene age, glaciers of lengths between 12 and 19 km flowed centripetally from its centre (Veyret 1978). Active volcanic processes occurred at the peripheries of the area in the Holocene.

Monts Dore features typical landforms of a high-mountain geomorphological system, including rock-ridges, cliffs and talus slopes. Most of the slopes, however, are covered with waste-mantles and relatively thick organic covers. The timberline, at 1550 (1450–1500) m, splits the area into two morphodynamic systems: a cryo-nival system above and an erosional-denudational system below (Kaszowski and Krzemień, 1989, Fig. 1).

A wet maritime climate predominates in the area. The highest weather gauging posts are located at 1000–1100 m a.s.l. Measurements are also taken in the local winter resorts, such as Super-Besse (1350 m a.s.l.) and le Mont-Dore (1050 m a.s.l.), but their record history is not long enough to allow for meaningful statistical processing (Valadas 1984, Lageat, Neboit-Guilhot 1989). The annual precipitation in the highest parts of the Massif Central exceeds 1500 mm (1980–1985 mean). According to P. Estienne (1989), Monts Dore receives between 1700 and 2000 mm (1980–85 mean). However, peak results reach 2098 mm at Super-Besse (1982). The Besse weather station (1060 m a.s.l.) recorded annual precipitation ranging from 940 mm in 1957 to 1811 mm in 1965. Most of the annual precipitation falls in wintertime (December-January) and least in summertime (June-July) (Krzemień 2004). However, the nature of the summer rainfall is such as to favour occasional rainstorms causing extensive torrential erosion. Peak singe-event rainfalls recorded at Besse reached 71.7 mm, and there are probably falls of more than 100 mm in the higher parts of the mountains. While these figures may not be particularly high in themselves, it is possible for such intensive rainfall to last for several days, allowing for an intensification of the rates of geomorphological processes. For example, in 1999, a three-day event yielded 181 mm of rain falling on snow-covered slopes. Similar conditions occur on average every 6 to 10 years, and are usually followed by the most significant slope transformation events and highest water levels in rivers. Some of the largest morphological transformations, however, are not linked with extreme rainfall events. On 5 November 1994, for example, the upper part of the Chaudefour Valley experienced debris flows and torrential streams and was hugely transformed in the process (Fig. 2).

Snow cover varies from year to year and from area to area, due to the effect of snow blow. The number of days with snow increases with elevation, the records from Super-Besse showing 138 days in the winter of 1977/78 and 132 days in 1978/79. The first snow falls in the second half of November, while the last snowstorms are sometimes recorded in late May (Valadas 1984). The snow stays on the ground until the end of June and even into July, and this provides material for morphogenetic processes. The marked variability of snowfall and temperature causes snow-cover thickness to vary greatly during each winter season. Snow thickness varies between 150 cm in the summit areas and

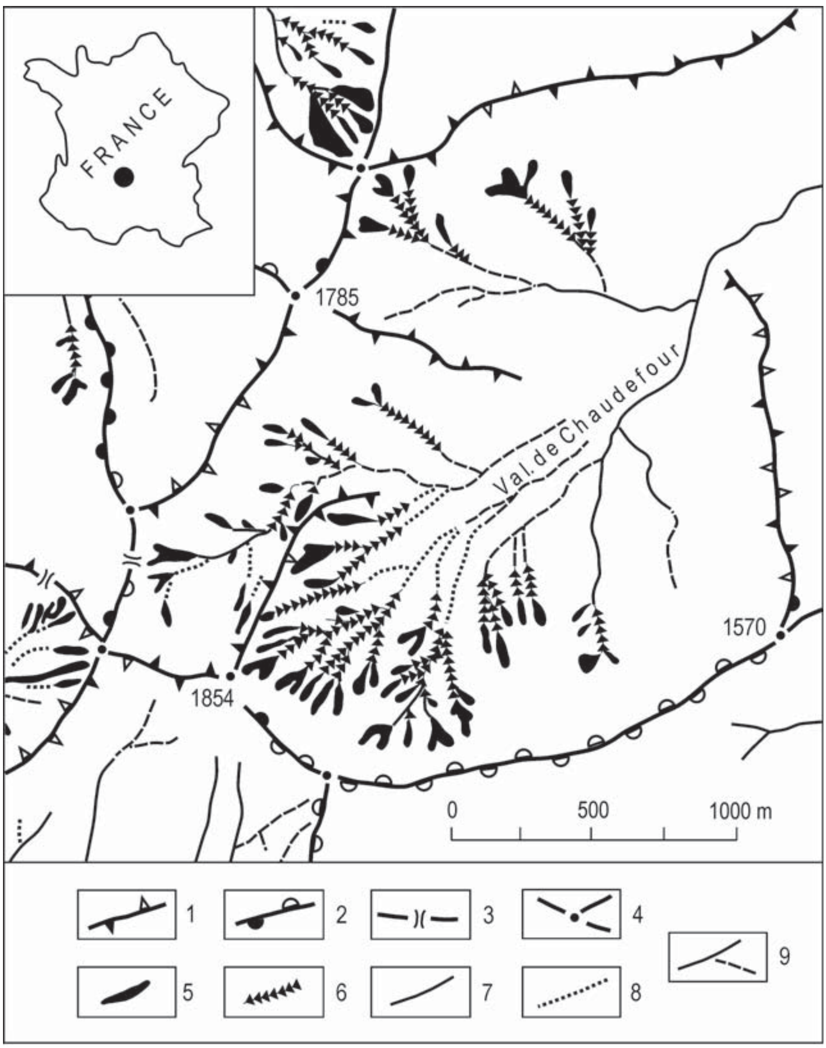

Figure 2. Map of debris flows in the Monts Dore massif:
1—narrow ridges; 2—broad ridges; 3—mountain passes, 4—summits, 5—debris-flow hollows, 6—gullies, 7- debris-flow chutes, 8—debris-flow toes, 9—periodic and permanent stream channels.

90 cm at the Monts Dore station. Between January and March, the Col de la Croix St. Robert (1450 m a.s.l.) is blocked by snow and the mountain road is normally closed, but there are years in which the snow disappears entirely during that period (Lageat, Neboit-Guilhot 1989). The number of days with negative temperatures, another important factor for morphogenetic processes, is more than 150. The average annual temperature in the upper parts of the massif is less than 5°C (Tort 1989).

Despite a lack of accurate meteorological records, it seems certain that the 1000 m elevation difference must produce an altitudinal climatic stratification in the Monts Dore massif. This stratification is reflected in altitudinal differentiation to the rates of morphogenetic processes and the role they play in the development of the massif's landforms. The general pattern is rather peculiar, and involves a large amount of winter precipitation, long-lasting snow cover, frequent thaws and numerous temperature freeze-thaw cycles—all combining to suggest a potentially important role for nivation processes and pipkrake activity (Figs. 3, 4). Additionally, as the massif is exposed to strong winds, a possible major role for deflation is indicated, wherever the waste mantle is not protected by vegetation cover (Figs. 4, 5).

Monts Dore has an annual morphogenetic pattern consisting of three distinct periods: 1) a dead winter season (November

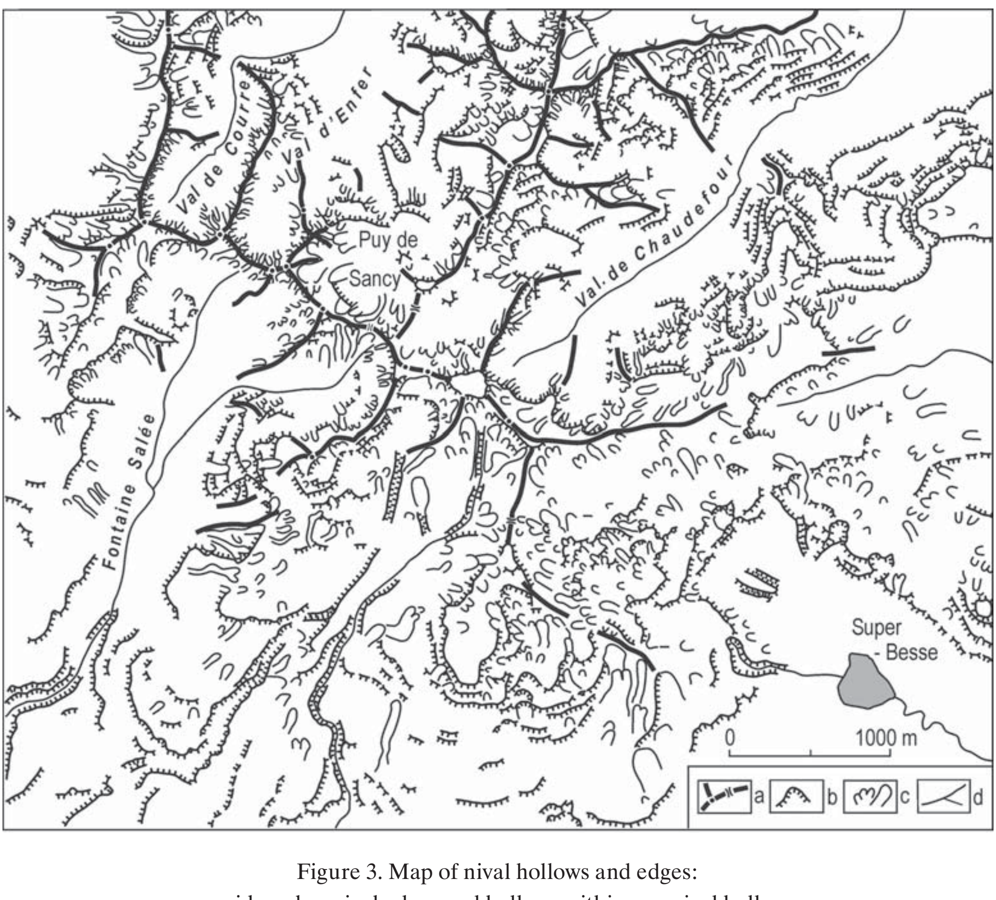

Figure 3. Map of nival hollows and edges:
a—ridges, b—nival edges and hollows within, c—nival hollows.

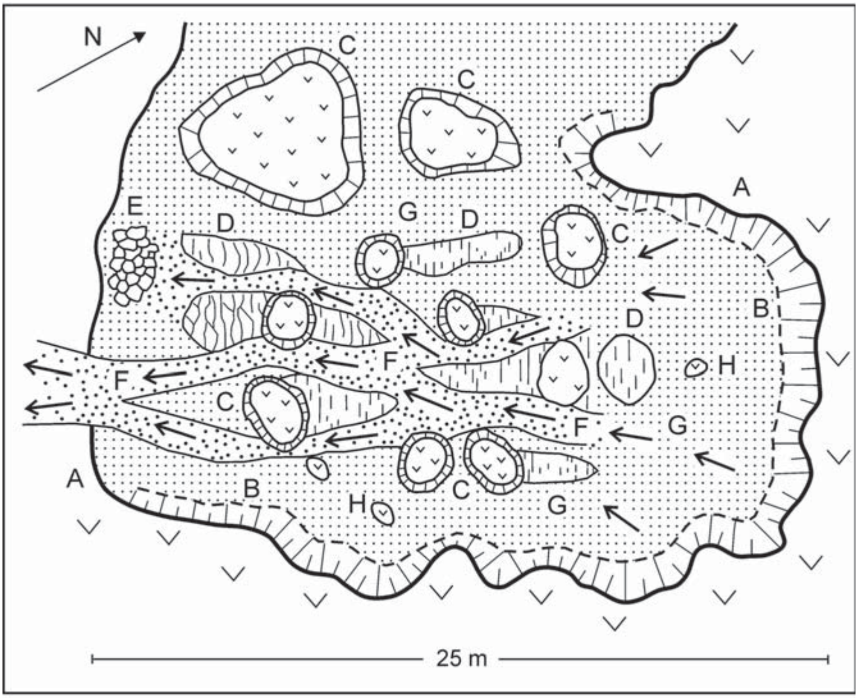

Figure 4. A nival hollow in 1989:
A—turf-consolidated surfaces, B—area subject to deflation and needle-ice,
C—remnants with a turf cover, D—area subject to rainsplash shaping with earth pyramids,
E—area covered with residual debris, F—area shaped mainly by concentrated wash,
G—area shaped mainly by dispersed wash and deflation, H—remnants with high content
of humus and no turf cover.

-December); 2) a nivation period as snow cover and isolated snow patches thaw (typically in April—June); and 3) a pluvial season involving runoff action (July—October).

According to M. Tort (1989), the summit belt at 900–1500 m a.s.l. is mainly covered with beech forest or (on the western side of the massif) by spruce forest. The subalpine belt is occupied by natural and human-induced glades in which pedoclimatic conditions have prevented trees from taking root.

The thickness and type of slope covers depend on the bedrock and its postglacial transformation. They vary widely (Libelt 1995). Gradual slopes (mainly in the southern and south-eastern part of the area) are covered by peat of 0.5–0.7 m thickness. Steeper slopes (around the Puy de Sancy and Vallée de Chaudefour) are covered by loam/peat/stone covers and loam/stone covers. Trachyandesite outcrops are covered by regolith, often subject to solifluction, while loamy or loam/stone covers have developed on pyroclastic formations. Where the vegetation cover is destroyed the covers have degraded towards the more stony types.

3. METHODOLOGY

Field research in the area involved geomorphological mapping of landforms, covers and the effects of morphogenetic processes on 1:10 000 maps (enlarged from 1:25 000 maps). Special attention was given

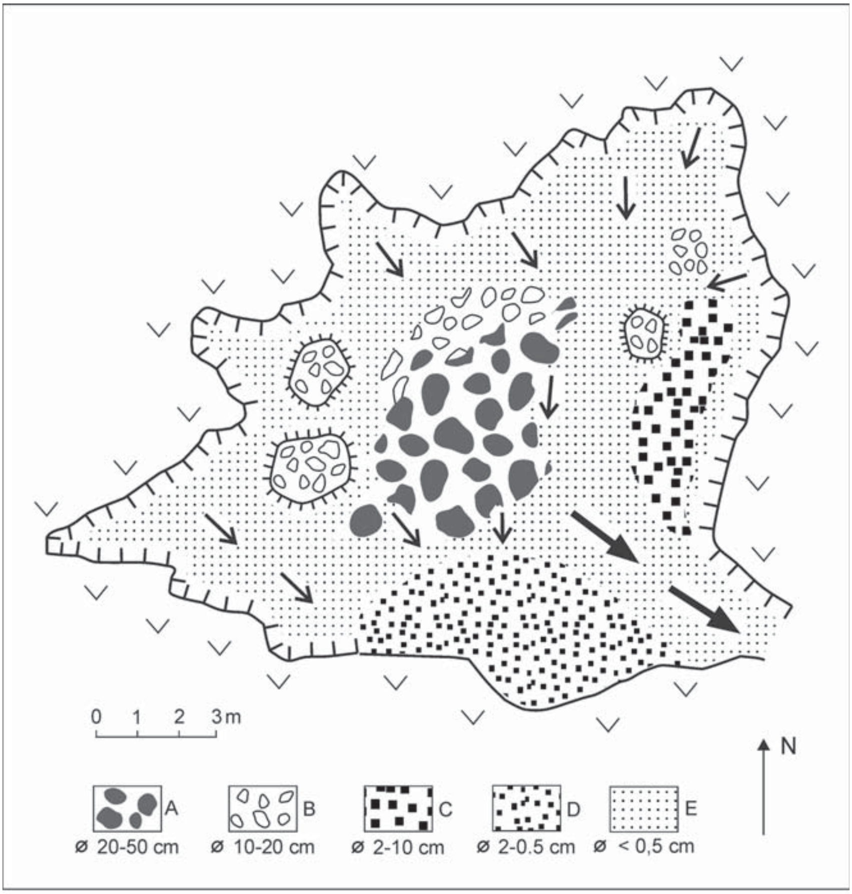

Figure 5. A nival hollow with a very active dissection process cutting into coarse waste mantle; waste mantle granularities:
A—20–50 cm, B—10–20 cm, C—2–10 cm, D—0.5–2 cm, E—0.5 < cm.

to: nival hollows, debris flow chutes, select stream channels and anthropogenic forms. Detailed drawings, cross-sections and longitudinal sections were made of the most characteristic features, and the entire effort was supplemented with sketches and numerous photographs. The study also used aerial photos at the scale of 1:25 000 from various periods, and unpublished research work from the Institute of Geography of Blaise Pascal University. The fieldwork providing the bulk of the material used in the study was carried out over several 2–3 week study trips, mainly in spring and summer, starting in 1984.

4. DOMINANT FORMS AND PROCESSES

The geomorphological maps of the Monts Dore massif present the main features as regards landforms and their dynamics. Three types of landform sequences are identified in the longitudinal section of the slopes and valley bottoms (Figs. 2 and 3, Kaszowski 1995):

1. nival hollows—small nival valleys—fluvial fans;
2. nival hollows—proluvial fans—nival hollows—(nival valleys);
3. nival hollows or hollows of debris

flow systems—debris flow cones—alluvial plains.

Medium and micro-scale forms are indicative of the area's contemporary morphodynamics. The following major examples of these were discovered during the fieldwork:

(a) systems of hollow-like dips in slopes above the timberline, mostly consolidated by vegetation and forming a characteristic microrelief of nivation origin;
(b) systems of gelideflation edges and hollows (Fig. 3);
(c) deflation floors in high-mountain passes and also accompanying roads and tourist trails;
(d) erosional dissection of slopes and systems of proluvial fans;
(e) debris flow chutes and cones forming extensive vertical morphodynamic systems starting at near-ridge hollows and funnel pits and continuing down to the timber line (Fig. 2);
(f) rubble fields and streams only at a local scale and associated with block-type rock weathering.

The landforms listed have developed throughout the Holocene, and continue to do so, though the morphodynamic role of certain processes has increased through human involvement (Kaszowski 1995, Krzemień 1995). Indeed, the most important type of contemporary process in the cryonival morphogenetic zone is the nivation type associated with long-lasting snow patches. (Figs. 3, 4, 5, 6, 7 and 8). This is followed by cryogenic processes and deflation (Figs. 8 and 9). Pipkrake plays a particularly important role (Figs. 4 and 5). All these processes have been intensified anthropogenically, especially along tourist trails and ski pistes, as well as long roads (Fig. 10). Finally, a very important role is played by rainwater in the form of rill-flow and debris flow (Fig. 2). The erosional role of water comes into play mainly where vegetation cover has been degraded (Fig. 10).

Figure 7. Nival hollows on the broad hilltop of the Plaine des Moutons.

Figure 6. The northern slopes of the Puy Fernand.

Figure 8. An actively modified nival hollow near the Col du Couhay Pass.

Figure 9. A partial view of a nival hollow under the influence of aeolian processes near the Puy Fernand. There is a visible zone of fine gravel accumulation.

Figure 10. A degraded slope occupied by a ski-piste between the Puy Fernand and the Puy de Gros.

Nival forms. There are two generations of nival forms:

1. hollows, kettles and little valleys consolidated by sod cover;

2. irregularly-shaped hollows without vegetation cover and with a degraded soil cover, subject to very intensive contemporary processes (Figs. 3 and 4).

The first generation forms are stable and older. Their origin may be explained by long-term postglacial development in a wet maritime climate, followed by accelerated destruction as a result of increased human activity in the past.

The second generation forms have developed during the last few decades. They are an effect of contemporary human activity, especially tourism and winter sports. New tourist paths, roads, ski-lifts and cable cars, as well as other structures, are the nuclei of accelerated erosion. Above the timberline, patches of snow are the most important geomorphological factor behind erosional activity.

Deflation and gelideflation hollows occur in slope flattenings, on ridges and mountain passes. The hollows range in size from roughly one by one metre to more than a dozen metres each way. They may be isolated or may form systems, and are irregular in shape, with numerous bays and peninsulas bearing marks of gelideflation edges. Inside each hollow there are normally between a few and several dozen gelideflation remnants, separated by barren areas reshaped by the action of pipkrake, sheet flow, aeolian processes and accumulation of the displaced material (Figs. 5 and 8). The destruction of the vegetation cover may be due to natural or artificial causes.

Deflation floors occur in high mountain passes and in association with roads and tourist trails, normally measuring ca. one by one metre (Fig. 8).

Erosional dissections accompany roads, ski-pistes and routes of former t-bar lifts. They take the form of long chutes (up to 700 m) between 0.5 and 5 metres wide (Krzemień 1995). They can develop as rapidly as during a single rainfall event, but may sometimes take a dozen years to develop, as for instance did the dissections on the slopes of Puy Ferrand. An intensive rainstorm can produce deep (up to 1.8 m) erosional chutes with systems of steps and kettles which dissect the slope covers down to the bedrock at points of concentrated water flow, such as where a road culvert opens onto a slope. The long and gradual development pattern prevails in areas where the vegetation cover is degraded by, e.g. skiers, snow-cats or bulldozers.

Debris flows occur primarily on the slopes of glacial cirques. Three types of debris flow systems can be identified in the study area (Krzemień 1991), each consisting of:

(a) a hollow, chute and toe;
(b) a long hollow, short chute and a toe dissected by a stream channel;
(c) one or more hollows and a rocky gully that gradually turns into a stream channel.

