

On the response of rockglacier creep to surface temperature increase

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Received 18 August 2005; accepted 21 July 2006

Available online 14 September 2006

Abstract

Besides its thermal characteristics creeping mountain permafrost is substantially defined by its kinematics. Due to the – in general considerable – ice content of rockglaciers, their dynamics respond sensitively to climate forcing. Questions arise how rockglaciers react to the current or recent climatic changes, and what the further consequences of such reactions could be. Using a one-dimensional thermo-mechanically coupled numerical model we simulate the potential response of rockglacier creep to a change in surface temperature (Section 2). It turns out that variations in temperature could indeed affect rockglacier creep in the currently observed order of magnitude. Other influences, however, clearly act as well. Among these, the occurrence and complex influence of liquid water in the frozen material might be the most important factor for permafrost close to 0 °C, though difficult to model. As a next step in this contribution, we plot globally observed rockglacier speeds against mean annual air temperature (Section 3). In fact, air temperature can be statistically identified as a major factor determining rockglacier speed. The remaining scatter clearly points to other influences such as slope, debris content, column thickness or liquid water. In a further step, we summarize current monitoring results on rockglacier speed (Section 4). A surprisingly large number of Alpine rockglaciers showed an increase in speed during recent years. This large number points to other than solely local influences, but rather to some regional-scale impact such as the observed increase in air temperatures. Our monitoring and modelling work clearly shows that rockglaciers with ground temperatures close to 0 °C creep in general faster than colder ones. Furthermore, our findings suggest that the creep of permafrost close to 0 °C is more sensitive to thermal forcing than the creep of colder one. From this, we conclude that increasing rockglacier temperatures may lead to a marked, but both spatially and temporally highly variable speed-up, before a significant loss of ice content by melt-out is able to reduce the deformation rate of the frozen mass towards its entire deactivation. By means of three scenarios, we exemplify the possible consequences of an increase in rockglacier temperature and subsequent acceleration: (1) increasing sensitivity of rockglacier creep to seasonal influences, (2) activation of so far stable frozen debris slopes, and (3) rockglacier destabilization.

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Keywords: rockglacier; creep; mountain permafrost; climate change; warming; slope stability

1. Introduction

Ice–debris mixtures under mountain permafrost conditions are able to form prominent creep streams — so-called rockglaciers. These features creep with surface speeds in the order of centimetres to several metres per year, and are thousands to ten-thousands of years old. Thus, they are important factors of landscape evolution in

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cold mountains, and important geo-indicators for mountain permafrost and related (paleo-)climatic conditions (e.g. Haeberli, 1985; Martin and Whalley, 1987; Whalley and Martin, 1992; Barsch, 1996; Humlum, 1998; Haeberli et al., 1999; Frauenfelder and Kääb, 2000; Haeberli, 2000; Kääb et al., 2003; Frauenfelder et al., 2005).

The deformation magnitude of the frozen bodies can change at temporal scales of millennia to seasons. Speed variations at sub-seasonal or diurnal scales are likely to occur, but might not have a relevant influence on the long-term evolution of rockglaciers and alpine landscape considered here. The following rockglacier speed variations have been investigated so far:

- (1) Millennia-scale speed variations represent fundamental stages in the rockglacier transport system, such as the origin of a rockglacier itself, impacts from glaciation, or melt-out of the ice content. Relictification of rockglaciers is in many cases obviously related to paleoclimatic changes (e.g. Brazier et al., 1998; Frauenfelder and Kääb, 2000).
- (2) Speed variations (including de- or re-activation) with frequencies in the order of centuries or several decades can result from general spatio-temporal changes in boundary conditions such as changes in material and water/ice supply or thermal regime, or from terrain topography (Barsch, 1996; Frauenfelder, 2005). Comparison of present-day flow fields of rockglaciers with their current shape – that cumulatively reflects their dynamic history – clearly shows such past temporal variability (Frauenfelder and Kääb, 2000).
- (3) Photogrammetric and geodetic monitoring series of up to several decades of duration suggest that velocity changes of mountain permafrost creep at pluriannual time-scales may be due to variations in weather conditions and/or climate conditions (e.g. Zick, 1996; Kääb et al., 1997; Schneider and Schneider, 2001; Kaufmann and Ladstädter, 2002; Roer et al., 2005b) (see Section 4).
- (4) At least on shorter time-scales, the sensitivity of the creep of frozen slopes to external climate forcing is strongly dependent on individual internal conditions of rockglaciers, as clearly shown by observations of seasonal velocity variations. Geodetic measurements in the Swiss Alps yield seasonal speed variations from nearly 0% to several tens of percents (Haeberli, 1985; Barsch, 1996; Arenson et al., 2002; Kääb et al., 2003) (more in Section 4).