The presence of the latter two types suggests that the flowing debris does reach the stream channels and can be gradually transported into the mountain foreland. The large number of partly masked and consolidated hollows, chutes and toes suggests that debris flows formerly played a much greater role in the development of the area's morphology. This is corroborated by the existence of large side-wall systems on the consolidated debris toes. Detailed information on the frequency of debris flows in the area is not available, though it is known that the last event of larger scale occurred in the Chaudefour Valley in November 1994.

Anthropogenic forms develop in association with the construction and operation of ski-pistes, slope roads, ski-lifts and cable-cars. Only tarmac roads and underground slope drainage are relatively effective in halting further slope transformation. Intensive linear erosion mainly occurs along road drainage ditches, ski-pistes and ski-lifts (Krzemień 1995, Fig. 10).

5. MORPHOGENETIC ALTITUDINAL ZONES

The structure and functioning of a mountain relief is determined by the sequence of morphodynamic systems, relief types and morphogenetic altitudinal zones. Morphodynamic systems consist of agent-process dependencies, and produce sets of landforms that can be classified as types of mountain relief. Morphogenetic altitudinal zones form a sequence of horizontal belts on a mountain, each spanning a certain vertical distance and featuring a set of morphodynamic systems that produces a defined type of mountain relief (Kaszowski 1984). Morphogenetic altitudinal zoning is proposed by a large number of publications. Working in the Zermatt massif, G. Galibert (1960) identified two high-mountain altitudinal zones: the Alpine and the Pyrenean zones featuring different morphogenetic activity. A. Kotarba and L. Starkel (1972) identified two such zones in the Carpathian Mountains: a cryonival system and a temperate forest system, divided by the timberline. L. Starkel (1980) went on to identify two morphogenetic zones occupying the belt between the timberline and the long-term snow-line in continental climate mountains: a congelifluction process zone (lower) and a debris zone with frost segregation processes (higher). L. Kaszowski (1984) identified four zones in the Hindu Kush Munjan Mountains: the perennial frost, glacial, cryonival and denudational-fluvial zones. He also employed the approach of G. Galibert (1960) in identifying four relief types: the Alpine, Hindu-Kush, Pyrenean and Iranian. In similar vein, M. Chardon (1984) notes patterns in morphogenetic processes and the relief product of the Western Alps by identifying four landscape zones (belts): 1) a glaciated high-Alpine zone; 2) a high-mountain debris and rock zone; 3) a mountain forest and Alpine glade area; and 4) a slope foot and valley bottom zone. Respective sets of morphogenetic processes were associated with these zones. In the context of published research from various high-mountain areas, as defined by C. Troll (1973), it is possible to identify major morphogenetic zones in the Monts Dore massif, and their dominant high-mountain relief features.

There are two morphogenetic zones in the Monts Dore massif: the cryo-nival and denudational-fluvial zones, as divided by the timberline at ca. 1500 m a.s.l. (Figs. 1 and 6).

The cryo-nival zone covers just 12% of the area of the massif and features high-intensity rates of morphogenetic processes. Nivation associated with long-lasting patches of snow is the main process (Varlet 1976, Kaszowski and Krzemień 1989, Kaszowski 1995). The locations and sizes of snow patches depend on their exposure to sun and wind. They are thicker on the leeward slopes and in deep, shady north-facing valleys (Varlet 1976). These are also the places in which nivation plays a greater role. The process is particularly intensive in springtime and in early summer when the snow patches eventually thaw, but also occurs during the frequent mid-winter thaws.

Gelifraction operates due to frequent freeze-thaw action. These processes occur during every thaw in wintertime, but are at their most intense at the end of March and beginning of April. The freeze-thaw cycles also involve the activity of pipkrake (Valadas 1984, 1991).

Aeolian processes also play an important role in the area (Fig. 9). They produce systems of gelideflation hollows that develop in a combination of aeolian and freeze processes (Fig. 8, Kaszowski, Krzemień 1989). High rates of deflation processes are recorded around nivation hollows, roads, tourist trails and ski-pistes (Fig. 4). In summer, deflation finds particularly favourable conditions, due to the fact that the rates of evapotranspiration exceed those of precipitation leaving slope covers dry.

An important role is also played by washdown in areas with a degraded vegetation cover, especially along tourist trails, roads and ski-pistes. Water action develops numerous dissections and deep chutes (Kaszowski, Krzemień 1989, Krzemień 1995).

Gravitational processes such as rock falls and rock failures occur on steep cliffs (Kaszowski, Krzemień 1989). Rock falls are facilitated by wind, especially on windward cliffs (Izmaiłow 1984). As a result of these processes, rock and boulder fields are developing actively at the foot of the cliffs (Kaszowski and Krzemień 1989).

Solifluction can be observed during the mid-winter and springtime thaws, but due to the small scale of the process, its overall role is limited (Valadas 1984).

Human activity has caused extensive damage to the vegetation cover distorting and complicating the boundaries between morphogenetic zones. However it is still possible to distinguish the intensity rates of individual morphogenetic processes in relation to altitude in the Monts Dore massif (Fig. 11).

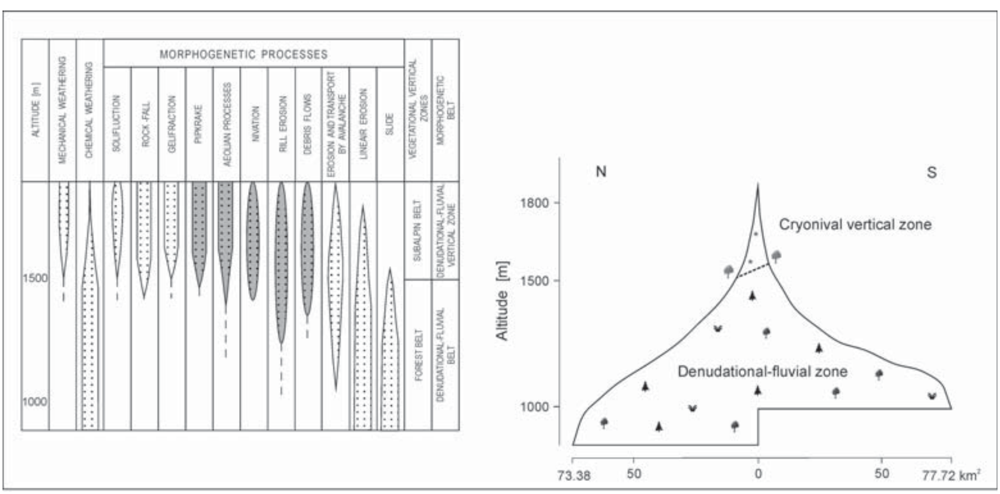

Figure 11. Vertical differences between intensity rates of morphogenetic processes and morphogenetic zones in the Monts Dore massif compared to the background of relative hypsographic curves of the N- and S-facing slopes.

Denudational-fluvial zone. This zone occupies the remaining 88% of the study area and normally features low rates of morphogenetic processes, except of debris flows that are activated every few decades. In average conditions the slopes remain in equilibrium (Valadas 1984, 1991). Due to high concentrations of organically-derived carbon dioxide, intensive chemical denudation occurs in the forest zone (Kotarba *et al.* 1987), at a rate proportional to the volume of water reacting with the rock (Kotarba 1972). In the Monts Dore massif this rate is at its highest in winter and spring, during the frequent thaws and falls of rain. The dominant processes of the fluvial-denudational zone are linear erosion and slow mass movements. Sliding is most likely to occur on heterogenic slopes, especially where volcanic rocks lie on top of pyroclastic formations (Valadas 1984). Allochthonous processes, such as avalanches and debris flows, occur in this zone, primarily in the Chaudefour Valley. Debris flows originate in one of three morphodynamic zones: i) a zone with one or more hollows; ii) a zone where water-saturated material is transported and where debris flows grow in volume and erosional power; and iii) a deposition zone. All three combine to form a morphodynamic system spanning the ridge zone and the forest zone (Krzemień 1991).

6. CONCLUSION

The topography of the Monts Dore slopes reveals both several generations of old and fully consolidated forms and some very fresh landforms, such as debris-flow hollows, chutes and toes. Major groups of landforms observed in the area have emerged at various points throughout the Holocene, and are continuing to develop up to the present day. Certain processes involved in this development have gained in morphodynamic importance due to human activity.

In the topmost morphodynamic zone of the Monts Dore massif nivation processes associated with long-lasting patches of snow are the most important. They are followed in order of importance by cryogenic processes, with a notable role for needle-ice, and by deflation. Finally, there is also a material role of runoff, taking the form of either rill-flow or debris flow, whose natural potential has increased recently as a result of a deterioration of the vegetation cover.

Geomorphological maps of the highest parts of the Monts Dore massif generated during fieldwork suggest that the most effective phase in its morphogenesis occurred in periglacial conditions, towards the end of the last Ice Age. The massif was shaped by nivation and runoff throughout the Holocene, but the last 30 years have altered this picture. The added human activity and accelerated rates of erosion and deposition have likened this period—in terms of the scale of transformation—to the periglacial morphogenesis.

Nival hollows of various sizes and generations have been found to be the hallmark dominating the relief of the area (Fig. 3). This suggests a long history during which this landform assumed sustained importance in the overall relief. Consequently, the Monts Dore massif might be seen as a typical high-mountain area with a nivation type relief dominant above the timberline, or simply a Monts Dore type of relief. It would complement such high-mountain relief typologies as those proposed by G. Galibert, M. Chardon, or L. Kaszowski. The "Monts Dore" type of relief is also found in other ranges of the Massif Central, e.g. Cantal and Mezanc, as well as in the Scottish Cairngorm mountains.

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FORTY YEARS OF ROCKFALL ACCUMULATION AT THE MOUNT WILCOX SITE, JASPER NATIONAL PARK, ALBERTA, CANADA

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Abstract: Rockfall accumulation was measured in 1981 and 2000 along ca. 850 m of the former Banff-Jasper highway that skirts a large talus cone from Mount Wilcox in Jasper National Park, Alberta. Approximately 34.5 m^3 of rockfall debris has accumulated on the roadbed since it was abandoned in 1961. The largest boulder was 4.65 m^3 and 5 others were >0.95 m^3. Estimated minimum accumulation rates from 20–30 m sections of highway adjacent to the base of the cone are mainly between 0.1–0.3 mm yr^{-1}, with rates >0.4 mm yr^{-1} reflecting the presence of individual large boulders. These rates correspond to a thickness of 1–5 m over the Holocene and may be reasonable estimates for deposition at the outer talus fringe of such large cones. Over the observation period, approximately 10 rockfalls > 0.25 m "a" axis were deposited on the road each year, but only one every two years was > 0.5 m. Much larger rockfalls (up to 6–10 m^3) have traveled across the road, creating bumpholes, and been deposited on the outwash. These results indicate some of the problems with trying to estimate rates of contemporary rockfall activity from sampling programmes based on relatively short periods of time and limited depositional areas because of the magnitude frequency distribution of the rockfall events.

Key words: rockfall accumulation, lichenometry, the Canadian Rockies

INTRODUCTION

In his early career Professor Kotarba pioneered studies of the magnitude and frequency of mass wasting and other geomorphic processes in the high mountain areas of his beloved Tatra Mountains (e.g. Kotarba *et al.* 1979). I enjoyed visiting some of these wonderful sites with him in conjunction with meetings of the IGU Commission on Field Experiments in Geomorphology in 1979, and we have also explored other mountain areas together in Scandinavia, Karkevagge and the Canadian Rockies. At that time I was also investigating the geomorphic activity, magnitude and frequency of rockfalls, debris flows and snow avalanches. I offer this little study to Adam in celebration of his career and on the happy occasion of his seventieth birthday.

THE MEASUREMENT OF ROCKFALL PROCESSES

The problem with measuring rockfall processes directly has always been one of sampling in time (for an adequate length of record) and over space (to obtain a reasonable sample of events) sufficient to allow the process to be characterized. Recent advances in dendrochronology offer new opportunities to study these processes indirectly (e.g. Stoffel and Perret 2006), but that will not be the main focus here. Initially work was carried out on the talus slopes themselves by means of simple, short term inventories (e.g. Luckman 1976; Gardner 1980) and studies of accumulation patterns on small mats, plastic sheets and/or boulders (see Luckman 1978, 1988 and references therein). The problem with using these techniques for rockfalls was that the rockfall events were too infrequent or of insufficient number to allow reasonable estimates of seasonal, long-term or spatial patterns of activity to be arrived at.

One attempt to address the above problem was to select sites at which a large reference surface could be identified, allowing for the isolation and recognition of recent rockfall events that could then be measured and dated to provide long-term rates of activity. An initial study of this type by Gray (1973) placed large areas of fish netting over talus slopes in the Yukon, but unfortunately did not give rise to any publication of results in full. An alternative approach to this problem was to use lichenometry to estimate long–term accumulation rates (André 1986; Luckman and Fiske 1995; McCarroll *et al.* 1998). In the Canadian Rockies talus slopes below high quartzite walls are often heavily colonized by *Rhizocarpon geographicum*, and these lichen-covered surfaces provide the necessary reference surface on which to identify new, relatively lichen-free boulders and also to date these recent additions to the surface. This has allowed for estimates of rockfall rates at these sites for periods of several hundred years. However, the rates

Figure 1. The Sunwapta outwash, toe of the Mount Wilcox Cone (left) and the present (far right) and former (left) highways, September 13th, 2000. The roadbed and outwash are covered with mountain dryas which is in seed. The view is eastwards towards Mount Athabasca and shows sections 5 (foreground, left) to 14 of the sampled road surface.

of debris accumulation were extremely low (ca. 1 mm/century) and clearly indicated that the slopes under study were largely fossil and, possibly, paraglacial forms, probably dating from the end of the last glaciation (Luckman and Fiske 1997). Unfortunately this approach cannot be adopted for many other sites in the Rockies because of geological constraints: most talus slopes in these mountains are developed from calcareous sedimentary rocks, while *Rhizocarpon* only colonizes siliceous substrates.

An alternative approach would be to locate a well-dated, stable reference surface on which new material could be identified clearly, and hence volumetric determinations of accumulation over time could be made. Situations such as the outwash plain of the Sunwapta River, downstream of the Athabasca Glacier could be ideal, as a large talus slope runs out onto the outwash plain. Many large boulders are clearly visible on the surface and have obviously resulted from contemporary rockfall activity (Fig. 1). Unfortunately, the age of this surface is unknown and diachronous due to channel migration across it during and subsequent to the Little Ice Age. However, one portion of this outwash plain does contain an abandoned surface of known age that can be used to estimate past rates of rockfall activity.

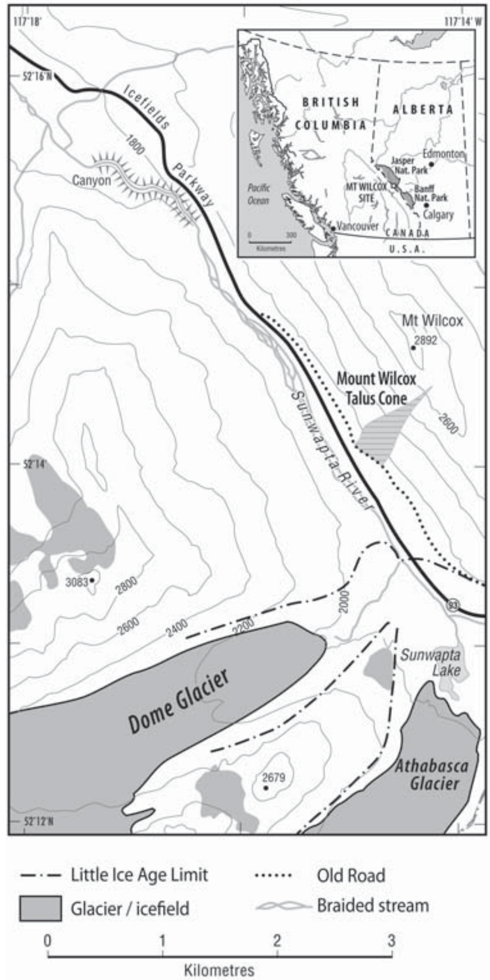

Figure 2. Location of the Mount Wilcox site.

THE MOUNT WILCOX STUDY SITE (FIG. 2)

The highway between Banff and Jasper was originally a gravel road constructed as a Depression Relief Project starting in 1931. The road was officially opened on June 15th 1940. As the road approached the Columbia Icefield and Athabasca Glacier from the north, it skirted the edge of the Sunwapta outwash before ascending gradually to cut through the forest on the lower slopes of Mount Wilcox and avoid the Little Ice Age terminal moraine complex of the Athabasca Glacier (Fig. 2). The northeastern margin of the outwash was flanked by major cliffs of Mount Wilcox that tower above the road and feed a number of talus cones, some minor rock-glacierized talus and avalanche tracks (Fig. 3, inset lower left). In the late 1950s and early 1960s, the old road was surfaced, widened and, in some places, realigned. The former roadbed below Mount Wilcox was abandoned and has slowly revegetated, accumulating rock-fall material after the new road opened in 1961.

North of the Athabasca Forefield the rerouted Icefields Parkway cuts across the Sunwapta outwash, the river having been redirected into a new straight channel south of the road. The former road followed the edge of the outwash, either flush with or raised a small distance above the former surface, and skirted the front of a large talus cone from the south-west facing slopes of Mount Wilcox. This cone heads in a series of gullies from the ridgecrest at 2700–2800 m (Fig. 3), descending onto the former outwash surface at ca. 1920 m. The cliff is formed of calcareous and dolomitic siltstones, sandstones and shales, interbedded with limestones of lower Ordovician and Cambrian age which dip gently back into the grading to the outwash at its base (Fig. 4). The old road on the outwash (Fig. 3) runs initially about 50–20 m beyond the base of a series of smaller talus cones, but then skirts the foot of the main cone for over 500 m. The main sampling area was along the toe of this cone, where it forms a sharp junction with the outwash. In some cases a short fringe of shrubby willows and small trees has colonized the lowest part of the cone.

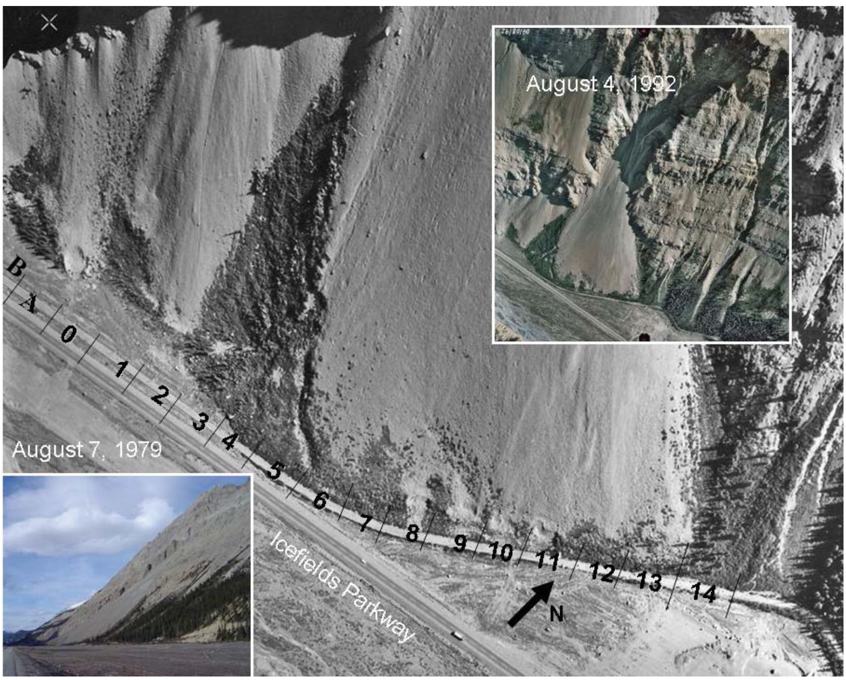

Figure 3. The Mount Wilcox Cone showing its local setting and the sampled section of the old Banff-Jasper Highway. The 1979 and 1992 aerial photographs are reproduced by permission of the National Air Photo Library of Natural Resources Canada, Ottawa. The photograph lower left shows the cone and highway from the east (photograph by Ben Gadd, Jasper, October 2007).

slope, forming an alternating series of cliffs and benches. The main cone begins at about 2300–2400 m, morphological considerations suggesting that this is basically a rockfall talus (see Luckman 2007) with minor amounts of snow-avalanche and debris-flow activity. It has a relatively simple, straight long profile (Fig. 3, inset), with a basal concavity

RESEARCH METHODS

The initial survey of accumulation was carried out over ca. 660 m of the former roadbed by the author and Gordon Frazer on July 6th, 1982. For most of this length the highway roadbed was between seven and eight metres wide. Abandoned telegraph poles,

Figure 4. The largest boulder on the former highway, estimated volume 4.65 m^3 in Section 11. View to the west, September 14th, 2000. The basal slope of the talus cone is visible to the right of the person.

spaced approximately 50 m apart, were still standing at this time and provided convenient reference points to subdivide the road into sections for measurement. The first few sampled areas (segments B to 4, Figure 3, Table 1) contained little new debris and were measured in ca. 50 m units. Sections 5–14 contained considerably more debris and were subdivided into two parts, the initial 30 m (the length of the measuring tape!) and the distance to the next pole (usually ca 20 m). Triaxial measurements were made with a 60 cm wooden caliper previously used by Luckman (1988). Essentially most boulders with an "a" (long) axis of > ca.30 cm were measured, while smaller volumes were estimated visually.

The measurements were repeated by the author and Helen Luckman on September 13th and 14th, 2000, along approximately 860 m of roadbed (Fig. 4), using a similar procedure but with an extendable wooden ruler to measure the boulders. In this inventory, measurements were made in 10 m increments between the stumps of the former telegraph poles. The measurement cutoff was somewhat smaller, generally about 0.2 m a axis. All axial measurements were made by the author. Boulder volume was determined from the product of the three axes and an empirically derived "shape" factor (see Luckman 1988)[1]. Only those boulders that had clearly come to rest on or against the former road surface were measured and included in the inventory, though occasionally large boulders on adjacent gravel pit surfaces and access roads between the old and new roads were also measured, as these had obviously arrived after the surfaces were abandoned ca. 1960. Volumetric estimates were made for all boulders sampled on both surveys to determine accumulation over the 1961–1982 and 1961–2000 intervals. Volumes of accumulation for the 1981–2000 interval were determined by differencing these two values.

[1] Four major shape classes were involved with values of 0.41, 0.51 (the most common), 0.61 and 0.76

Table 1. Rockfall accumulation data for the Mount Wilcox site.

			Volume (m^3)			Boulders[1]		Accumulation rate (mm yr^{-1})						
											N boulders[2]			R.I[3]
Section	Length m	Area m^2	1961–1982	1982–2000	1961–2000	largest %	top 5 %	1961–1982	1982–2000	1961–2000	>1 m	>0.5 m	all	years
B	51.0	357			0.000					0.000	0	0		
A	53.0	411			0.283	57.8	99.9			0.017	0	2	6	35
0	50.0	375			0.000					0.000	0	0		0
1	49.0	367			0.843	90.8	100.0			0.057	1	2	3	65
2	52.0	416	0.013	0.001	0.014	92.7	100.0	0.001	0.000	0.001	0	0	3	69
3	50.0	375	0.066	0.054	0.121	37.0	99.8	0.008	0.008	0.008	0	2	6	33
4	56.0	225	0.908	0.173	1.081	16.0	48.0	0.183	0.040	0.120	3	19	61	3.7
5	30.0	255	0.890	0.465	1.355	9.0	36.0	0.159	0.096	0.133		25	51	2.4
5.3	18.5	148	0.390	0.228	0.618	16.7	46.0	0.120	0.081	0.104		14	28	2.6
6	30.0	240	1.008	0.297	1.305	10.0	32.0	0.191	0.065	0.136		24	61	2
6.3	28.0	224	0.337	0.329	0.666	6.1	35.0	0.068	0.077	0.074		17	31	3.6
7	30.0	247	0.850	0.544	1.393	16.8	40.0	0.156	0.116	0.141	2	34	62	1.9
7.3	15.0	120	0.497	0.020	0.517	21.4	52.0	0.188	0.009	0.108	1	15	29	2.1
8	30.0	240	1.989	0.082	2.071	20.8	43.0	0.377	0.018	0.216	3	32	59	2
8.3	20.0	160	0.691	0.367	1.058	13.6	44.0	0.196	0.121	0.165		19	43	1.9
9	30.0	240	0.664	0.546	1.211	12.6	40.0	0.126	0.120	0.126		27	30	4
9.3	12.3	98	0.437	0.279	0.716	16.5	43.0	0.203	0.150	0.183		16	40	1.2
10	30.0	240	0.552	0.809	1.361	7.2	23.0	0.105	0.177	0.142		30	112	1.1
10.3	21.0	168	0.564	0.603	1.168	11.1	30.0	0.153	0.189	0.174	1	21	76	1.1
11	30.0	240	5.549	0.542	6.090	76.3	81.0	1.051	0.119	0.634		31	76	1.6
11.3	22.0	154	1.472	0.312	1.784	49.5	76.0	0.435	0.107	0.290	1	16	35	2.5
12	30.0	210	0.352	0.583	0.935	18.0	53.0	0.076	0.146	0.111		19	49	2.4
12.3	23.5	164	1.474	2.199	3.672	39.5	93.0	0.408	0.706	0.560	3	10	24	3.9
13	48.0	350	0.204	0.140	0.345	24.7	63.0	0.027	0.021	0.025	2	6	20	9.6
14	30.0	180	1.947	1.027	2.974	52.0	96.0	0.492	0.300	0.413	0	7	8	15
14.3	22.0	121	1.640	1.599	3.240	41.8	80.0	0.616	0.696	0.669	3	14	42	2.1
TOTAL	861.0	6,522	22.497	12.042	34.539						20	402	955	
MEAN								0.157	0.097	0.132				

Notes Accumulation on sections B- through 1 was only measured in 2000

[1] percentage of the total volume contributed by the largest or largest 5 boulders in the section

[2] number of boulders on each section > 1 m, >0.5 m, and ca. >0.22 m “a” axis length in 2000

[3] R.I. number of years between rockfalls >0.22 m “a” axis length on this section calculated as the inverse of the number of boulders >0.22 m “a” axis length per year per 10 m of section

The boulders from the first survey were obviously re-measured during the second survey. As the road segments often contained a few larger boulders and many smaller ones, the total volume of accumulation for individual segments is frequently dominated by the volumetric estimates for one or two boulders. These "largest" boulders may comprise 10–95% of the total estimated volume of accumulation for an individual transect over a given period (see Results and Table 1). As these boulders are irregular in shape (Figs. 4 and 5), triaxial measurements are approximate, and replicate estimates of the volume of the same boulder 18 years apart, can show significant differences. To eliminate possible discrepancies from this source, measurements for the largest boulders from each section of road in the initial survey were matched against the dimensions of the largest boulders from the same section in the second survey and similar pairs of boulders identified. Where volumetric estimates for these pairs differed by >ca. 10%, the two estimates were averaged and this volume was used in calculating the volume of accumulation for both the 1961–1982 and 1961–2000 time periods.

RESULTS

Table 1 summarizes the results broken down by the individual sections of roadbed identified on Fig. 2. Results for sections 4–14 are reported in two parts, the first 30 m and the residual distance to the next "pole" (usually ca 20 m). The estimated volumes of rockfall accumulation are given for the 1961–1982 and 1961–2000 periods and interpolated 1982–2000 values. Accumulation rates are also presented as the deposition for each segment for all three time periods, calculated as a thickness (depth) of accumulation in mm yr^{-1}, averaged over the segment. Details are also given of the number of boulders

Figure 5. Fresh "Bumphole" and debris on the road surface. The view is eastwards from section 6 towards section 14 (large boulders in the distance). Note the heavy dryas cover that has colonized the former road surface.

measured on each segment, categorized by the length of the longest "a" axis. The proportion of the total volume of accumulation on each segment represented by the single largest and five largest boulders measured on that section is also listed. Details of the largest boulders and boulder numbers are only presented for the 2000 data.

Over the 40-year period, an estimated volume of 34.5 m^3 accumulated on the road surface, approximately 22.5 m^3 between 1961 and 1982 and an additional 12 m^3 between 1982 and 2000. These volumes correspond to rates of 0.157 and 0.097 mm yr^{-1} for the two periods, and an average of 0.132 mm yr^{-1} for the total period. Approximately 450 boulders were measured in 1982 and over 1000 in 2000[2]. The apparent difference in rates between these two intervals is somewhat misleading, because of the influence of individual boulders on these totals. The largest boulder measured in 1982 (also the largest in 2000) was of approximately 4.65 m^3, and the cumulative volume of the five largest boulders measured during the 1982 survey is approximately 9.48 m^3. Ignoring these five boulders, the volume of accumulation over the two intervals is very similar (ca. 13 and 12 m^3, respectively).

Table 1 shows the measurements broken down into individual segments along the transect, allowing for some discussion of spatial variability. From section B to section 3 (Fig. 3) the road is raised slightly above the outwash and between 20–50 m from the base of the smaller talus cones below the low cliff north of the main cone. This part of the former road has few boulders, and accumulation rates are of ca. 0.01–0.02 mm yr^{-1}—except where a single boulder is present in section A. Sections 4 and 5 are on the margins of the main cone, but separated from the open talus surface by a narrow belt of shrubs and a few small trees. Accumulation rates here range from ca. 0.1 to 1.13 mm yr^{-1}. This area also has limited snow avalanche deposition. Broken tree branches were noted in 1982, but many fresh rock chips and a veneer of small "avalanche" or "perched" boulders (Luckman 1988) were seen in 2000, though volumes were too small to contribute measurably to the volumetric estimates. In addition, comparison of the 1979 and 1992 air photographs shows 4 or 5 fresh debris flow channels on the later image that extend from the cliff, down the northern flank of the talus cone, terminating in the trees above section 4, but not extending onto the road.

From sites 6–13 the road is a short distance from the base of the talus (Fig. 3) and in some places is cut slightly into the toe of the slope. The base of the talus also has small shallow pits from which material was quarried for the road between sections 8 and 11. Deposition on these sections of roadway averages between 0.1 and 0.3 mm yr^{-1}, except where larger boulders "inflate" these values (sections 11a and 12b). Just over half of these sections show greater accumulation values in the earlier record, mainly as a result of individual large boulders. Section 14 is at the edge of the talus cone, and has a number of larger boulders with average deposition of ca. 0.5 mm yr^{-1} (Fig. 6). The sites lying at the foot of the talus cone (sections 4–14) have mean accumulation rates of 0.253 mm yr^{-1} and 0.146 for the 1961–1982 and 1982–2000 periods, respectively, and an average of 0.209 mm yr^{-1} over the 40-year period.

Although large numbers of boulders were measured, the mean volume of accumulation recorded for each section or time interval is significantly influenced in many cases by the volume of the largest boulders. Over the 40-year period, 50% of the total measured accumulation is accounted for by the 29 largest boulders (less than 3% of those measured), whereas the smallest 50% of boulders measured only comprise about 7.5% of the total volume of accumulation. Table 1 and Figure 7 show the percentage of the total volume accounted for by the single largest and five largest boulders in each section for the 2000 survey. Below the main cone the largest boulder in each section accounts for 10–20% of the total accumulation, with the largest 5

[2] These numbers include some smaller boulders not included in the calculations in Table 1

usually accounting for 30–50% of the total, except where the largest boulders are found (between 11a and 14).

10 m length of road (Fig. 7) Along the foot of the main talus slope (sections 4–14) these results are relatively uniform, with most 10 m

Figure 6. Large boulders and a "bumphole" at the distal end of section 14. Beyond section 14 the road is cut into the toe of a vegetated talus and has little fresh debris.

equivalent sections averaging between 6 and 13 events over the 40-year period. Additional estimates of rockfall frequency over this section were derived from the total number of measured boulders with a axes greater than or equal to ca 0.22 m (Table 1), and expressed as recurrence intervals. Based on the total number of boulders measured in each section plus standardization by section length, the sample areas between sites 4 and 14 had recurrence intervals of between 1.1 and 15 years, i.e. they would receive rockfalls >0.22 m once every 1–15 years for each 10 m section. Sections 4–12 have a higher frequency of once every 2.33 years.

A better estimate of relative rates of activity can be achieved by examining the number of boulders of a given size range, rather than the volume of accumulation. The sedimentary rocks comprising the talus material are relatively thinly bedded and tend to be platy or elongate in form. Average axial ratios for the measured boulders >0.5 m a axis are 1.5 (a/b), 3.2 (a/c) and 2.2 (b/c). Over the 40 year period only six of the boulders measured on the road had estimated volumes >0.98 m^3, and twenty had long "a" axis lengths of >1.0 m. As these numbers were too few for a meaningful analysis, the population of 402 boulders with a axes >0.47 m were used as indicators of the number of rockfalls (Table 1). To account for differences in the length of the individual sections sampled, these numbers were converted to the mean number of rockfalls per

These figures are obviously only minimum estimates for rockfall activity, as they simply record those rockfall boulders that have come to rest on the former road surface over the last 40 years. Many other boulders

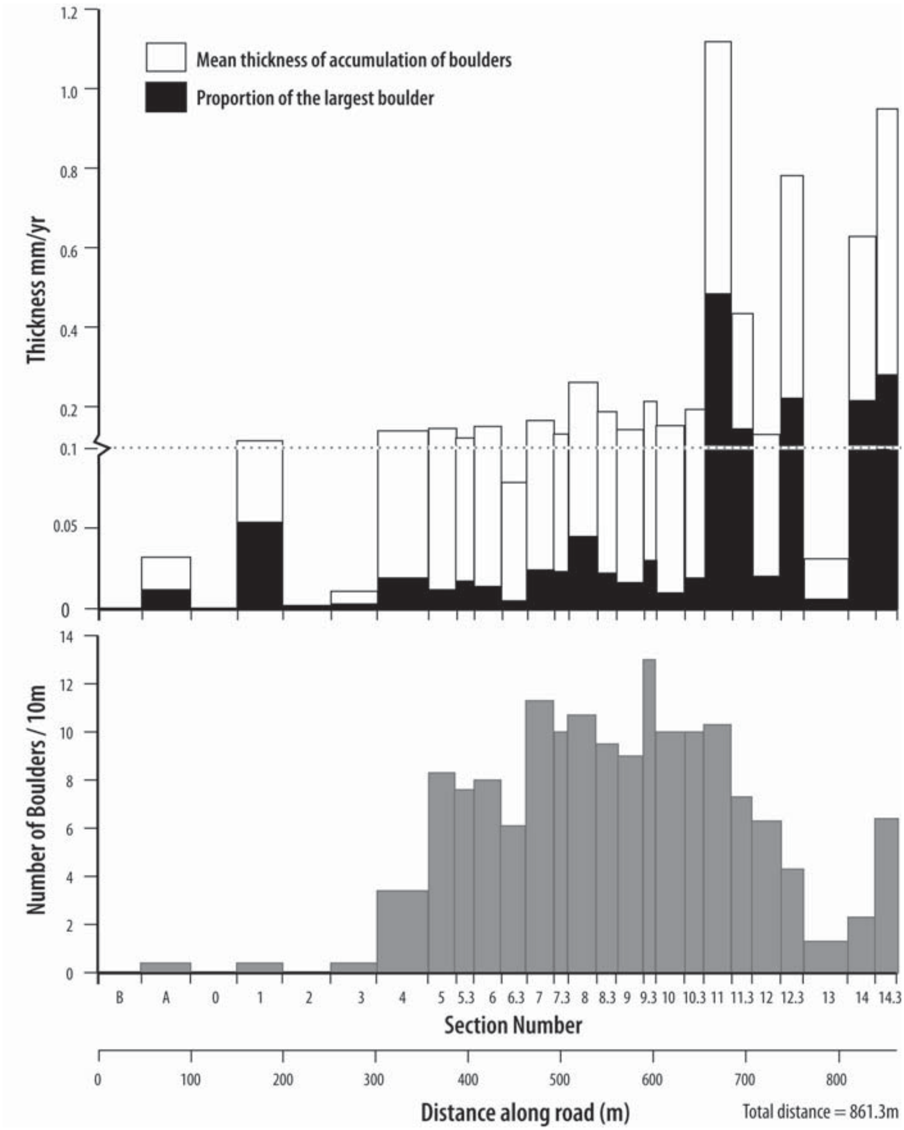

Figure 7. (UPPER) Rockfall accumulation rates on the former highway.
The accumulation rates are derived for each segment by dividing the volume of accumulation (1961–2000) by the area of roadbed sampled and are expressed as mm "thickness" per year (divided by 40). The column widths are proportional to segment widths. The open bars are the mean accumulation on each segment and the black bars represent the proportion of the total contributed by the largest boulder on that segment. (LOWER) Relative numbers of rockfalls on each sampled segment. The " relative number" of rockfalls is defined as the number of boulders >ca. 0.22 m a axis measured on each segment over the 1962–2000 period per 10 m unit of segment width. The column widths are proportional to the varying segment widths.

have clearly travelled across the road and been deposited on the outwash between the old and new roads (Fig. 1). There are three large bumpholes in the road surface, the largest of which is approximately 5 m x 2.5 m wide and up to 0.75 m deep at its deepest point, (Fig. 6). One smaller bumphole can be traced to a boulder of ca. 4.9 m^3 about 30 m beyond the road. Although no systematic search was carried out, the largest boulder measured on the outwash surface was of ca. 10.8 m^3; about 70 m from the road, while several other boulders in the 3–6 m^3 range can be seen on access roads, former gravel pit surfaces or the abandoned bed of the Sunwapta River.

COMPARISON WITH OTHER ESTIMATES

Two other sets of direct estimates of rockfall accumulation are available from the Canadian Rockies. Estimates of debris accumulation on squares and boulders by Luckman (1988) in Surprise Valley involved both snow-avalanche and rockfall dominated sites. Accumulation rates at the base of the rockfall dominated Scree No 1, Lower Palliser and Entrance sites averaged 0.29, 0.51 and 0.1 mm yr^{-1}, respectively, over a 13-year period. However, like the results reported here, these figures are dominated by the volumes of a few large boulders averaged over a relatively short period of time. In most cases these represented only 3–4 events with rates of approximately one event every 3 or 4 years. Nevertheless, given these limitations, the results seem broadly comparable between the two studies, despite the much smaller sampled areas for the sites in Surprise. The study by Luckman and Fiske (1995) on lichen-covered quartzite sites reported rates averaging 1.1 mm/century (range 0.25–1.77) over an estimated 300-year period from sampled areas of 25–200 m^2. Rockfall deposition at the base of the talus was estimated to have a recurrence interval of once every 15–100 years per 100 m^2 area. These rates are orders of magnitude lower than the Surprise or Mount Wilcox data, indicating that these quartzite talus slopes are relatively inactive at present (Luckman and Fiske 1997). It would appear that obtaining reasonable estimates of contemporary rockfall activity (volumes) remains an intractable problem, because of the magnitude frequency spectrum of these events. Mean values tend to be skewed by the largest events (often single boulders) and frequency observations would seem to provide the best estimates. The rates reported for the Mount Wilcox site are probably reasonable for a large talus in this environment. It is clear however that several boulders in the range of 3–5 m^3 arrived at the talus foot in a 30–40 year period, but came to rest on the outwash beyond the road. With hindsight it would have been interesting to map and determine the size of the largest boulders on the outwash at this site, just to provide a size distribution for these large events. Ultimately the best estimates of long-term accumulation amounts could be derived from volumetric measurements of the talus itself, using techniques such as ground-penetrating radar (e.g. Sass and Wollny 2001). However, such measurements only provide mean rates over the Holocene period and are inadequate for determining contemporary rates of activity in the absence of recognizable dated horizons with the talus itself. Accumulation on abandoned and dated outwash surfaces such as the Sunwapta outwash provide the best potential to estimate rates on decade-century timescales.

CONCLUSIONS

Approximately 34.5 m^3 of rockfall debris accumulated along ca. 550 m of the former Banff-Jasper Highway between 1961 and 2000, at the toe of the Mount Wilcox Cone (sites 4–14, Fig. 3). This corresponds to minimum rockfall accumulation rates that vary between ca. 0.1 and 0.55 mm yr^{-1} over the last 40 years at or immediately beyond the base of this active talus cone. The distribution of deposition around the base of the cone appears relatively uniform at ca. 0.1–0.3 mm yr^{-1}, with rates >0.4 mm reflecting the presence of individual large boulders on the 20–30 m long sampling sections. These rates correspond to a thickness of 1–5 m over the Holocene (last 10,000 years) and may be reasonable for the outer talus fringe of such large cones. Over the observation period, approximately 10 rockfalls > 0.25 m "a" axis were deposited on the road each year, but only one every two years was > 0.5 m a axis. Much larger rockfall boulders of up to 6–10 m^3 have traveled across the road, creating bumpholes, and been deposited on the outwash. However the majority of these events cannot be dated and probably reflect deposition over a period of several hundred years. These data indicate that extrapolation of long-term accumulation rates from short-term or small-area sampling programmes may be biased by the inclusion the few large events recorded over the observation period. Even with 40 years of accumulation on an area of >6000 m^2, 13% of the total accumulation was a single boulder, and half of the accumulated volume was accounted for by less than 3% of the thousand boulders measured in 2000. These results caution against estimating contemporary rates of activity on the basis of casual observation or short-period monitoring.