The above mentioned frequencies of speed variability form a continuum of different changes overlaying

each other. While long-term variations are mostly relevant to landscape evolution, short-term changes provide e.g. insights into the creep mechanisms of rockglaciers and frozen ground.

Recently, questions arose whether and how the atmospheric warming, which is currently observed in many mountain regions (McCarthy et al., 2001; Böhm et al., 2001) influences the creep of perennially frozen debris. On the one hand, this issue is of large scientific interest, for instance, in view of climate change projections or landscape evolution. On the other hand, geotechnical changes of frozen slopes have notable impact on slope stability, and are thus of interest for dealing with applied problems such as construction stability, debris-flow hazards, or rockfall hazards (Haeberli, 1992a; Haeberli et al., 1997; Arenson and Springman, 2000; Noetzli et al., 2003; Kääb et al., 2005a,b).

Some authors have already investigated the (potential) reaction of perennially frozen slopes to ground temperature variations. Haeberli (1992b) developed a theoretical model on mountain permafrost reaction to climate change. Azizi and Whalley (1996) modelled the development of an ice-rich scree slope and found significant differences for ice temperatures of -5°C and -1.5°C . Davies et al. (2001) found minimum shear strength of frozen rock joints at temperatures between -1°C and 0°C using centrifuge modelling. Arenson (2002) showed a clear temperature dependency of the deformation of ice and ice-rich soils during triaxial tests. Further authors have collected various indications that the deformation of frozen debris or rockglaciers increases with increasing ground temperatures (e.g. Hoelzle et al., 1998; Irving, 2000; Rea et al., 2000; Kääb et al., 2002; Frauenfelder et al., 2003; Ikeda et al., 2003) or have proposed creep laws which include a dependency on ground temperature (for overview see Arenson, 2002; Ladanyi, 2003).

In this study, we investigate the potential influence of ground temperature increase on rockglacier creep using a simple numerical model (Section 2), a global set of rockglacier speed data (Section 3), a regional-scale compilation of pluriannual speed variations in the Swiss Alps (Section 4.1), and local-scale observations of seasonal to pluriannual speed variations on three individual rockglaciers, again in the Swiss Alps (Sections 4.2, 4.3 and 4.4). In the discussion (Section 5) these data are compared to each other.

2. Model

2.1. Model scheme

In this section, an initial estimation of the influence of surface temperature variations on permafrost creep is

numerically deduced by combining heat diffusion with a temperature-dependent rate factor for ice deformation. The model is constructed as follows (Fig. 1):

A body of pure ice is assumed. This assumption has little effect on the computed heat diffusion since the respective heat diffusivity coefficients for ice and rock are similar (Lee, 1970). The assumed isotropic heat diffusivity would, however, not apply for a significant air (or water) content of the frozen ground. Further, the ice/debris proportion certainly has an influence on the creep process. Among others, three important factors counteract: (i) the density of the mass and, thus, the basal stress which increases with increasing debris content. (ii) On the other hand, the amount of ice, which can be deformed between the rigid debris particles decreases with increasing debris content and with it the deformation of the mass (Azizi and Whalley, 1995). (iii) In addition, the influence on overall deformation from processes of ice/debris interaction increases with increasing amount of debris particles (e.g. Arenson, 2002; Ladanyi, 2003). These three effects are neglected here as they are considered not to alter the outcomes of the general sensitivity study intended in this investigation in a substantial way.

Water penetration and other advective mechanisms of heat transport are thus neglected. These processes could indeed play a significant role in relation to permafrost temperatures close to and at the melting point (Delaloye, 2004). Yet, the available knowledge about their possible effects is not sufficient for our modelling purposes (Davies et al., 2001; Arenson, 2002; Ladanyi, 2003). They might have non-linear (phase transition ice–water), highly individual consequences for different rockglaciers. The model presented is one-dimensional, i.e. it neglects any lateral influences as described, for instance, by Delaloye (2004).