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PRESENT-DAY RETREAT OF SLOPES ABOVE THE UPPER TIMBERLINE IN THE SLOVAK PART OF THE WESTERN CARPATHIANS

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Abstract: This article focuses on the rate of present-day retreat of slopes, as determined by direct measurement, both above and below the upper timberline, and in relation to erosion, mass movements, cryogenic processes, nivation, aeolian processes, etc.
Rates of surface erosion processes—the most widespread geomorphic processes operating at high altitude in the Western Carpathians—range from low to high. In the forest belt below the upper timberline, in the dwarf pine scrub and on surfaces with grassland, the mean rate of soil removal is of only 0.001–0.007 mm yr^{-1}. In the case of denuded surfaces with destroyed cover this rises to 3.4 mm yr^{-1} on average, while linear erosional landforms lose as much as 17.4 mm yr^{-1}.
As a whole, the area above the upper timberline in the Slovak geomorphic units of the Western Carpathians is characterised by medium to relatively high real denudation rates, the mean rate of slope retreat being 0.27 mm yr^{-1}.

Key words: present-day geomorphic processes, upper timberline, rate of slope retreat, Slovakia, Western Carpathians

INTRODUCTION

The data needed to evaluate present-day geomorphic processes participating in the morphogenesis of high mountains are lacking, since they must mainly be obtained by way of direct and long-term measurement on stationary plots. The lack of such data is in fact perceptible worldwide, since the measurement in question involves strenuous efforts made under difficult geomorphic and climatic conditions. Many of the data on rates of morphogenetic processes ongoing in high mountains that are present in geomorphological papers have in fact been deduced from indirect measurements, estimations or observations (e.g. Lukniš 1973).

The data upon which to base evaluations of present-day geomorphic processes in the high-altitude parts of the Western Carpathians were obtained—mostly by direct measurement—by both Slovak geomorphologists (Midriak 1983, 1993, 1996 etc.) and their Polish counterparts (Gerlach 1964, etc.; Kotarba 1970, 1971, 1972, 1976, 1992, 1997, 2005, etc.; Dobija 1973; Kłapa 1980; Kotarba *et al.* 1983, 1987; Rączkowska 1995; and others).

STUDY AREA

The high-altitude areas of Slovakia are taken to be those at least 1500 m a.s.l., in which a natural upper timberline is therefore present (Lukniš and Plesník 1961; Kotarba and Starkel 1972). Isolated areas of this kind within the Slovak part of the Western Carpathians are found in the Eastern Tatras (Vysoké Tatry and the Belianske Tatry), the Western Tatras (Západné Tatry), the Low Tatras (Nízke Tatry, i.e. Ďumbierske Tatry and Kráľovohoľské Tatry), the Mala Fatra Mts. (Krivánska Fatra and Lúčanská Fatra), the Veľká Fatra Mts., the Chočské vrchy Mts. (though only at Mt. Veľký Choč) and the Oravské Beskydy Mts. (Babia Hora and Pilsko). Geoecological characteristics of these high-mountain areas in Slovakia are supplied in Table 1, the data in part being adapted from Midriak (1983).

Research on present-day geomorphic processes involving the use of direct methods of measurement was done in all the high-mountain areas of Slovakia except the Chočské Vrchy. The results are here compared with those obtained for the sub-montane belt (i.e. the area below the upper timberline).

TYPOLOGICAL DIVISION OF THE RELIEF IN THE HIGH MOUNTAINS OF SLOVAKIA

The cold climate morphogenetic area in the Slovak part of the Western Carpathians (only above the upper timberline) occupies about 1% of the total area of Slovakia, or some 48,900 ha. Across the entire area occupied by different types of relief, Mazúr (1980) derived the following typological division for Slovakia:

- 66% of the surface corresponds to fluvial dissected relief,
- 13% of the surface corresponds to high-mountain gentle relief (with grassland),
- 14% of the surface corresponds to mixed high-mountain sharp and high-mountain gentle relief (with grassland)
- 7% of the surface corresponds to high-mountain sharp relief.

Accumulation relief is nearly absent above the natural upper timberline, so the erosional-denudational relief is widespread in the cryonival belt. According to Plesník (1971) and Midriak (1983, 2005), the upper timberline has been lowered anthropogenically by an average of 100–265 m. The present-day timberline thus only reaches to 1185–1438 m a.s.l., though it may still be present above 1700 m. Typical relief above the natural upper timberline is high-mountain gentle relief with grassland, as well as high-mountain sharp relief. However, lowering of the natural upper timberline led to deforestation over an area twice as large as it once naturally was (when confined to the upper timberline and enlarged cryonival belt). The fluvial dissected relief (predominantly in the subalpine belt) is the most widespread in the altitudinal zone of the enlarged cryonival system.

METHODS AND INTERPRETATION

The work described in this paper utilised many methods to determine both potential and real retreat of slopes.

The water erosion threat posed at high altitudes in the Western Carpathians was evaluated using the Frewert method, as modified by Stehlík (1970) and the author. In the calculation of potential soil (earth) losses by surface runoff it was assumed that running water together with gravitation are the most important erosive factors on the discussed high-mountain slopes, as elsewhere.

Results of indirect measurement or estimations of surface retreat (lowering of the surface) in the highest part of the Western Carpathians (the Slovak part of the Tatra Mts.) are as published by Lukniš (1973).

The direct methods of measurement of the present-day retreat of slopes at high altitude in the Western Carpathians included micro-nivelation and translocation, as well as volumetric, deluometric, pedologic, photogrammetric and other methods. Detailed descriptions of these are to be found in the monographs and papers by Gerlach (1964),

Table 1. Geoecological characteristics of high mountains in the Western Carpathians of Slovakia.

Geomorphological unit -or part thereof	Max. altitude (m a.s.l.)	Altitude differences between summits and basin bottoms (m)	Mean altitude of present timberline (PT) including "struggle" zone (m a.s.l.)	Length of existing upper timberline (km)	Prevailing type of landform (geoecological belt)
Tatry	2655	1700–2000	1482	364	sharp and gentle high-mountain relief (alpine + subalpine)
-Vysoké Tatry	2655	2000	1487 (PT 1392-1438)[1]	161	sharp high-mountain relief (alpine + subnival)
-Belianske Tatry	2152	1400	1485 (PT 1385)[1]	27	gentle high-mountain relief (alpine)
-Západné Tatry	2248	1500	1478 (PT 1380)[2]	176	gentle high-mountain relief (alpine)
Nízke Tatry	2043	1400–1500	1423	402	gentle high-mountain relief (alpine)
-Ďumbierske Tatry	2043	1400–1500	1417	275	gentle high-mountain relief (alpine)
-Kráľovohoľské Tatry	1948	1200–1300	1431	127	gentle high-mountain relief (alpine)
Malá Fatra	1709	1300	1253	129	fluvial relief (subalpine)
-Krivánska Fatra	1709	1300	1238 (PT 1185)[3]	95	fluvial relief (subalpine)
-Lúčanská Fatra	1476	1000	1298	34	fluvial relief (subalpine)
Veľká Fatra	1592	1100	1252	108	fluvial relief (subalpine)
Chočské vrchy	1611	1100	—	—	fluvial relief (subalpine)
-Veľký Choč	1611	1100	1460 (PT 1274)[4]	3	fluvial relief (subalpine)
Oravské Beskydy	1725	1000	1403	15	fluvial relief (subalpine)
-Babia hora	1725	1000	1400	7	fluvial relief (subalpine)
-Pilsko	1557	800	1406	8	fluvial relief (subalpine)

Note: [1]Plesník (1971), [2]Myczkowski, Plesník (1974), [3]Plesník (1958), [4]Plesník (1966)

Kotarba (1970, 1972, 1976), Zachar (1982), Midriak (1983) and others.

RETREAT OF SLOPES AT HIGH ALTITUDE IN THE SLOVAK PART OF THE WESTERN CARPATHIANS—THREAT AND RATE OF REAL LOWERING

POTENTIAL RETREAT

As regards the threat of potential erosion on slopes posed by water running on a slope surface, it is possible to refer to a predominance of very strong erosion (according to the classification by Šály and Midriak 1995), with potential lowering of slopes by 5–15 mm yr^{-1}. Such potential erosion has threatened nearly 62% of the surface above 1000 m a.s.l. in the high mountains of Slovakia. The role of this erosion increases with elevation because the average inclination of slopes increases from the forest belt through to the alpine (or up to the seminival) belt.

Average values for the potential retreat of slopes by water running on the slope surface (based on a calculation of potential soil loss)—in particular the geoecological belts of the discussed geomorphic units in the high mountains in Slovakia—are presented in Table 2.

and morphodynamic processes connected with differentiated relief and climate.

Table 2. Average potential retreat of slopes (mm yr^{-1}) due to surface runoff by geoecological belts in high mountains within geomorphic units of the Slovak part of the Western Carpathians.

	Tatry			Nízke Tatry		Malá Fatra				Oravské Beskydy		
Geoecological belt	Vysoké	Belianske	Západné	Ďumbierske	Kráľovohoľské	Krivánska	Lúčanská	Veľká Fatra	Choč	Babia hora	Pilsko	Average
Supramontane	4.75	8.60	8. 51	7.95	5.58	8.80	6.60	6.97	7.28	3.74	3.74	6.83
Subalpine	8.62	9.45	9.32	8.60	5.27	7.87	4.87	7.98	8.94	5.43	4.12	8.25
Alpine	9.70	9.76	9.64	8.49	4.40	—	—	—	—	—	—	9.37
Average	7.08	8.89	8.93	8.09	5.54	8.44	6.42	7.13	7.32	3.98	3.78	7.29

REAL RETREAT BASED ON INDIRECT METHODS

Lukniš (1973) calculated that the thicknesses of the layer carried down by glaciers in the Slovak part of the Tatra Mts. range from 28 to 40 m (33 m on average), while the volume of moraines is 4.918 km^3. The average rate of glacial erosion during the last glaciation in the southern part of the High Tatras can thus be determined as ca 0.55 to 1.10 mm yr^{-1} (this data not including material washed away by flow from surfaces covered with glaciers).

By indirect measurement, the same author also established (at 0.5 mm) the mean rate of surface retreat during the Holocene, in the Slovak part of the High Tatras.

REAL RETREAT OF SLOPES BASED ON DIRECT MEASUREMENT

The most elevated parts of the supramontane belt, including the subalpine and alpine belts, occur in the higher parts of the high mountains of Slovakia. The area of land at altitudes between 1000 m a.s.l. and the highest summits is about 184,000 ha. Each of the quoted belts is characterised by a different group of present-day morphogenetic and morphodynamic processes connected with differentiated relief and climate.

The area between the 1000 m isohypse and the present-day upper timberline (the supramontane belt) accounts for 73.4% of the total. It is mainly covered by protective forest at between 70 and 90% forest cover. Weak erosion processes related to running water prevail here and their rates are as follows:

- high-mountain spruce forest:
- slopes with inclines under 25°: average total soil loss = 9 (relative loss 43) kg ha^{-1} yr^{-1}
- slopes with inclines above 25°: average total soil loss = 17 (relative loss 134) kg ha^{-1} yr^{-1}
- forest free area (cutting, windfall, grassland): average total soil loss is 33 (relative loss 153) kg ha^{-1} yr^{-1}
- denuded surfaces with the destroyed vegetation cover or without it in the high-mountain forest belt (ski path, timber yard, skid road): average total soil loss is 161 (relative loss 2239) kg ha^{-1} yr^{-1}
- superficial chemical denudation in forest belt: retreat of slopes of 0.09 mm yr^{-1} (Kotarba 1971). According to the author and Stankoviansky (1983), superficial chemical denudation in the forest belt is also related to overland flow.

The subalpine belt accounts for 19.5% of the area of the zone between summits and the 1000 m isohypse. Dwarf pine scrub covers 16–62% (mean 33%) of the area (total area 21,000 ha). The rest of slope surfaces consists of permanent grassland on podzols, cambisols, rankers, rendzinas or pararendzinas (Šály 2006), to a lesser extent also bare rock surfaces of glacial relief, or block fields of periglacial relief.

Only a smaller part of slope surface in the alpine belt (from 1800–1950 m a.s.l. to about 2300 m a.s.l.) is rocky, while debris (the greater part mobile) accompanied by soil with disrupted a turf mantle prevails. Gravitation processes, nivation, cryogenic and aeolian processes dominate. Soil cover is missing and the surface consists of outcrops of granitization rocks that are subject to disintegration by frost and gravitation processes—collapsing rockwalls, rockfall and frost creep, but also the activity of debris flows and debris avalanches (Midriak 1983; Kotarba 1992).

Erosion processes also dominate in the subalpine belt, their rates being as follows:

- dwarf pine stands above the upper timberline: the average absolute soil loss is 12 (relative losses 20) kg ha^{-1} yr^{-1}
- subalpine meadows: average absolute soil loss is 25 (relative losses 41) kg ha^{-1} yr^{-1}
- denuded and destroyed soils (or surfaces) above the timberline: average absolute soil loss is 253 (relative loss 4812) kg ha^{-1} yr^{-1}
- chemical denudation of surface above the timberline: retreat of slopes of 0.04 mm yr^{-1} (Kotarba 1971).

However, running water-related processes in both the subalpine and alpine belts are also accompanied by gravitation, nivation and cryogenic processes in particular forms of water-gravitation-, nivation-gravitation-, cryogenic-gravitation- and partially aeolian processes. Their rates were estimated as follows:

- rockfall: retreat of carbonate rockwalls (of limestone, dolomite) about 0.01 to 0.43 mm, *in extremis* up to 3.00 mm yr^{-1}(according to Kotarba 1972, 1976)
- creep: retreat of slopes about 0.4 mm yr^{-1} (author and Dobija 1973)
- landslide: retreat of slope by a single event on average 55 mm
- snow avalanches: retreat of slopes of about 1.6 to 4.5 mm yr^{-1}, local retreat (as a microform) by a single event up to 150 (max 350) mm
- gelisaltation: retreat of slopes by needle-ice (pipkrake) activity during one total regelation cycle from 0.48 to 6.62 mm (average 2.23), with the number of regelation days 62 to 123 per year
- cryogenic processes: retreat of rock walls by frost weathering about 0.003 to 0.019 mm yr^{-1}
- nivation: retreat of slopes of about 2.5 mm yr^{-1}
- wind erosion: retreat of denuded parts of slopes by deflation from 0.00003 to 0.5 (on average 0.18) mm yr^{-1}
- anthropogenic destruction of surface: retreat of tourist trail surfaces on slopes above the timberline of about 5.9 to 26.6 (average 13.8) mm yr^{-1}

DISCUSSION AND CONCLUSION

The rate of surface erosion, as the most widespread geomorphic process, varies from low to high. In the forest belt, and also in the dwarf pine scrub and where surfaces are covered by turf, the mean soil removal is of only 0.001–0.007 mm yr^{-1}. In the case of denudation, this reaches 3.4 mm yr^{-1} on average, and in erosional rills and small gullies even up to 17.4 mm yr^{-1}.

However, the area above the upper timberline in the Slovak part of the Western Carpathians (as a whole) is characterised by a medium to relatively high real denudation rate. The mean rate of surface lowering (or slope retreat) is 0.27 mm yr^{-1} (in particular high mountain ranges this fluctuates between 0.10 and 0.72 mm yr^{-1}—see Table 3).

In the case of the average rate of slope retreat the situation is complicated (see also Kotarba and Starkel 1985). Up to 99% of the volume of the material removed from eroded landforms above the upper timberline comes from only ca. 8% of the total area—from

Table 3. Average real retreat of slopes (mm yr^{-1}) due to synergic activity of various (all) geomorphic processes in the area above the timberline in geomorphological units of the Slovak part of the Western Carpathians.

Geomorphological unit – *or part thereof*	Average real retreat of slopes (mm yr^{-1})
Tatry	0.27
– *Vysoké Tatry*	0.12
– *Belianske Tatry*	0.72
– *Západné Tatry*	0.39
Nízke Tatry	0.21
Malá Fatra	0.41
– *Kriváņska Fatra*	0.48
– *Lúčanská Fatra*	0.10
Veľká Fatra	0.27
Choč	0.64
Oravské Beskydy	0.34
Average	0.27

bare rock surfaces. If processes of water erosion below the timberline are taken into consideration (the rate there being less than 0.007 mm yr^{-1}), we can consider that the average rate of present-day lowering for the whole area above an altitude of 1000 m a.s.l. is 0.08 mm yr^{-1}.

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HIGH-MOUNTAIN ELEMENTS IN THE GEOMORPHOLOGY OF THE SUDETES, BOHEMIAN MASSIF, AND THEIR SIGNIFICANCE

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Abstract: The forested mountains ranges of the Bohemian Massif, including the Sudetes, typify moderately high mountain geomorphology (*Mittelgebirge*). However, the most elevated parts of the Sudetes also have landscape elements more readily associated with high-mountain relief. These include sub-alpine meadows and bare regolith-covered slopes, a multitude of relict periglacial landforms, as well as inherited Pleistocene glacial landforms. The present-day geomorphological activity in the terrain located at and above the timberline is much more evident than that recorded in the lower forested belt. Debris flows triggered by occasional downpours are the most potent geomorphic agents, also influencing hydrological conditions and vegetation patterns. Avalanches play a further role in determining the position of the timberline, but rockfalls are very rare. The contemporary development of certain small-scale periglacial landforms has been recognized, and close correspondence is found to exist between the tiered structure of morphogenetic domains recognized in the Carpathians and the highest massifs in the Sudetes.

Key words: high mountains, geomorphology, glaciation, debris flows, Sudetes, Bohemian Massif

INTRODUCTION

Mountain environments occur in a variety of types that often differ markedly from one another. This explains why there have been frequent attempts to establish a coherent classification scheme for mountains, in the hope that such a scheme would not only facilitate comparative analyses, but would also be related to basic physical characteristics of areas of high relief. In Central Europe, such classifications have been dominated by German terminology, hence the frequent references to the so-called *Hochgebirge* and *Mittelgebirge* classes. The type example of *Hochgebirge* came from the Alps, while the lower, largely forested mountain massifs of southern Germany and Bohemia were referred to as *Mittelgebirge*. The former term is easily translated as 'high mountains', but there are problems with adequate translation of the latter which may be variously termed 'middle mountains', 'medium-high mountains' or 'mid-mountains'. Nevertheless, these regions do constitute a specific environment which needs to be distinguished from the high mountains.

Characteristics of the Central European *Mittelgebirge* include altitudes of between 400 and 1500 m a.s.l., the general absence of distinctive glacial landforms, the dominance of gently rolling skylines, and the weak development of vertically differentiated vegetation belts due to insufficient height. In fact, the *Mittelgebirge* rarely rise above the timberline, unless severe human interference has taken place. Most of the terrain in the Central European Bohemian Massif conforms with the above definition of *Mittelgebirge*. However, a few mountain massifs, particularly the Karkonosze (*Krkonoše*) of Poland and the Czech Republic, show additional features which require special consideration. Jeník (1973) reviewed several existing classification schemes and argued that the presence of 'high mountains' elements in the Karkonosze is distinctive enough to warrant giving special consideration to this particular range. His principal arguments related to the number of altitudinal zones (climax ecosystem belts) and the existence of the most elevated parts above the timberline. The presence of four distinctive vegetation belts in the Karkonosze, and the large vertical extent of the zone above the timberline (up to 400 m) are decisive in ascribing a 'high mountain' character to this mountain range. Jenik proposed that the Karkonosze 'be regarded as middle-mountains with strongly represented high-mountain elements' (Jenik, 1973 p. 98–99).

While Jeník did not consider the dynamics of geomorphic systems as he framed his argument, the geomorphological perspective clearly complements his ecosystem approach. In Central Europe, slopes and valley floors below and above the timberline usually show very different rates of geomorphic change (Kotarba and Starkel 1972; Kotarba *et al.* 1987). Even during extreme hydrometeorological events (e.g. Czerwiński and Żurawek 1999) the forest belt shows considerable slope stability unless developed on bedrock that is highly susceptible to landsliding (e.g. Ziętara 1999), or on slopes heavily disturbed by human activity. In contrast, the sub-alpine belt is characterized by a much

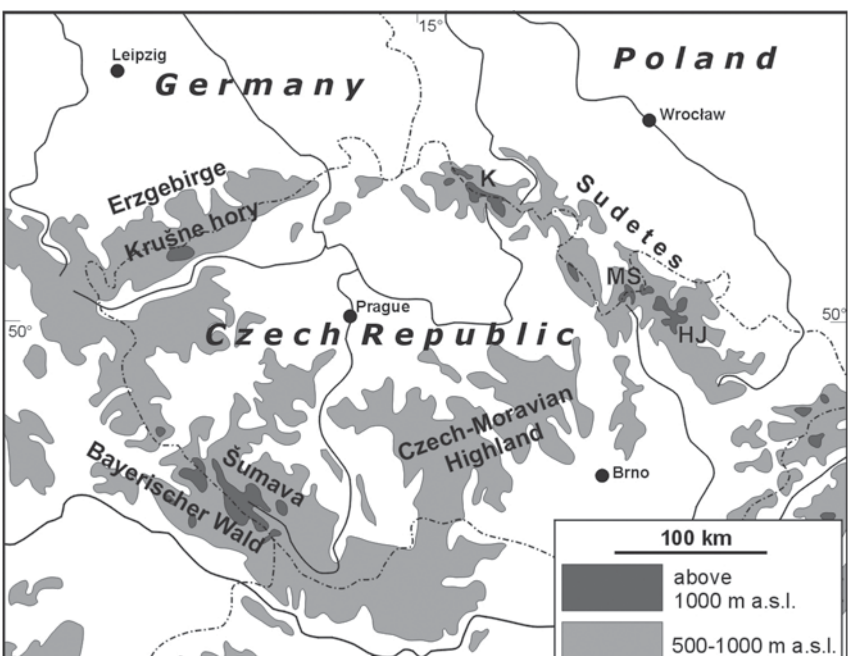

Figure 1. Location map and altitudes within the Bohemian Massif.
K—Karkonosze, MS—Śnieżnik Massif, HJ—Hrubý Jeseník

higher frequency of landforming events. Likewise, the geomorphic and geoecological effects of such events are much more evident and may propagate down into the forest belt (e.g. Krzemień *et al.* 1995; Łajczak and Migoń 2007).The presence of this sub-alpine domain, even be this of limited spatial and vertical extent, is very significant from the geomorphological viewpoint.

This paper thus reviews the existing regional literature, and develops the argument that the most elevated areas in the Sudetes are very distinctive geomorphologically and bear a number of affinities to high mountain environments. Particular attention is given to their inherited glacial relief, present-day cold-climate phenomena, and debris flows as indicators of slope instability.

STUDY AREA

The Bohemian Massif is a large tract of terrain located in Central Europe, north-east of the Alps and north-west of the Carpathians, mainly in the Czech Republic, though with outlying parts in Austria, Germany, and Poland (Fig. 1). Its most elevated parts occur around the perimeter, whereas the central part is low-lying, only occasionally exceeding 600 m a.s.l. The marginal mountain ranges are the Bayerischer Wald/Šumava in the south-west, the Krušne hory/Erzgebirge in the north-west, and the Sudetes in the north-east. These mountain ranges rise considerably above 1000 m a.s.l., culminating in Grosser Arber (1457 m), Klinovec (1244 m), and Śnieżka (1603 m), respectively. The latter have the most distinctive geomorphology, which is the subject of this paper.

The Sudetes have a long and complicated geological history which can be traced back to Precambrian times (Żelaźniewicz 2005; Mazur *et al.* 2006). Consequently, there is a great variety of geological structures and rock types present, with a corresponding diversity of general relief (Czudek *et al.* 1972; Migoń 2005; Balatka and Kalvoda 2006). However, the most elevated terrain is invariably developed on crystalline basement rocks of Palaeozoic age, mainly granite, gneiss, and mica schist.

Relative elevation of close to or more than 1000 m is typical for the highest parts of the Sudetes, and it influences contemporary climatic and vegetation patterns. The altitudinal zoning is best developed in the Karkonosze/Krkonoše and Hruby Jeseník (Fig. 2). These massifs have their highest areas above the timberline, which reaches an average altitude of 1230 m a.s.l. in the Karkonosze, and c. 1310 m a.s.l. in the Hruby Jeseník, although with much vertical variation locally (Treml and Banaš 2002; Treml 2007). Sub-alpine meadows and bare rock and regolith-covered surfaces occupy 55 and 10.5 km^2, respectively. Within the sub-alpine belt there are two distinct terrain types.

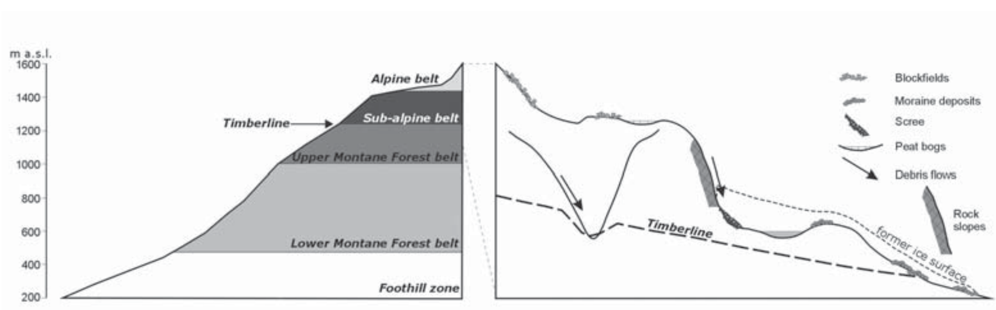

Figure 2. Altitudinal vegetation belts in the Karkonosze (left, after Fabiszewski 1985) and selected geomorphic characteristics of the uppermost parts of mountains (right). Note lowering of the timberline in certain fluvial valleys and in Pleistocene cirques, due to frequent avalanches and debris flows. No direction is implied in the right hand figure.

Firstly, there are summit surfaces of very low relief and negligible gradient. Secondly, these low gradient surfaces give way to very steep slopes, often greater than 30°, flanking former glacial cirques or deeply incised valley heads (Fig. 3).

The Karkonosze are the most extensively glaciated mountain range in the Bohemian Massif, with clear geomorphological evidence of cirque and valley glaciers having been present. The research on the glacial history of the range started in the late 19th century and the most recent summaries have been provided by Engel (2003) and Migoń and Pilous (2007). These show that the glacial geomorphology of these area is much better known than their glacial geochronology, attempts only recently having been made to delimit the ages of successive glacial advances or recessional phases with greater precision (Bourlès *et al.* 2004; Engel 2007). The most distinctive glacial landforms are the cirques which fed major valley glaciers (Fig. 4). There are six cirques on the northern side, including two overdeepened ones, and four compound cirques on the southern side, each consisting of several slope hollows adjacent to each other. All of the southern cirques, except Lábský důl, have their floors

Figure 3. High-mountain elements of the present-day landscape of the Karkonosze Mountains (the vicinity of Mały Staw Cirque: an elevated summit surface with active and relict periglacial landforms, a glacial cirque with steep slopes moulded by rockfalls, debris flows and snow avalanches, and the timberline; by R. Rąpała).

PLEISTOCENE INHERITANCE IN THE PRESENT-DAY RELIEF

One of the differences between the two most elevated mountain ranges of the Sudetes, the Karkonosze and the Hruby Jeseník, as opposed to remaining areas, lies in the occurrence of inherited Pleistocene glacial features. In neither range are these particularly extensive, unless one accepts the former existence of a glacier tongue more than 15 km long in the East Karkonosze, on the basis of a somewhat controversial interpretation of a few outcrops of till-like deposit (Carr *et al.* 2002). Even then, however, the geomorphic impact of this postulated glacier would have been negligible.

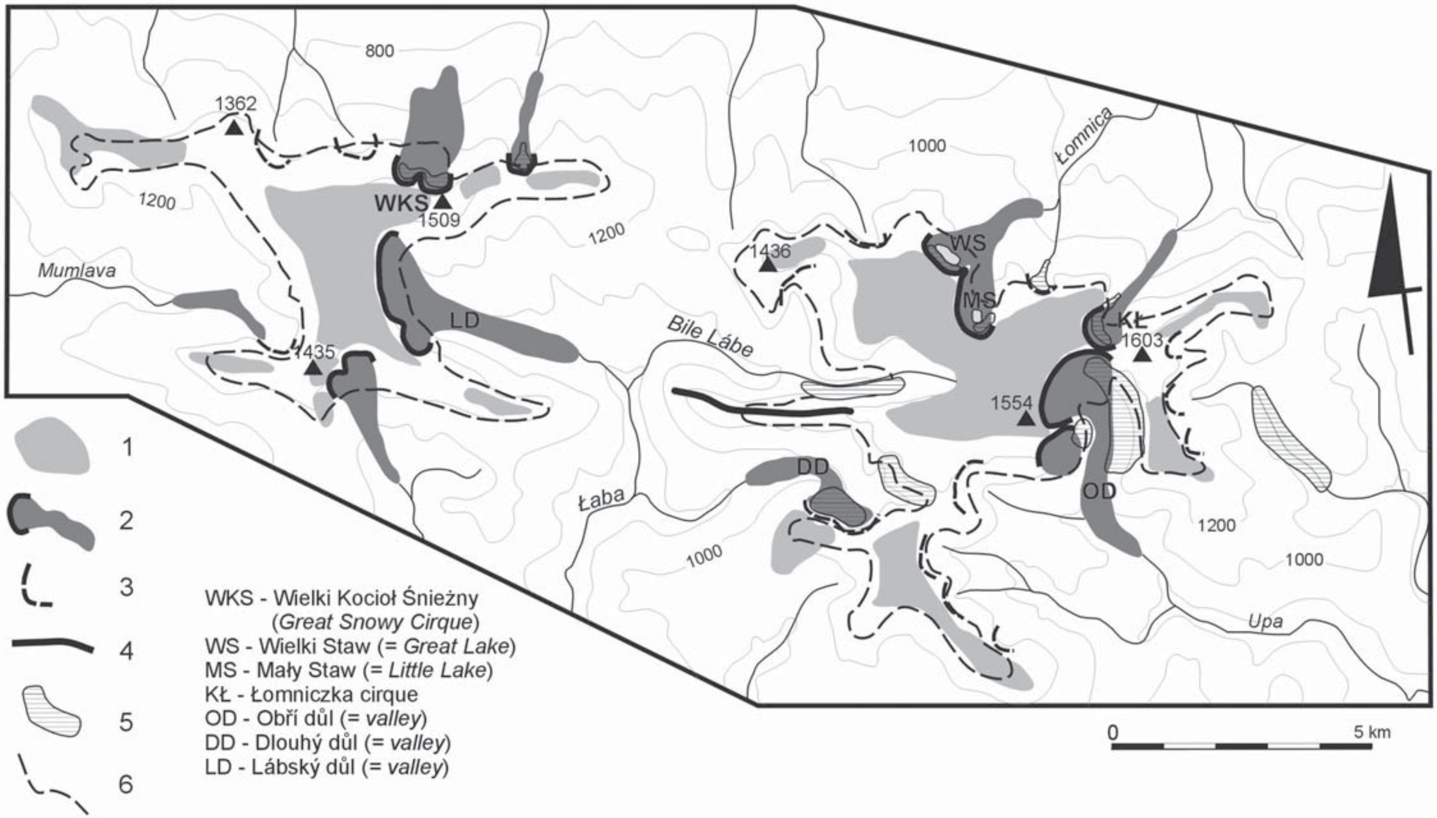

Figure 4. Distribution of high-mountain landscape phenomena in the Karkonosze.
1—summit plateau, 2—cirques and the extent of Pleistocene glaciers, 3—steep valley heads (possible nivation hollows), 4—sharp crests, 5—areas of frequent debris flow occurrence, 6—timberline.

sloping towards outlet valleys. Two north-facing cirques have deep lakes with respective depths of 24.4 m (Wielki Staw = Great Lake) and 7.3 m (Mały Staw = Little Lake). Although rock slope sections occur in each cirque, their extent varies. The Wielki Kocioł Śnieżny (Great Snowy Cirque) is the best developed cirque, with a 180 m-high amphitheatre headwall, dissected by numerous fracture-controlled ravines (Fig. 5). In other cirques, the planar or step-like rock slopes are of lesser extent, and cirque morphology is dominated by steep slopes covered by a thin regolith veneer. All the well-developed cirques are located within the subalpine belt. In addition, there are about 30 steep valley heads and slope hollows that possibly hosted glaciers during the Pleistocene (Šebesta and Treml 1976), though clear geomorphological evidence for this is not available, and the features lack the characteristic rocky headwalls and flat floors of classical cirques. Some of these problematic landforms are located within the subalpine belt, while others lie within the upper montane belt.

Glacial depositional forms also show varied levels of development. Morainic terrain is most evident in the forefield of the Śnieżne Kotły (Snowy Cirques) and in the Łomnica valley, where there are arcuate and longitudinal ridges with intervening areas of hummocky ground, large boulder piles, and closed depressions with peat bogs (Traczyk 1989). Local relief in the accumulation zone attains 20 m in elevation. Both Obří důl and Lábský důl hosted typical valley glaciers 4–5 km long, and remnants of recessional moraines occur (Engel 2007). In contrast to the glacial erosion zone, located largely above the timberline, depositional land-

Figure 5. The geomorphological landscape of the Wielki Śnieżny Kocioł, Karkonosze Mountains, with rocky headwalls, fracture-controlled ravines, debris cone and debris flow tracks, bears much resemblance to typical high-mountain terrain. The debris flow track in the left part of the photo originated in August 2006 (by P. Migoń).

forms occur within the natural forest belt and are best seen in areas of forest clearance or recent forest dieback.

The extent of glaciation in the Hruby Jeseník was considerably more limited than in the Karkonosze. Only the valley head of Velká kotlina bears relatively clear signs of glacial action from a glacier which was c. 1.5 km long, although it does not have a continuous rocky cirque headwall (Prosová 1973). Other steep-sided valley heads lack comparable geomorphological evidence and long-lived snow patches rather than true glaciers probably existed in these locations. The reasons for the relative under-development of glaciers in the Hruby Jeseník relate to the topography. As in the Karkonosze, valley heads were mainly fed by snow blown from the summit flats. Therefore, the origin and potential size of a glacier was critically dependent on area supplying snow. In the Hruby Jeseník, narrow ridges and broad convexities typify the highest surfaces, providing less favourable source areas for snow and subsequent accumulation in lee-side slope hollows (Migoń 1999).

In the highest parts of the Sudetes, the geomorphic legacy of periglacial conditions is much more widespread than the inherited glacial relief (Sekyra 1961; Żurawek 1999; Traczyk and Migoń 2003; Křížek 2007). This includes a variety of landforms, such as tors, mid-slope rock cliffs and associated benches (interpreted as frost-riven cliffs and cryoplanation terraces), block fields and block streams, relict rock glaciers, solifluction lobes and terraces and patterned ground. In addition there are extensive Pleistocene cover deposits in the form of both solifluction mantles on slopes, as well as *in situ* mountain-top detritus on summit flats. However, many of these landforms and deposits are also known from the less elevated parts of the Bohemian Massif (Czudek 2005), even if partly masked by forest, hence their lack of any greater distinctiveness for the Karkonosze or Hruby Jeseník.

The impact of Pleistocene glacial and periglacial re-modelling of the landscape on contemporary geomorphic systems has been considerable. First, although cirques have developed from preglacial valley heads that were partially controlled by the pre-existing refief (Jeník 1961; Migoń 1999), glacial erosion has enhanced local relief. The resulting steep rock slopes with ravines, bedrock spurs, discontinuous coarse regolith cover and scree accumulation are new geomorphic features formed through glacial erosion and subsequent mass wasting processes. Second, a blanket of loose deposits has been left perched on steep slopes within cirques, glacial troughs, and deeply incised fluvial valleys. Third, in some of the formerly glaciated areas lakes have formed in the cirques, and these now act as both local denudation base-levels and sediment traps interrupting the sediment transfer cascade.

DEBRIS FLOWS

Debris flows are undoubtedly the most spectacular contemporary surface processes acting in the most elevated parts of the Bohemian Massif, even if their recurrence intervals are rather low. They are invariably generated by very heavy precipitation events, near or in excess of 100 mm daily, such as those in 1882 and 1897 in the Karkonosze (Pilous 1973). However, establishing a rainfall threshold value for debris flows remains a challenge in these areas because of the scarcity of adequate meteorological data. The available evidence suggests that the required rainfall amounts are lower than those identified in the Tatra Mountains, and may be as low as 15–20 mm per hour (Migoń *et al.* 2002).

Within the Bohemian Massif the largest and most frequent debris flows are those occurring in the Karkonosze. More than 200 individual cases have been recorded on the Czech side (Pilous 1973; Migoń and Pilous 2007), while there is evidence for more than 70 debris flows of various ages on the northern (Polish) side (Parzóch *et al.* 2007). About 90% of the flows were generated above the timberline at 1200–1450 m a.s.l., though many individual flows reach the forest boundary and travel downslope to valley floors or mid-slope benches through the upper montane forest (Fig. 6). Two geomorphic settings are most favourable if debris flows are to occur. These are within glacial cirques, such as Obří důl in the Czech part or Wielki Kocioł Śnieżny and Kocioł Wielkiego Stawu in Poland, and on steep valley sides (e.g. the northern slope of Dlouhý důl in the Czech part). The cirques are characterised by sharp slope gradients and a widely-distributed loose regolith cover. In addition, fracture-controlled ravines collect debris liberated by weathering and act as suitable routes for debris transfer (Fig. 5). Valleyside debris flows are known mainly from the part of the Karkonosze underlain by metamorphic rocks (Pilous 1973), but the reasons for this apparent lithological control are probably complex. On one hand, regolith covers derived from gneiss and schist contain a greater proportion of the fine fraction, and are hence more susceptible to landsliding. However, the overall progress of erosion has been more efficient in metamorphic rocks, resulting in deeper and more incised river valleys than in the granite terrain.

Figure 6. Shallow landslide scars, debris flow tracks and the irregular upper forest boundary in the Łomniczka cirque, Karkonosze (by K. Parzóch).

Debris flows in the Karkonosze have varied in length. The longest ones travelled for 900 m, though the mean length is a little more than 200 m (Pilous 1977; Parzóch *et al.* 2007). Two types of flow may be recognized, these in part corresponding with the two geomorphic settings described above. The most common are flows on regolith-covered slopes which typify the valleys and the majority of glacial cirques. These were usually initiated by shallow debris slides and then channelized into pre-existing gullies and other slope concavities. Most flows have a clear tripartite morphology, with a wide shallow scar in the upper slope, followed by a long erosional furrow bordered by levee ridges up to 3–4 m high, and an irregular deposition zone near the lower end. Steep toes and terminal lobes have only formed rather infrequently.

The other type is found almost exclusively in the Wielki Kocioł Śnieżny cirque, which is the site in the Karkonosze most similar to classic high-mountain cirques. Here, flows are initiated within debris-filled ravines and travel onto the scree cones that are the areas of ultimate deposition (though some erosion also occurs in the upper cone). Debris flows are important agents of material delivery and redistribution to and within the cones which, though primarily built by rock fall, offer clear exemplification of the concept of alluviated slopes developed in the Tatra Mountains by Kotarba (1989). The most recent flow occurred in early August 2006, within the eastern wall of the cirque (Migoń *et al.* 2006). The source area was located in the upper part of a ravine, c. 150 m above the foot of the rock slope, and was initially confined to the narrow ravine. After reaching the cone, the liquefied debris mass bifurcated and followed several parallel routes downslope (Fig. 7). The longest track on the cone was of 190 m, while the total length of all tracks was about 330 m. The most distinctive landforms produced by the flow were boulder trails in the upper cone and levees up to 1 m high in the distal part. Significantly, the

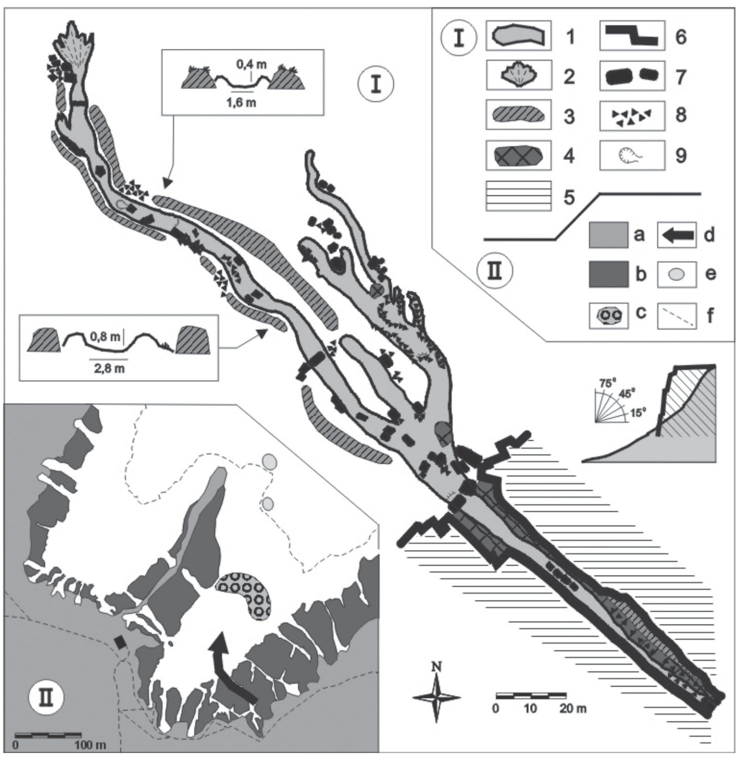

Figure 7. Groundplan of the 2006 debris flow in the Wielki Śnieżny Kocioł, Karkonosze Mts. (after Migoń *et al.* 2006; drawn by M. Kasprzak).
I—debris flow geomorphology: 1—flow track, 2—alluvial fan, 3—levees of ancient flows, 4—granite bedrock outcrops, 5—rock slopes, 6—rock walls, 7—boulders >1 m long, 8—stone runs, 9—hollows (the uppermost part of the track is not indicated).
II—geomorphic setting: a—summit plateau, b—rock slopes, c—moraine ridge, d—debris flow track, e—lakes, f—footpaths.

main track was confined by older boulder-rich levees, overgrown with shrubs and the occasional rowan, which testify to previous episodes of even higher–magnitude flows. A flow was triggered by extreme precipitation in the western part of the Karkonosze between 3 and 7 August 2006 (Fig. 8). The total 5-day rainfall was 404 mm, including 241 mm falling on 7 August (Vaškova and Metelka; pers. comm.). However, as the hourly intensities only occasionally exceeded 10 mm h^{-1}, and only once were greater than 20 mm h^{-1}, the threshold intensities to initiate flows are evidently less the 30–40 mm h^{-1} recorded for the Tatra Mountains (Kotarba 1992).