Since the model in question aims at simulating variations over some years, the interplay between energy flux through the surface and geothermal heat flux is neglected too. As a consequence the temperature is kept constant at the base of the ice body. This procedure is deduced from measured thermal permafrost behaviour under seasonal temperature variations (Vonder Mühll et al., 1998; Vonder Mühll et al., 2003).

Model input is a variation in ground surface temperature. The latter ranges between $-3\text{ }^{\circ}\text{C}$ and $0\text{ }^{\circ}\text{C}$ for the model runs depicted in Fig. 2. Such thermal conditions are typical e.g. for the European Alps (Hoelzle et al., 1998; Vonder Mühll et al., 2003).

Heat diffusion in the ground is computed from

$$\frac{\partial T}{\partial t} = \kappa \frac{\partial^2 T}{\partial z^2} \quad (1)$$

(Carslaw and Jaeger, 1959), where T is temperature, t is time, κ is heat diffusivity of ice and z is depth. In the model, heat diffusion is calculated for layers with a finite thickness (e.g. 3 m).

For $T(t)$ obtained for each layer, a temperature-dependent rate factor A is interpolated. For this purpose an exponential curve is fitted into the A factors given by Paterson (1994) for various ice temperatures and the flow-law exponent $n=3$ (see also Eq. (3) for the parameter n):

$$A = 6.6e^{0.4 \cdot T} 10^{-15} [s^{-1} (kPa)^{-3}] \text{ for } T \leq 0 \quad (2)$$

(Cf. Azizi and Whalley, 1996; Arenson, 2002; Ladanyi, 2003 for other descriptions of temperature-dependent viscosity of frozen ground). An exponential dependency of rate factor from temperature is theoretically expected from the Arrhenius equation, at least for cold ice temperatures (Paterson, 1994; Arenson, 2002). A can then be written as a function of time and depth ($A(t, z)$). An exponential-like dependency of deformation from temperature seems also to apply for rockglaciers (Davies et al., 2001; Arenson et al., 2002; Kääb et al., 2002; Frauenfelder et al., 2003) (see also Section 3).

The rate factor A is increased for selected layers to simulate shear horizons as found in most rockglaciers with borehole deformation measurements available (called “soft layer” in Fig. 2) (see Wagner, 1996; Arenson et al., 2002). Thus, the model only simulates ice deformation, though with optional soft layers, but neglecting any kind of basal processes such as sliding. Borehole measurements suggest that sliding *sensu strictu* might play a less important role for rockglaciers compared to e.g. glaciers (Arenson et al., 2002).

$A(t, z)$ is then introduced into Glen’s flow law for an infinite slab (Paterson, 1994):

$$1/2 \frac{\partial v_x}{\partial z} = -A(\rho g \sin \alpha z)^n \quad (3)$$

Eq. (3) is then integrated over all model layers. For instance, for an A that is constant with depth the surface velocity would be

$$v_x^s = \frac{2A}{n+1} (\rho g \sin \alpha)^n h^{n+1} \quad (4)$$

where v_x^s is the horizontal surface velocity, ρ is the density of ice, g is the acceleration due to gravity, α is the surface slope, h is the total column thickness, and z is the depth-axis (perpendicular to the surface). Slope is set at 12° ($\sin \alpha = 0.2$), but 20° ($\sin \alpha = 0.35$) was also tested. The flow-law exponent n is set at 3, but model