In the Hrubý Jeseník about 100 individual debris flows have been recorded (Gába 1992), all within valley head/valley-side settings, and the vast majority of them near the upper boundary of the forest belt. Their length is similar to those recorded in the Karkonosze, attaining maximum lengths of 800–900 m. As in the Karkonosze, structural and lithological controls show clearly. The Keprník area, notorious for frequent debris flows, is a structural dome built predominantly of mica schists which weather by hydration and are easily disintegrated by the slaking that facilitates creep and downslope bending (Hrádek 1984). Moreover, many potential failure planes along cleavage and

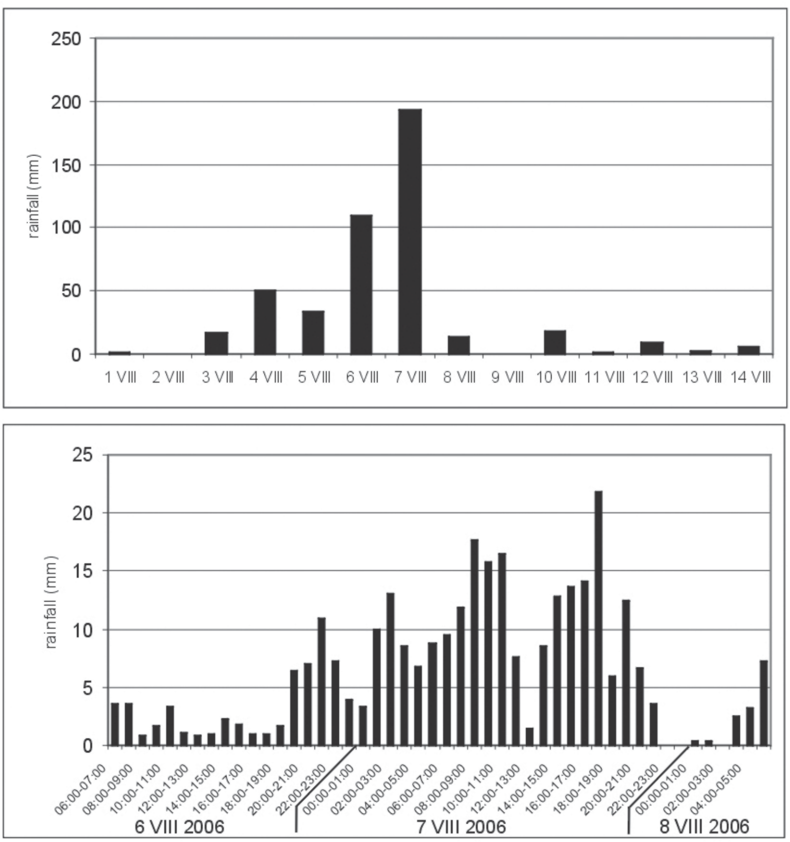

Figure 8. Rainfall characteristics during August 2006 at Labská louka, Karkonosze Mts. (after Migoń *et al.* 2006; based on data kindly made available by the Czech Hydro-Meteorological Office).

schistosity surfaces exist within the weathered mica schists, and some steep slope surfaces are parallel to schistosity.

Here debris flows are again caused by heavy rainfall events. For example, on 1 June, 1921, close to this debris flow area, a rainfall of 196.5 mm occurred in 24 hours, including 118 mm in only 90 minutes (Sokol and Vavřík 1971). On 4 July, 1991, there was a huge debris flow in the valley of the Keprnícký potok, and precipitation is likely to exceed 100 mm (Gába 1992). Unfortunately, no hourly precipitation records are available.

OTHER RAPID GRAVITATIONAL PROCESSES

Under contemporary conditions, rockfalls and rock slides are very infrequent in the Sudetes. An exception is the larger of the twin glacial cirques of Śnieżne Kotły in the Karkonosze. Its headwall is a semi-circular, serrated rock slope up to 180 m high, with a gradient of 50–90°. Below the rock walls, and particularly at the outlet of the numerous ravines, active scree cones occur (Fig. 5). The presence of little vegetation in the proximal parts of the cones, as well as of occasional piles of large angular blocks, indicates a continuous supply of debris to the cones, otherwise moulded by debris flows. However, extensive talus deposits, now largely overgrown by either spruce forest or dwarf pine, rowan, and grass communities, can be identified in all cirques. Collectively, they show that gravity-driven geomorphic processes must have been more efficient in the past. At the moment, no data are available to specify if the re-modelling of cirques was essentially a paraglacial phenomenon, conditioned by glacier decay and rapid environmental change during deglaciation and close to the end of the Pleistocene (Engel 2007), or continued well into the Holocene, including in the Little Ice Age.

The reasons for the minor role of rock-slope failures in the present-day geomorphic system are likely to be complex. Among factors to consider are an insufficient gradient of the steepest slopes (gradients >50° are very rare), the limited area occupied by bedrock outcrops, the protective role of vegetation, and material availability after a period of more frequent rock slope failures in the past.

The sub-alpine and alpine belts in the Sudetes are known for their avalanches. In the Karkonosze, there are 56 avalanche tracks in the Czech part and 51 tracks on the Polish side (Spusta *et al.* 2007). There are striking similarities in locations with debris flows. They occur within all glacial cirques, and the most pronounced valley heads (sometimes referred to as nivation hollows), and follow deeply-incised ravines in steep valley sides. The longest run-out zones of avalanches are well in excess of 1500 m. As only a fraction of avalanches incorporate mineral material these do little geomorphological work overall, their effects being minor in comparison with those of debris flows.

Avalanches are much less frequent in the Hruby Jeseník and Śnieżnik Massif and are confined to a few localities. They occur in steep valley heads, partly remodelled by Pleistocene glaciers, and located above the timberline. However, the direct role of avalanches as geomorphic agents is negligible (Kříž 1995).

PRESENT-DAY PERIGLACIAL PHENOMENA

Another important boundary in the mountains is the lower limit of permafrost. In the Bohemian Massif, even the most elevated mountain ranges lie well below this limit. Nevertheless, frost processes are active above the timberline and give rise to an assemblage of minor yet distinctive landforms, particularly pronounced in the Karkonosze. According to Křížek (2007), who considered both frost susceptibility of the substratum and the frequency of regelation cycles, non-sorted patterned ground—earth hummocks, small stone circles and stripes, and low solifluction steps are developing under present-day conditions. In addition, many ploughing blocks littering the summit surfaces of the

Karkonosze show the evidence of movement, including disturbance of man-made structures dating back to the 18–19th centuries (Sekyra *et al.* 2001).

Recent frost heaving and frost sorting activity has also been documented in the Hruby Jeseník. Earth hummocks (thufurs) are widespread above 1350 m a.s.l., their ongoing development gaining confirmation by AMS radiocarbon-dating of one hummock base as dating back to 2,090±35 BP, as well as by detection of a core of segregation ice in the summer period of 2003 (Křížek 2007). Irregular sorted nets and some ploughing blocks may also be active. On the summit dome of Śnieżnik small stone circles exist and ground is frost-susceptible (Klementowski 1998). Nevertheless, conclusive evidence for current activity is lacking.

GEOMORPHOLOGICAL AND GEOECOLOGICAL CONSEQUENCES

Debris flows and avalanches in the Karkonosze and Hruby Jeseník, though very localized within the respective mountain ranges, contribute to the geomorphological and geoecological distinctiveness of these areas. Pleistocene glacial cirques and steep valley heads and valleysides above or at the timberline constitute a very specific landscape type that differs from otherwise forested and thick regolith-covered slopes typical of the 'middle mountains'.

From the geomorphological viewpoint, these areas are the most dynamic in the entire Bohemian Massif. The evidence for this is twofold. First, debris flows and avalanches rarely occur outside former cirques or away from particularly steep valley sides, and are almost unknown to originate within the forest belt. In certain localities, such as in Obří důl in the Karkonosze, debris flow tracks of variable age cover c. 15% of the west-facing upper valley side area (Pilous 1973, 1977). The frequency of debris flows varies. In the Łomniczka cirque and Wielki Kocioł Śnieżny in the Polish Karkonosze and Obří důl in the Czech part, major debris flows occur at least every 20–30 years, often with several flows triggered by the same event. In other places, only a few events have been recorded in the last ca. 150 years. However, a recent dendrogeomorphological study of debris flow deposits in the Hruby Jeseník (Malik and Owczarek 2006) suggests that the frequency of events is higher than previously recognized, and that some episodes of debris re-mobilization may have gone unnoticed. Second, debris flows accomplish much more geomorphic work than other mass-wasting processes active in the sub-alpine domain (Bieroński *et al.* 1992). Likewise, landform changes associated with flows are both spectacular and persistent, even if revegetation obscures part of the evidence over the long term (Parzóch and Dunajski 2002). Slope lowering in the source zone, where the entire regolith cover is known to be stripped away, may be of up to 1 m, whereas the depth of erosional furrows in the transit zone varies from 1 to 5 m (Pilous 1973). This magnitude of change far outweighs the geomorphic effects of deflation, surface wash or piping (Parzóch 1994).

Debris flows and avalanches have important geoecological consequences, particularly when they penetrate the forest belt. Landscape effects include local modifications of the timberline position (Treml 2007), the origin of linear damage zones within the upper montane forest stands (Sokol and Vavřík 1971; Pilous 1973; Parzóch *et al.* 2007), enhanced biodiversity in areas frequently affected by gravitational phenomena and the origin of mosaic vegetation patterns (Jeník 1961; Dunajski 1998). Conservation approaches to damage inflicted by debris flows and avalanches vary between different mountain ranges, depending on their status of protection. In the Karkonosze, within the National Parks, it is generally recommended that terrain be left without intervention (Pilous 1977), although erosional furrows in the forest belt may undergo "soft management" entailing the construction of timber-made check dams across furrows (Parzóch and Dunajski 2002). In the Hruby Jeseník, debris flows are considered a serious threat to forest management, and much effort is expended on reducing their impact. In order to prevent further erosion, stabilization works including the construction of check dams and aforestation are carried out on the tracks shortly after the flows have occurred. Scars, levees and terminal lobes may be obliterated in the course of engineering work (Sokol and Vavřík 1971).

Although slope gradient and hydrological conditions are important factors controlling the occurrence of debris flows and other types of mass movement, the availability of suitable material is decisive. An instructive case if afforded by the most recent event in Wielki Śnieżny Kocioł (Migoń *et al.* 2006). The 2006 flow here was generated in a ravine that did not generate flows during the heavy rainfall of July 1997, hence it may be hypothesized that the debris accumulated in the ravine was not at that time yet at the yield point. On the other hand, an adjacent fracture-aligned cleft (Fig. 5) generated a major flow in 1997, but did not produce any significant debris movement in 2006, apparently because of debris exhaustion. In 1897 some of the debris flows in Obří důl were of such magnitude that regolith was completely stripped away and bare rock exposed. No subsequent debris flows have been recorded in the last 110 years at these localities (Pilous 1973).

The summit surfaces above the timberline constitute a very different high-mountain environment, bearing more similarities to northern European, cold-climate mountain plateau, rather than to alpine mountain ranges. Low gradients, apparently inherited from pre-Quaternary times, account for considerable slope stability and allow for the development of a range of periglacial processes and landforms. In the Karkonosze, the notion of 'Arctic-Alpine tundra' is widely used to describe the environment of the most elevated parts (Soukopová *et al.* 1995). In the Tatra Mountains, Jahn (1958) distinguished three altitudinal belts, each typified by its own active periglacial process. These were the earth hummock belt (1500–1800 m) with least activity, the solifluction belt (1800–2000 m) and the sorted patterned ground belt (>2000 m). By contrast, in the Karkonosze, periglacial processes appear active above 1400 m a.s.l. (Křížek 2007). The lower altitudinal limits of these belts in the Karkonosze may reflect differences in terrain conditions rather than climate: the lower slope inclinations in the Karkonosze provide a greater area for sorting and heaving processes to develop with less disruption by gravitationally driven downslope processes than occurs in the Tatras.

CONCLUSIONS

The mountains of the Bohemian Massif are generally low- to medium altitude ranges, located almost entirely within the natural forest belt. As such, and leaving chemical denudation aside, they are characterized by rather low levels of geomorphic activity at present, except for sporadic extreme hydro-meterological events capable of inducing considerable landform change along valley floors (Czerwiński and Żurawek 1999; Migoń *et al.* 2002). Forested slopes in particular appear stable, unless locally disturbed by logging operations.

However, the status of the most elevated parts of the Sudetes is different. They meet certain geoecological criteria for high mountains according to Troll (1973), rising above the timberline and above the lower altitudinal limit for active periglacial phenomena. Altitudes do not exceed the contemporary snow line, but inherited glacial and nival landforms are widely present. In addition, the sub-alpine belt differs significantly from the forest belt in terms of rates of geomorphic change and the efficacy of the denudation system. Summit surfaces above 1300–1400 m a.s.l. are the domain of cryonival processes, whereas the adjacent steep slopes are subjected to rainfall-triggered debris flows and avalanches. The former, although not as frequent as in the high mountains, are the most efficient geomorphic drivers of change, producing new landforms on these slopes and creating new pathways of sediment transfer. Occasionally, the impact of gravity-driven processes initiated in the sub-alpine belt extends into the forest belt, influencing the position of the timberline. Thus, there is close correspondence between the tiered structure of morphogenetic domains recognized in the Carpathians (Kotarba and Starkel 1972; Kotarba *et al.* 1987) and the highest massifs in the Sudetes. The extent of high-mountain components in their present-day landscape is small, but they give a very distinctive character to the otherwise typical 'middle mountain' type of relief, influencing their biodiversity as well (Jeník 1961).

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ARE THERE GEOMORPHIC INDICATORS OF PERMAFROST IN THE TATRA MOUNTAINS?

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Abstract: This paper concerns itself with the issue of relief and permafrost in mountains in which the latter phenomenon is only present in the form of isolated patches, as is the case in the Tatra Mountains. Thus the slope morphology and morphodynamics at three permafrost sites in the Tatra Mts. are discussed in the context of the presence of periglacial landforms, especially indicative forms of permafrost. No distinct morphological evidence as to the presence of permafrost was in fact noted.

Key words: permafrost, high-mountains at mid latitudes, debris slope, periglacial landforms

INTRODUCTION

Permafrost is an important feature of periglacial environments as its presence modifies significantly the conditions under which periglacial relief develops. Permafrost is defined as soil or rock that has remained continuously at a temperature below 0°C for more than two years (Harris *et al.* 1988).

Recently, there have been a number of reports of the presence of permafrost in the highest belts of many unglaciated mountains at mid latitudes (Urdea 1992; King and Ackerman 1993; Kędzia *et al.* 1998; Ishikawa *et al.* 2003,; Dobiński 1997, 1998, 2005; Kern *et al.* 2004; Gude *et al.* 2003; Julián and Chueca 2007; Zacharda *et al.* 2007). The Tatra Mts. belong to this group (Dobiński 1997, 1998, 2004; Kędzia *et al.* 1998; Mościcki, Kędzia 2001; Gądek, Kotyrba 2003; Gądek, Żogała 2005). J. Warburton (2007) after J. Brown *et al.* (1997) includes the Tatras, as along with the Fagăraš and Retezat Massifs in the Southern Carpathians (of Romania) within the category of high European mountains featuring isolated patches of permafrost (accounting for less than 10% of their area).

The Tatra Mountains form one of the highest ranges in the Carpathian arc (peaking at Gerlach, 2666 m a.s.l.). Their relief is of alpine character and the climate is severe enough (Table 1) for the middle periglacial belt to be present above the upper timberline, i.e. 1550 m a.s.l. (Jahn 1975). On the basis of climatological analysis and indices of freezing and thawing, W. Dobiński (1997, 2004) asserts that there are potential conditions for permafrost to be present in this belt. Patches of permafrost are expected above 1930 m a.s.l. on north-facing slopes, and above 2300 m on south-facing ones

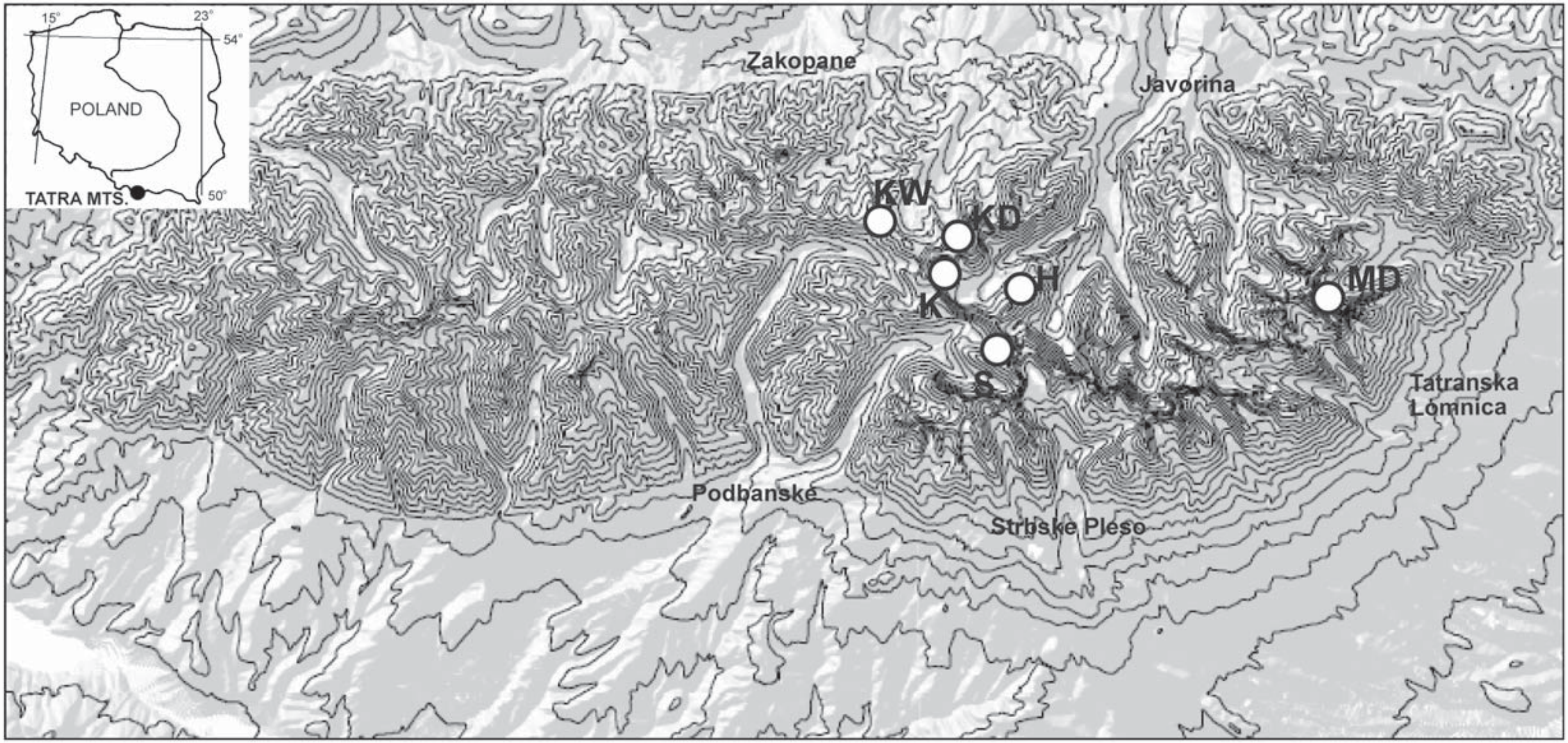

Figure 1. Occurrence of permafrost in the Tatra Mts., as documented by geophysical and georadar examination and direct observations. KW—Kasprowy Wierch (1990 m a.s.l.), K—Dolinka pod Kołem (2100 m a.s.l.), H—Hruby Piarg (1800 m a.s.l.) S—Śnieżny Kocioł (1900 m a.s.l.) (Dobiński 1997, 1998, 2004), KD—Kozia Dolinka (1950–2000 m a.s.l.) (Mościcki, Kędzia 2001; Kędzia 2004); MD—Medena Kotlinka (2000 m a.s.l.) (Gądek, Żogała 2005)—all as shown in Fig. 2.

(Dobiński 1997, 2004). Additionally, the lower climatic limit of permafrost patches corresponds with the altitude at which perennial and longlasting snow patches exist in shadowy fragments of talus slopes, usually in north-facing glacial cirques.

Table 1. Climate characteristics for the Tatra Mts.