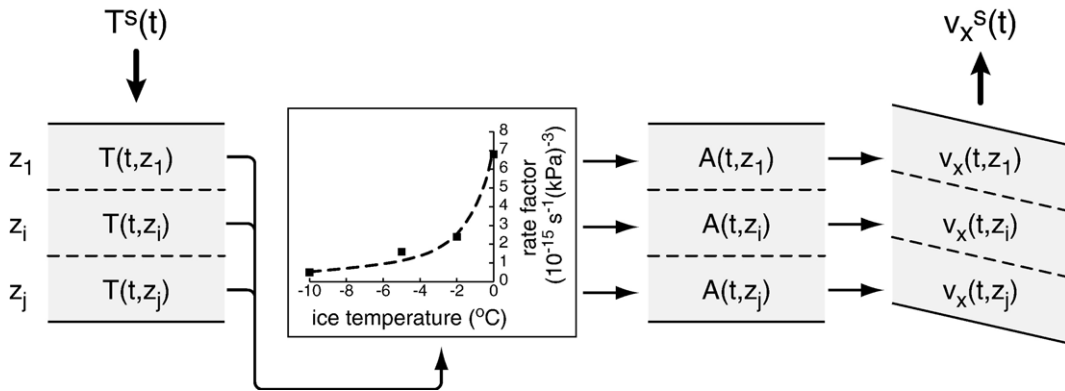


Fig. 1. Numerical scheme of an inclined ice column for simulating changes in surface speed due to variations in surface temperature. The temperature of individual layers (T) is computed using heat diffusion, and a temperature-dependent rate factor (A) is then estimated empirically for each layer. Integration of the deformation of the individual layers (v_x) then gives the total speed at the surface (v_x^s).

runs with $n=1$ were also performed (Wagner, 1996; Kääb, 2005).

The final model output gives variations of surface speed depending on surface temperature variations (Fig. 2).

Inserting Eq. (2) in Eq. (4) gives

$$v_x^s = c_1 \cdot e^{0.4 \cdot T} [ma^{-1}] \quad \text{for } T \leq 0 \quad (5)$$

where c_1 ranges, for instance, from 0.04 for $h=15$ m and $\alpha=12^\circ$, to 4.5 for $h=30$ m and $\alpha=20^\circ$. Note that Eqs. (4) and (5) assume A to be constant with depth, which is not the case in our model. The latter equations are rather given for comparison with field data (see Section 3).

2.2. Model results for sinusoidal temperature variations

A first set of model runs was performed using sinusoidal surface temperature variations of one-year and five-year cycles, and a column height of 15 m. Two model runs were computed for a surface temperature variation with a one-year wavelength. Run (a) with an average temperature of -1.5°C throughout the entire column and with a variation magnitude of $\pm 1.5^\circ\text{C}$ at the surface (Fig. 2a), and run (b) with an average ice temperature of close to 0°C (Fig. 2b). Run (a) simulates thermal conditions similar to those for Murtèl rockglacier (for example Vonder Mühll et al., 2003). Run (b) simulates thermal conditions for rockglaciers close to 0°C . Both runs aim to cover the range of climate conditions found so far for rockglaciers in the European Alps. For run (b) the temperature of the lower boundary was kept constant at 0°C . Temperature at the surface was varied as a sine-curve around $T=0^\circ\text{C}$ with an amplitude of -1.5°C . Temperatures above 0°C were cut off and set to 0°C . The aim of

this procedure is to simulate ice-melt at the permafrost table.

The modelled heat diffusion with depth is comparable to that measured in boreholes (Vonder Mühll et al., 1998; Vonder Mühll et al., 2003). For runs (a) and (b) the temperature maximum reaches a depth of 15 m with a delay of approximately half a year compared to the surface.

The modelled surface speeds depend, among other factors, significantly on the viscosity of the soft layer, the vertical position and the thickness of the soft layer, and the flow-law exponent n . Thus, in each of the following model results a range of speed variations is given, rather than exact numbers. This range reflects increases of A by the factors 5 and 10 for the soft layers, flow-law exponents of $n=3$ and $n=1$, and surface slopes of 12° and 20° . For the runs depicted in Fig. 2 $n=3$ and a surface slope of 12° was set. Also, tests with slight changes of the overall A have been made. Accordingly, the rate of deformation changes significantly with these model parameters. As a consequence it is important to interpret the model results only qualitatively and only in comparison of the different model runs to each other.