Meteorological station		Łomnica	Kasprowy Wierch	Skalnate Pleso	Dolina Pięciu Stawów Polskich	Hala Gąsienicowa	Strbske Pleso	Hala Ornak
Altitude (m a.s.l.)		2635	1991	1786	1670	1520	1330	1109
Mean annual air temp. (°C)		-3.8	-0.8	1.6	1.1	2.3	3.4	3.2
No. of days with temp. min <0°C and max >0°C		88	78	111	94	93	118	126
No. of days without frost		4	48	93	82	102	120	120
Precipitation (mm)	January		142		71	70		74
	July		215		260	247		206
	Mean annual	1645	1889	1323	1692	1664		1490

after T. Niedźwiedź (1992) and M. Konček, ed. (1974)

The 1990s saw studies of permafrost in the Tatra Mts. carried out on the basis of such methods as BTS measurement, electroresistivity sounding, georadar, seismic sounding, infrared imaging, etc. The results of these studies pointed to the possible existence of permafrost in isolated patches within the Tatra Mts., and specifically at: Kasprowy Wierch (1990 m a.s.l.), Dolinka pod Kołem (2100 m a.s.l.), Hruby Piarg (1800 m a.s.l.), Śnieżny Kocioł (1900 m a.s.l.) (Dobiński 1997, 1998, 2004), Kozia Dolinka (1950–2000 m a.s.l.) (Mościcki and Kędzia 2001; Kędzia 2004) and Medena Kotlinka (2000 m a.s.l.) (Gądek and Kotyrba 2003; Gądek and Żogała 2005) as is shown in Fig. 1. In the Medena Kotlinka Valley, the author even observed an outcrop of massive ice on a slope of a nival niche (Fig. 2) during the extremely hot summer of 2003. The results of the above mentioned studies indicate that patches of permafrost are usually located in the bottoms and on the slopes of high-elevation glacial cirques. At sites in which permafrost was found, coarse debris of different sizes, including even boulders, predominated in slope covers. Grain-size composition of slope covers probably facilitates cooling of the ground (via the so-called chimney effect), as has been noted in the Alps (Delaloye *et al.* 2003; Gude *et al.* 2003).

B. Etzelmüller *et al.* (2001) declared that morphological indications of the presence of permafrost in the mountains relate to:

- the creep of coarse debris, which is ice-saturated, and leads to the formation of rock glaciers and push moraines;
- ice bodies buried by debris thicker than the maximum active-layer thickness, as revealed in the formation of thermokarst, ice-cored moraine and palsas.

Periglacial forms, such as patterned ground and solifluction lobes, can be treated as morphological indicators of permafrost, but they are not unambiguous because these forms also develop in areas without permafrost.

Fig. 3 presents the location of periglacial forms in the Tatra Mts. on the basis of earlier geomorphological examination (Rączkowska 2007). The area was found to support periglacial forms of different types and sizes, including solifluction lobes, patterned ground and rock glaciers. Yet the rock glaciers occurring there are relict forms (Kotarba 1991–1992), and so cannot be treated as evidencing permafrost, even

Figure 2. Outcrop of massive ice body (4 m high) in Medena Kotlinka Valley (Slovak part of the High Tatras) in August 2003 (indicated by arrow).

if ground temperature is very low inside glacier tongues, as was the case in the relict rock glacier near the tarn called Hińczowy Staw (Kędzia *et al.* 2004). Moreover, solifuction lobes and patterned ground are not attributed to the presence of permafrost, their present-day activity being due to diurnal and seasonal freeze-thaw cycles (Rączkowska 2007).

The aim of this paper has been to discuss—by reference to the Tatra Mountains—the relationship between landform development and the presence of permafrost in mountains in which the latter is only present in isolated patches. To that end, answers were sought to the following questions:

- Are there any geomorphic indicators of permafrost being present?
- Does permafrost influence present-day relief development?
- Does permafrost influence slope morphodynamics?

Answers to the above questions were expected to arise from examinations of slope morphology and morphodynamics in those areas of the Tatra Mts. in which permafrost could be detected. Three sites with permafrost—the Kozia Dolinka Valley, the Medena Kotlinka Valley and the Kasprowy Wierch summit—were investigated geomorphologically by author. In this, detailed geomorphological mapping was used as the basic technique.

SLOPE MORPHOLOGY AT PERMAFROST SITES

THE KOZIA DOLINKA VALLEY SITE

The Kozia Dolinka Valley is the uppermost glacial cirque in the Sucha Woda Valley of the Polish High Tatras (Figs. 4 and 5). Permafrost was evidenced there using indirect methods, mainly geophysical and BTS. Isolated patches of permafrost exist in the lower part of shadowed talus slopes (Fig. 4) below the high rockwall of the Kozi Wierch summit. The depth of the active layer is an estimated 2 m (Kędzia *et al.* 1998; Mościcki and Kędzia 2001; Kędzia 2004). Fig. 5 presents a geomorphological sketch of the Kozia Dolinka Valley at the scale 1:10 000. High rockwalls surround a relatively narrow valley bottom,

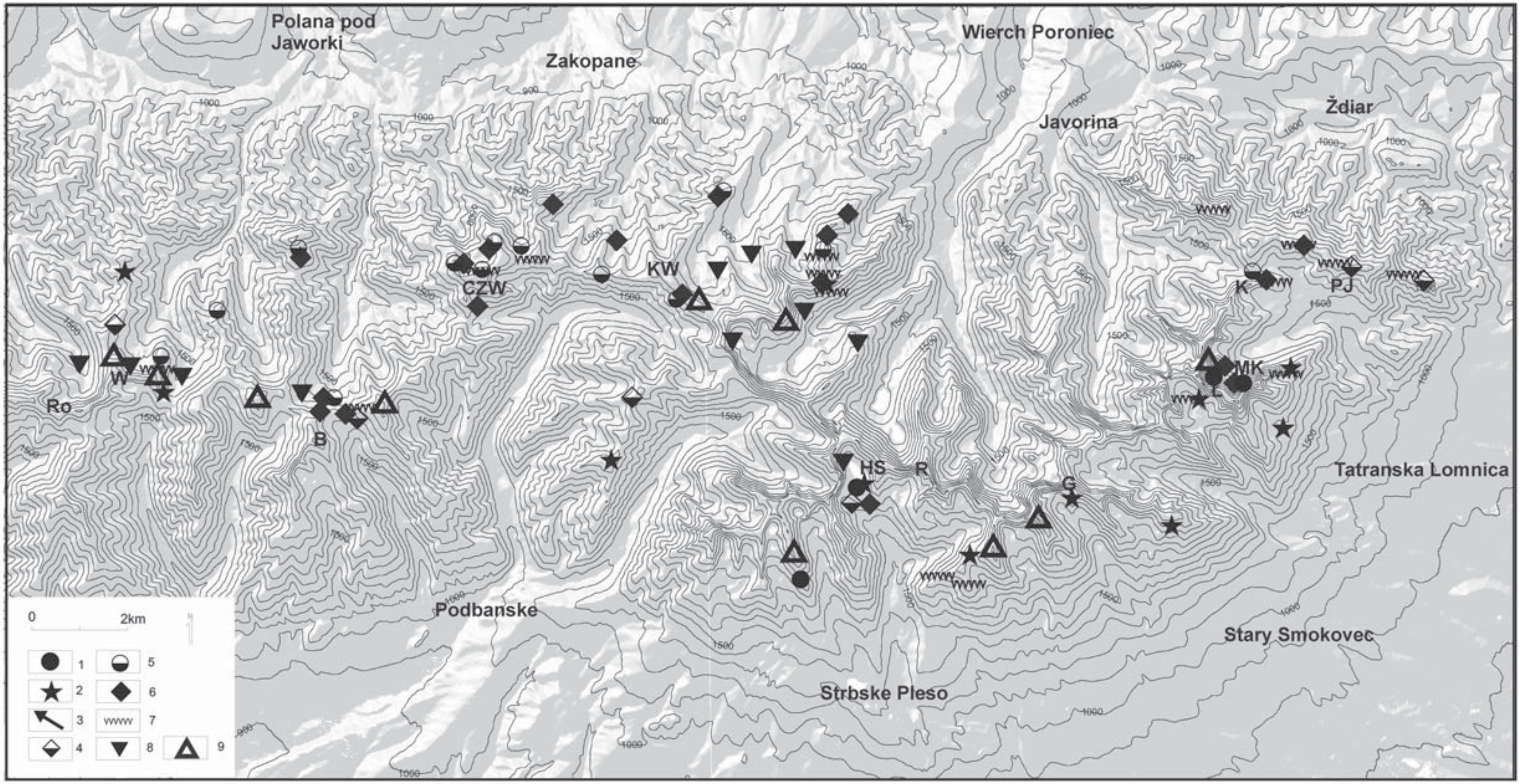

Figure 3. Periglacial landforms in the Tatra Mts. 1- sorted circles, 2—sorted polygons, 3—sorted strips, 4—miniature patterned ground, 5—thufurs, 6—solifluction lobes, 7—solifluction garlands, 8—rock glaciers, 9—blockfields (after Rączkowska 2007).

Figure 4. View of the Kozia Dolinka Valley and northern slopes of the Kozi Wierch summit (2291 m a.s.l.) (by S. Kędzia).

filled with glacial drift deposits, which in the southern parts consist mainly of an accumulation of giant boulders. Talus slopes between rockwalls and the valley bottom form heaps and cones. Permafrost was found at altitudes between 1900 and 2000 m a.s.l. on north-facing talus slopes, which are very much shaded year-round. Slopes there are of about 25–35°, and there upper parts have perennial snow patches with nival niches developing (Rączkowska 1997). The slopes are modelled by gravitational processes like falling, rolling or sliding, as well as by the accumulation of debris. The effects of debris-flow activity are visible in the eastern part. There are no other distinct periglacial forms beside the nival niches (Fig. 5).

THE KASPROWY WIERCH SITE

Kasprowy Wierch summit was indicated by W. Dobiński (1997) as an area whose climatic features are such as to sustain a presumption that permafrost is even now present. The results of geophysical soundings confirm this, and point to their being two permafrost layers (Dobiński 2004).

The slopes of Kasprowy Wierch are smooth and debris mantled, sustaining the vegetation cover of alpine meadows (Fig. 6). Coarse, angular boulders are significant components of slope covers. The relief in the area was analysed on the basis of a morphodynamic map of the Kocioł Gąsienicowy basin (Fig. 7), as devised within the framework of 1:1000-scale geomorphological mapping (Rączkowska 1999). The Kocioł Gąsienicowy basin is located on the north-eastern slope of Kasprowy Wierch. Bare crystalline rocks visible at the very summit are not portrayed on the map. Patches of relict blockfields are typical features, especially of the western slopes of Kasprowy Wierch, though small blockfields also occur on the eastern slopes (Fig. 6). Periglacial processes, mainly solifluction, dominate on the upper parts of slopes, while soil and debris creep prevail in the remaining parts. It is contended that both processes model slopes

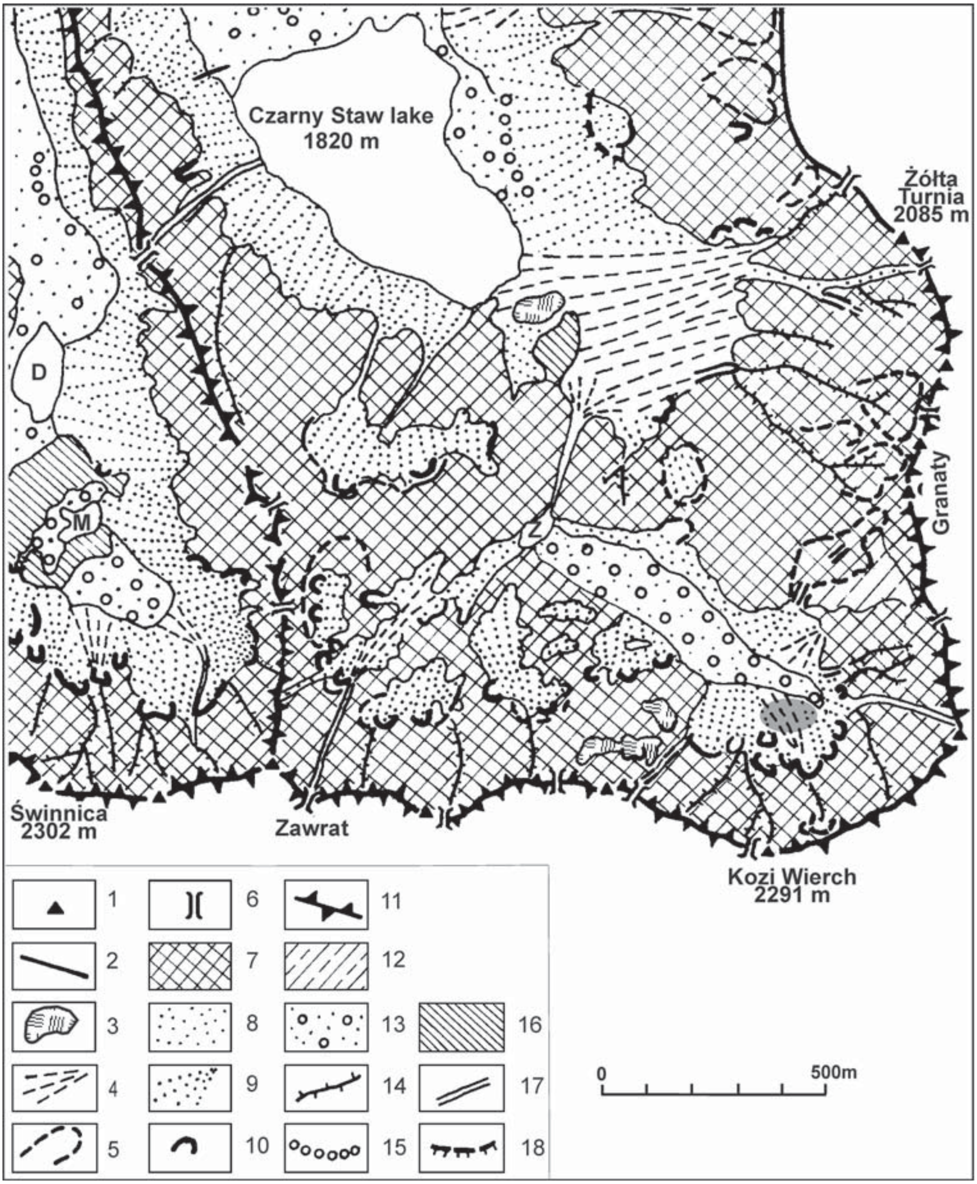

Figure 5. Geomorphological sketch of the upper part of the Sucha Woda Valley. Permafrost was found on the north-facing talus slope below the Kozi Wierch summit.
1—summit, 2—rounded ridge, 3—long-lasting snow patch, 4—alluvial-talus slope or cone, 5—relict nival niche, 6—pass, 7—rockwalls and rocky slope, 8—debris slope, 9—talus cone, 10—active nival niche, 11– knife-like ridge, 12—debris-mantled slope partly covered by vegetation, 13—glacial drift deposit, 14—chute, 15—moraine ridge, 16—roche moutonée, 17—debris flow gully, 18—protalus rampart; D—the Długi Staw lake, Z—the Zadni Staw lake (after Rączkowska 1997, modified).
Grey ellipse denotes area with permafrost found by S. Kędzia (2004).

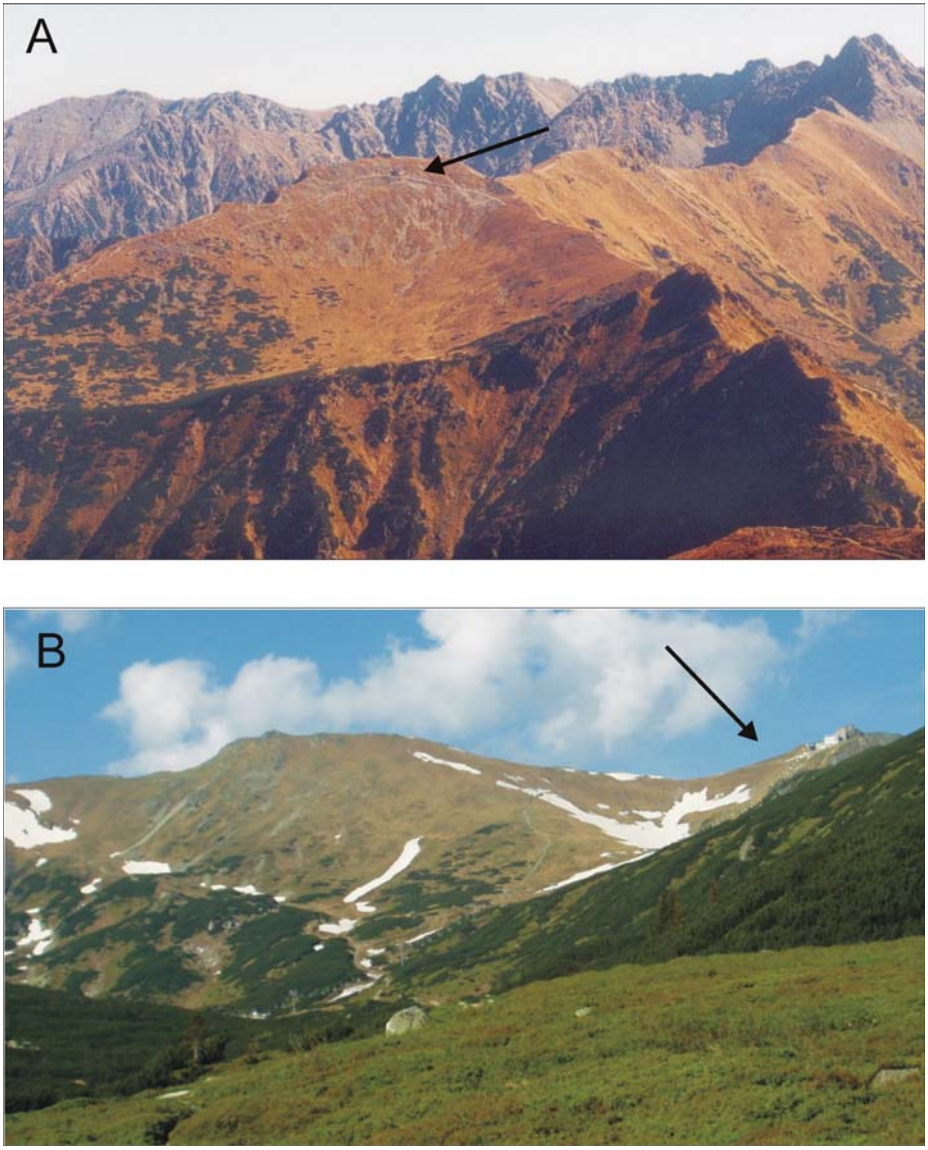

Figure 6. Slopes of Kasprowy Wierch (1987 m a.s.l). A—western, B—eastern. Arrows show summit, on which permafrost is said by W. Dobiński (1997, 2004) to occur (by Z. Rączkowska).

intensively. Rates of movement, evaluated on the basis of measurements for a few years are of 0.0–5.3 cm yr^{-1}. Solifluction has only resulted in the generation of microforms such as terracettes. Their development is not linked with permafrost, unlike that of the large solifluction lobes, which are even treated as geomorphic indicators of permafrost, if not very substantial ones (Etzelmüller *et al.* 2001). Unfortunately, solifluction lobes

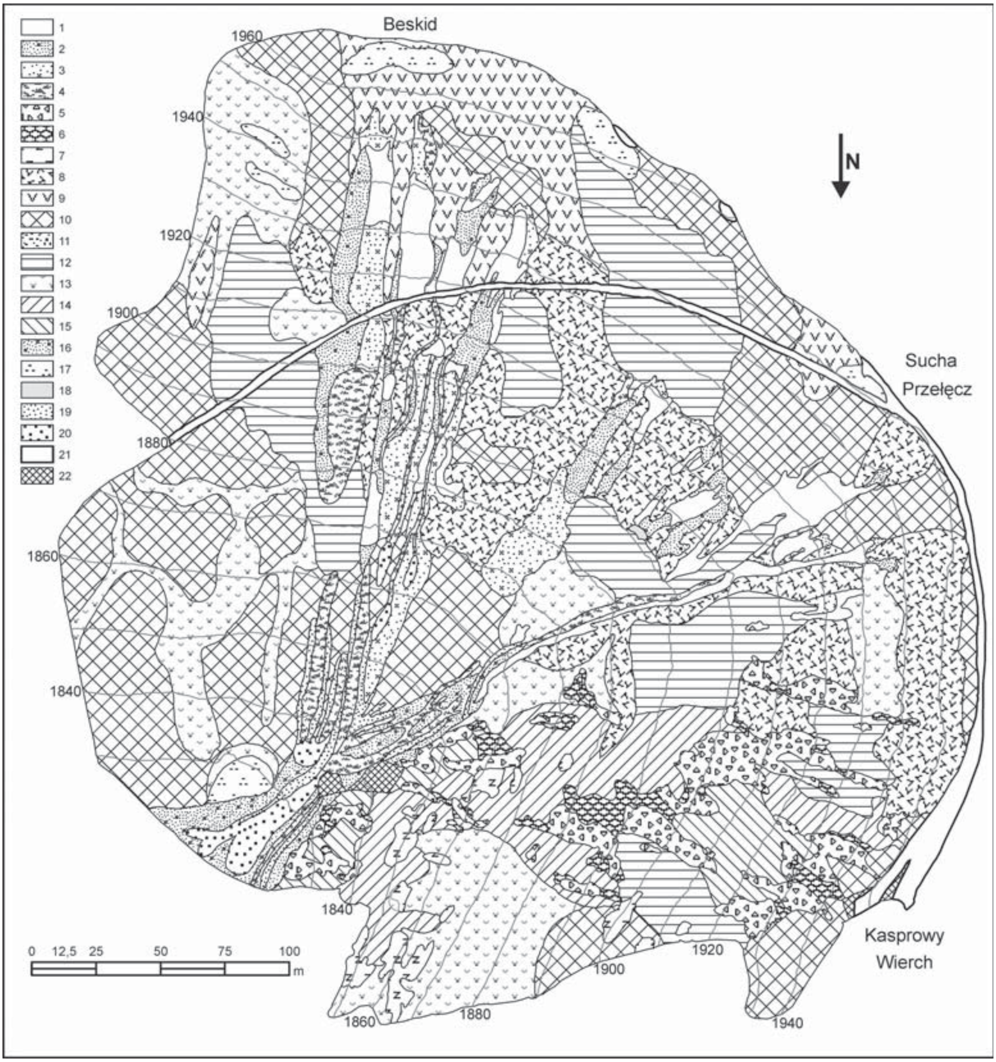

Figure 7. Morphodynamic map of the Kocioł Gąsienicowy area.
1—slope modelled by erosional processes, mainly sheetwash and rill erosion (I), 2—slope modelled by erosional processes, mainly sheetwash and rill erosion, with some sparse grasses, (II), 3—slope modelled by erosional processes, mainly sheetwash and rill erosion, partly stabilized by vegetation (III), 4—slope modelled by erosional processes, mainly sheetwash and rill erosion, stabilized by vegetation (IV), 5—slope modelled by creep of weathering cover (II), 6—slope with block cover modelled by creep of soil and vegetation layer over blocks (II), 7—slope with block cover modelled by creep of vegetation layer over blocks (III), 8—slope modelled by soil creep and solifluction (III), 9—slope with terractes modelled by solifluction (II), 10—slope modelled by solifluction (II), 11—nival niches modelled by nivation, 12—smoothed, stable slope modelled by deflation (III), 13—inactive slope (IV), 14—slope stabilized by *Pinus mugo* (IV), 15—stable slope with block cover (IV), 16—stable slope with block cover occupied by vegetation during period 1975–1995 (IV), 17—slope with fresh accumulation at debris flow levee (II), 18—stable slope with accumulation at debris flow levee (IV), 19—stable slope with accumulation at debris flow tonque (IV), 20—accumulation on alluvial cone and plane (IV), 21—slope with anthropogenic erosional and accumulational processes (II), 22—buildings. Roman numerals I to IV denote intensity of processes from very low to high.