For model run (a) (Fig. 2a) average speeds of the runs with a soft layer are significantly higher than without such a layer. Average speeds are also higher with a soft layer at greater depth. The resulting variations in surface speed are about 3–11% (total range) of average speed if no soft layer is assumed. Surface speed varies by approximately 2–4% if a soft layer is introduced for a depth of 12–15 m, and by about 5–13% if the soft layer is in 6–9 m depth. For run (b) (Fig. 2b), i.e. ice of 0°C with penetrating winter coldness, speed variations are on the order of 5–10% without soft layer, 4–5% for a

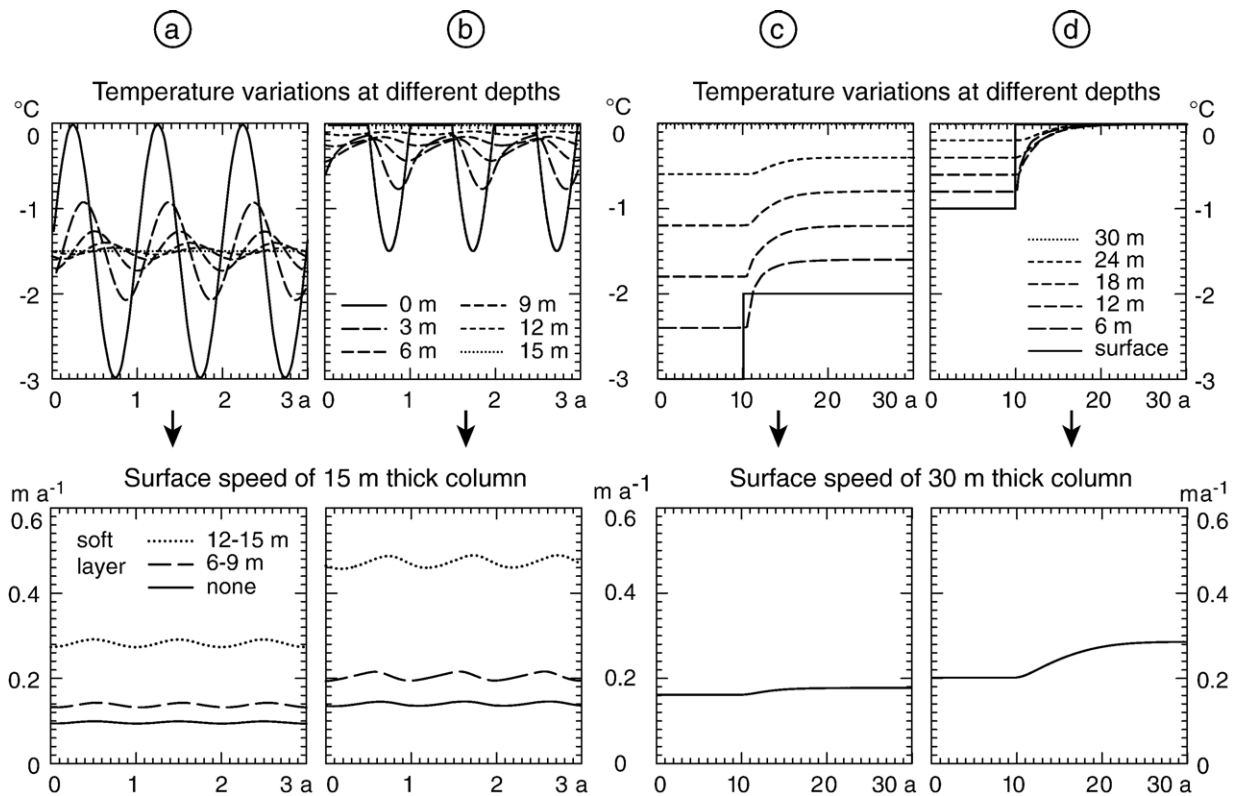


Fig. 2. Results from four runs of the model presented in Fig. 1. Left (a and b): surface temperature variations with a one-year cycle (i.e. seasonal variations); (a) cold permafrost; (b) permafrost close to 0 °C. Right (c and d): temperature steps of +1 °C. Upper row: modelled temperatures at selected depths; lower row: modelled response of surface speed to temperature variations. In runs (a) and (b) different scenarios with soft layers within the column were computed. (See text for detailed description).

12–15 m deep soft layer, and 9–12% for a 6–9 m deep soft layer.