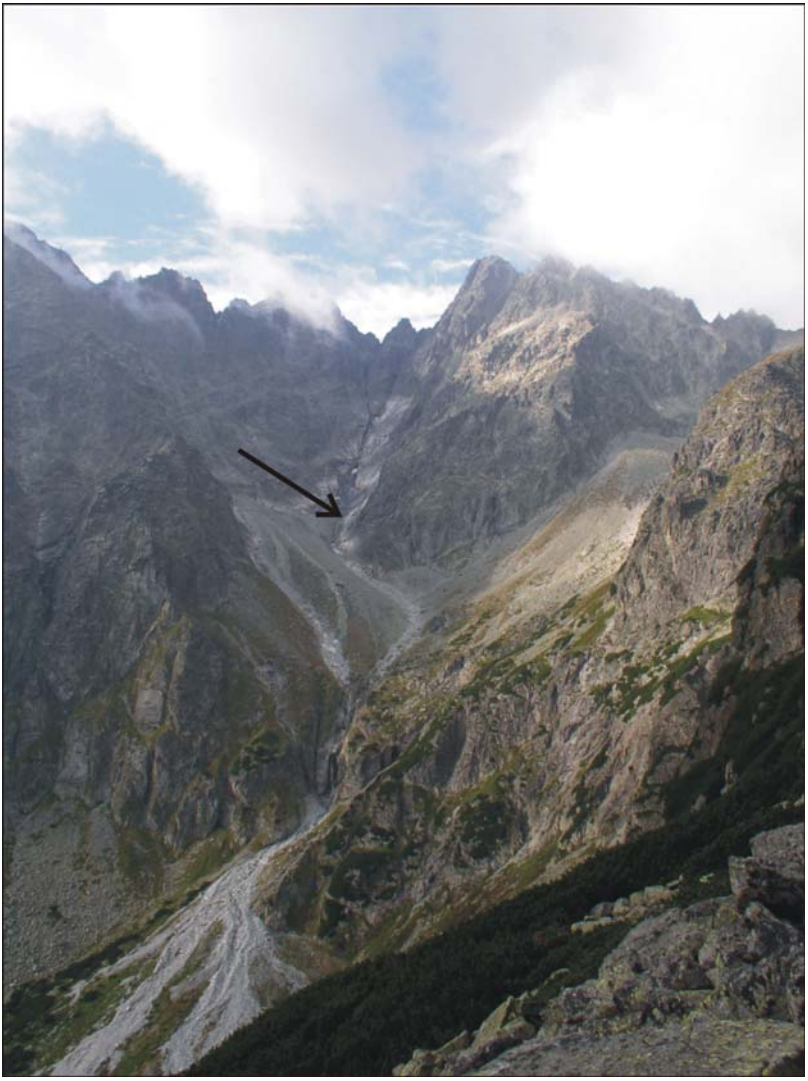

Figure 8. The Medena Kotlinka Valley with glacierette at the bottom indicated by the arrow (by B. Gądek).

are not found in the area. Apart from solifluction, nivation in the vicinity of the snow patches persisting until June that are present in small hollows on the upper parts of slopes has contributed to the development and simultaneous preservation of the above hollows. Erosional niches and fresh accumulation in the form of alluvial cones account for just small fragments of slopes, though erosional processes are presently involved in the significant modification of slope morphology. Overall, forms indicative of the presence of permafrost have been not found in association with slope morphology.

THE MEDENA KOTLINKA VALLEY SITE

The Medena Kotlinka Valley is a small, hanging, glacial cirque in the upper part of the Kežmarska Biela Voda Valley, in the Slovak part of the High Tatra Mts. (Fig. 8). Narrow and surrounded by c. 500 m high rockwalls, the valley opens up to the north. A small firn/ice body—a glacierette—rather than a perennial snow patch—is present at altitudes 2020–2350 m a.s.l. The glacierette is the largest form of this type in the Tatra Mts. A layer of ground ice was detected in the bottom of the valley, just to the right of the glacierette, at depths ranging from 2–5 m near the moraine ridge to 10 m at the talus cone. The presence of massive ground ice was evidenced by different geophysical methods (Gądek and Kotyrba 2003; Gądek and Żogała 2005). Moreover, ground ice outcropped at the surface even in an extremely hot summer (Fig. 2).

The analysis of slope morphology in the area under study was based on detailed geomorphological mapping at 1:1000 (Fig. 9). Gravitational processes, alluviation, avalanche activity and periglacial processes all model debris slopes at the bottom of the Medena Kotlinka Valley (Rączkowska 2005). Large alluvial cones and distinct talus cones occur at the valley bottom. Fresh debris flow gullies of various sizes dissect not only the alluvial cone but also those of gravitational and avalanche origin. Avalanche

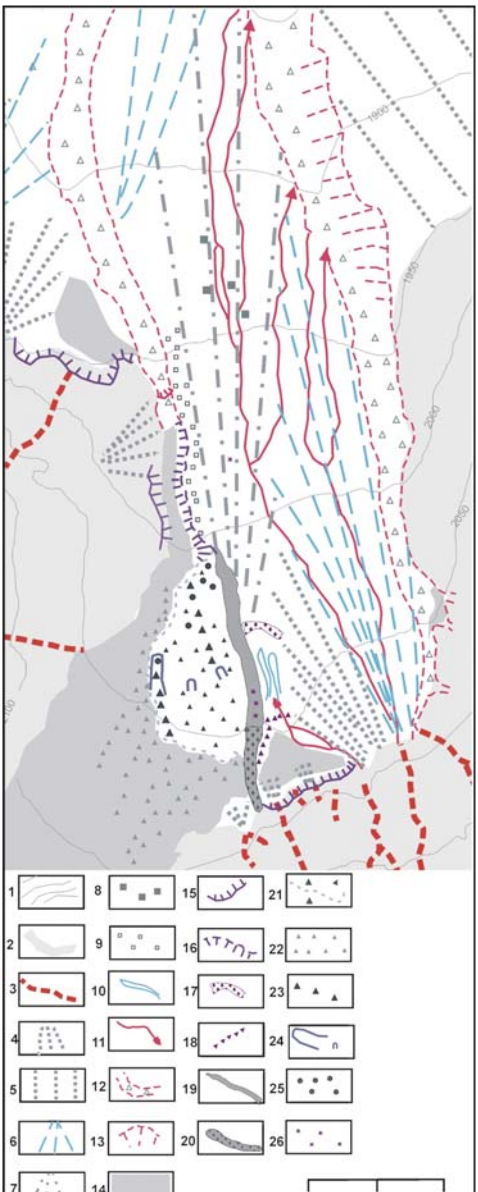

Figure 9. Geomorphological map of the Medena Kotlina Valley. 1—contour lines, 2—rockwall and rocky slopes, 3—chutes, 4—talus cones, 5—talus heaps, 6—alluvial cones, 7—debris slope modelled by avalanches, 8—rocky blocks (height >1 m), 9—accumulation of fresh debris, 10—debris flow tongues, 11—small gullies of debris flows (0.3–1.0 m depth, up to 3 m wide), 12—large gullies of debris flows (up to 6 m wide), 13—erosional edges, height >5 m, 14—perennial snow patches and tiny glacier (glacierette), state in summer 2003, 15—nival niches, 16—edges of nival niches cut in debris cover, 17—relict protalus ramparts, 18—active protalus ramparts, 19—ridge of frontal-lateral moraine, formed in the Little Ice Age, at present modelled by frost action, 20—active ridge of frontal-lateral moraine, 21—surface moraine, 22—periodical surface moraine, 23—moraine ridges within surface moraine, 24—lobes of free solifluction, 25—thermokarst hollows with diameters from 1 to a few metres, 26—patterned ground.

activity significantly affects the whole valley bottom through the transportation and accumulation of large amounts of debris and the simultaneous smoothing of the debris slope. Periglacial processes are at present resulting in the development of landforms. Nivation niches of different types and sizes with a distinct protalus rampart up to 1–1.5 m high are most common. Small polygons, 0.4–0.5 m in diameter, and miniature structural soils also exist, especially on the 7–10 m high moraine ridge rising along the right side of the glacierette. On the surface of the ridge it was possible to observe small crevasses (a few centimetres deep) related to ground thawing, as well as a large (> 1 m diameter) boulder that had plunged into the moraine deposits and been split into two pieces by frost weathering. In this case the occurrence of latest forms or phenomena can be linked with processes in the active layer, as the presence of permafrost in the morainic ridge was documented by the two-year BTS study (Gądek and Kędzia 2006). Morphological indicators of permafrost presence are found on the surface moraine covering the right half of the glacierette with a layer more than 3 metres thick. Small thermokarst hollows and a relatively large lobe of free solifluction have developed there. The thermokarst hollows are one to a few meters in diameter and less than 1 m thick. The fresh debris lobes are over ten metres long and a few metres wide, and have fronts more than 1 m high. Aside from the surface moraine, similar forms have not been found in the Medena Kotlinka Valley, even in the area in which the existence of permafrost patches is stated. The situation thus resembles that in the Tatra Mountains as a whole.

The results of the morphological analysis reveal either a weak relationship or almost no relationship at all between slope morphology and the presence of permafrost in the area. This is also confirmed by the results of the studies on slope morphodynamics and slope cover texture. The nature of the disturbances to the lines marked on the debris combine with the distance travelled by debris to point to frost creep or avalanche activity (Rączkowska 2004, 2007). The analysis of clast microfabric on the debris slope also indicates the possibility that permafrost is only present in surface moraine (Fig. 10). The azimuths of the clast longest axes and directions of clast axis dip vary from 0 to 360°. The slope of the superficial moraine generally faces north and is inclined to around 20°. The angle of inclination of debris ranges from 15° up to 90°. Clast microfabric on the debris slope, occurring below

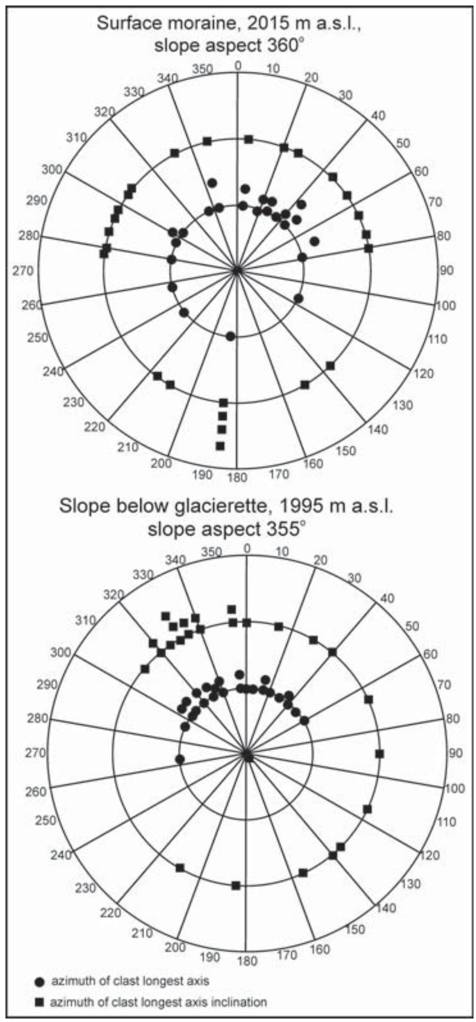

Figure 10. Microfabric of clast on a debris slope in the Medena Kotlinka Valley (after Rączkowska 2007, modified).

the glacierette, indicates that the debris slope is mainly modelled by avalanches. There is a definite difference in clast microfabric on the slope below the glacierette and on the superficial moraine. Azimuths of the clasts longest axes are between 260° and 60°, and are in good agreement with the slope orientation. Axes of 25% of clasts on this slope are inclined in opposite directions. This suggests that, while the slope is modelled by debris creep or maybe frost creep, it is not modelled by permafrost creep.

DISCUSSION AND CONCLUSION

The analysis of the slope relief at the three discussed permafrost sites does not provide unequivocal evidence as to the presence of permafrost and its influence on slope morphology or morphodynamics. Neither permafrost-indicative landforms nor large periglacial forms were found, other than the fragments of glacierrette covered by surface moraine.

Relatively large sorted polygons (the best developed examples anywhere in the Tatra Mts.) occur in the Mengusovska Valley, at an altitude similar to that at which permafrost has been found in other locations (i.e. 1950 m a.s.l.). However the polygons are zonal forms (Fig. 11), their contemporary activity being related to seasonal and diurnal freeze–thaw cycles, as documented by experimental studies (Rączkowska 2007). Active, indicative permafrost landforms are not found at other permafrost sites in the Tatra Mts.

Generally, the periglacial landforms developing in the Tatra Mts. are small, and only

Table 2. Occurrence of active periglacial forms in the high mountains of Europe.
(after Z. Rączkowska 2007, modified).

Forms	High Tatras	Western Tatras	Belanske Tatras	Scandinavian Mountains	Alps	Retezat	Fagaraš
Blockfields	–	–	–	±	+	–	–
Rock glaciers	–	–	–	+	++		–
Ice-cored moraine	–	–	–	+	±	–	–
Palsas	–	–	–	+	–	–	–
Non-sorted polygons	–	–	–	+	±	–	–
Sorted polygons	+	–	–	++	++	–	–
Sorted circles	+	–	–	++	++	±	±
Sorted strips	+	+	+	++	+	–	–
Miniature patterned grounds	+	+	+	+	+	+	+
Thufurs	+	+	+	+	+	+	+
Solifluction lobes	±	+	+	++	++	±	+
Solifluction sheets	–	–	–	+	–	–	–
Solifluction garlands	±	+	+	+	+	+	+
Terraccettes	+	++	++	+	++	+	++
Ploughing blocks	+	+	+	++	+	+	+
Gelideflation forms	–	+	±	+	+	–	–
Nival niches	+	+	+	++	+	+	+
Protalus ramparts	±	–	–	+	+	±	±

”**–**”—absent, ”**±**”—sporadic, ”**+**”—common, ”**++**”—very common

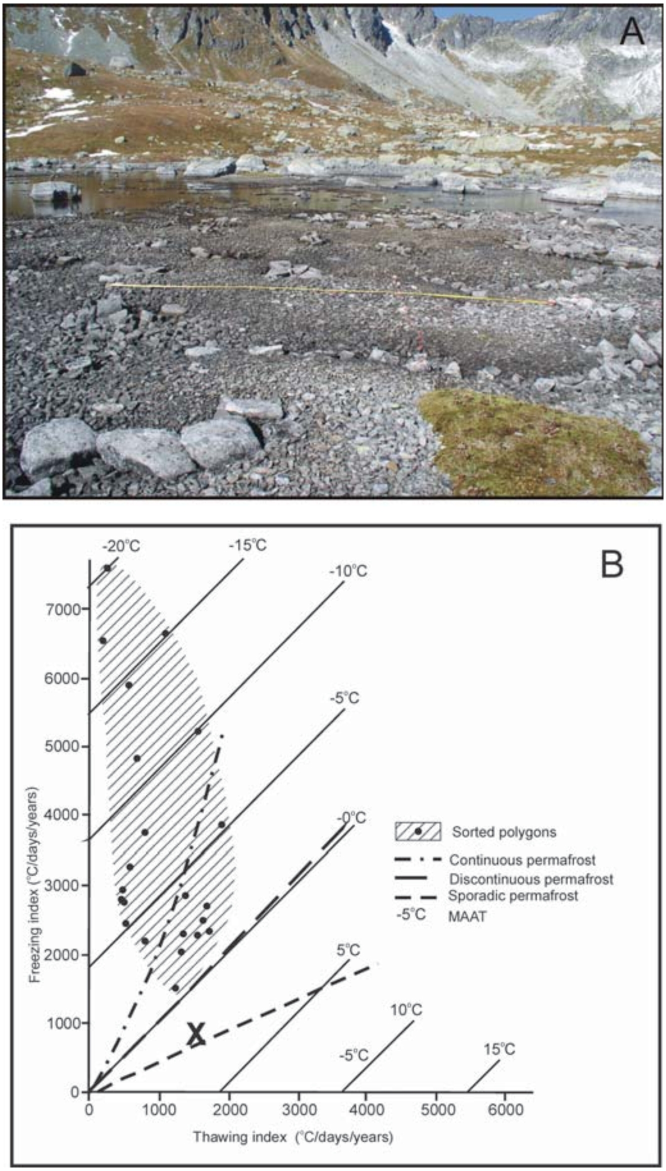

Figure 11. A. sorted polygons in the Hińczowe Oko lake at 1950 m a.s.l., Mengusovska Valley, south slope of the High Tatras, Slovakia (Photo. Z. Rączkowska);
B. Zonality of patterned ground in the Mengusovska Valley, on the diagram of S. Harris (1981);
X—sorted polygons at the Hińczowe Oko lake.

occasionally large, as is the case for mountains in which the presence of permafrost was only suggested, rather than evidenced, as in the Retezat or Făgraš Massifs of the southern Carpathians. The types of periglacial forms are less diversified/numerous than in the mountains where permafrost is more widespread and comprises a variety of forms from isolated patches to a continuous layer, as in the Alps or the Scandinavian Mountains (Table 2).

The high activity of geomorphic processes observed in the areas with permafrost is likely to be another reason for the absence of geomorphic evidence regarding the presence of permafrost. The activity of avalanches and gravitational processes seems to efface the results of periglacial morphogenesis on debris slopes.

It is the nature of permafrost that causes related active landforms to be absent from the Tatras, as well as other mountain regions in which similarly developed permafrost can be found. According to Kędzia (2004) and Dobiński (2004), the patches of permafrost are small. Therefore, they seem inadequate to set permafrost creep in motion, for example. Only a loosening of soils can be stated. The relatively great (2–6 m) depth of the active layer might also help explain why permafrost does not affect slope morphology.

The absence of active permafrost-related forms from the Tatra Mts. is not exceptional. There are other mountain regions with isolated patches of permafrost occurring or suggested to occur (i.e. Urdea 1992; Kern *et al.* 2004; Gude *et al.* 2003; Julián and Chueca 2007; Zacharda *et al.* 2007) in which no distinct geomorphic evidence of the presence of permafrost is reported. Thus, geomorphic forms are not particularly useful identifiers of the existence of permafrost in mountains in which the latter forms nothing more than isolated patches. Even the presence of periglacial zonal forms could not always offer a basis for a determination of the extent of permafrost in given areas, as S. Harris (1981) suggested.

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