In addition, a five-year cycle was introduced in the model in order to simulate the effect of one or several especially warm or cold years, or years with different snow depths or timing (inter-annual changes). Two runs, (e) and (f) (not depicted), were computed with the same temperature variations as above runs (a) and (b), but with a five-year cycle. For a temperature variation with a five-year wavelength, the delay of temperature maximum at 15 m depth increases to about 1 yr. For run (e), surface speeds vary by about 17–20% without soft layer, 11–14% for a 12–15 m deep soft layer, and 23–30% with a 6–9 m deep soft layer. For run (f), i.e. ice of 0 °C, the respective numbers are 40%, 35–37% and 42–47%.

2.3. Model results for a step-wise temperature increase

For the five-year cycle the surface temperature variations are still too fast in order to allow for the ground temperatures to fully adjust to the temporal changes. The

surface temperatures have already decreased significantly before the temperatures at greater depths are at their maximum. Thus, the body cannot reach its potential maximum temperature and speed. Therefore, a temperature step of +1 °C (cf. observations by Böhm et al., 2001) was introduced in the model in order to investigate the behaviour of a creeping mass, which thermally adjusts completely to a change in surface temperature (Fig. 2c and d). Model runs were performed for cold and for “warm” permafrost conditions. For all runs the temperature at the lower model boundary was fixed at 0 °C. The depicted runs are for a 30 m thick column.

The surface speeds rise by roughly 10% when the surface temperature is increased from –3 °C to –2 °C (Fig. 2c). This percentage order of magnitude applies both for 15 m and 30 m thick columns and also for higher speeds when soft layers are introduced like in runs (a) and (b). For a step-wise increase in surface temperature from –1 °C to 0 °C (Fig. 2d) the surface speed increases by about 40–50% for all model types with 12° slope. For 20° slope the increase in speed is up to 100%. The time it takes

for the columns to fully dynamically adjust to the new surface thermal conditions is roughly 5–7 yrs for 15 m thickness and about 15–25 yrs for 30 m thickness.

3. Rockglacier speed and mean annual air temperature

To some extent, rockglacier speed under conditions of different mean annual air temperatures (MAAT) can be viewed as an analogy for speed changes expected due to MAAT changes. Fig. 3 shows a scatter plot of measured rockglacier maximum speeds as a function of measured or estimated MAAT at the individual sites. The data cover sites in North America, Svalbard, Greenland, and the European Alps. Reference to data sources is given in Frauenfelder et al. (2003).

Rockglacier speed is a function of a number of factors such as ground temperature, slope, thickness of the deforming layer, marginal friction, density, debris content and their distribution, ice softness, layering, water content, etc. (cf. also Eq. (3)). Therefore, caution is needed for interpretation of Fig. 3:

- (i) Due to the above factors the speed at individual points on a rockglacier ranges from zero to the

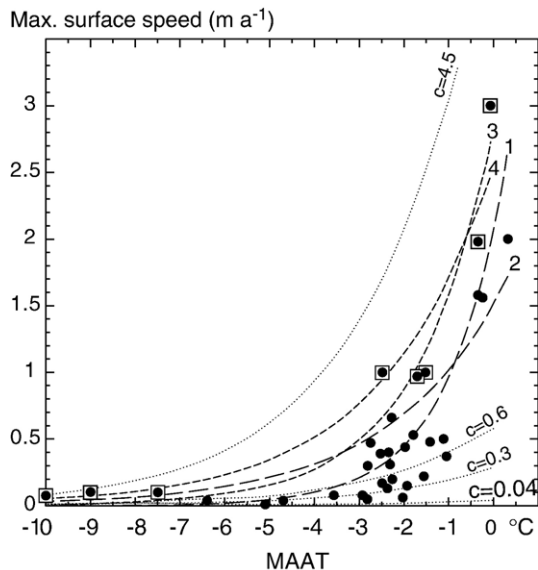


Fig. 3. Maximum surface speed of a global sample of rockglaciers as a function of mean annual air temperature (MAAT) at the rockglacier front (black dots). For data sources see Frauenfelder et al. (2003). Curves 1 and 2 are exponential fits through all points, curves 3 and 4 are exponential fits through the upper maximum points only (points marked with rectangles). Curves 1 and 3 are of the form $a \cdot e^{(b \cdot \text{MAAT})}$, curves 2 and 4 are of the form $c \cdot e^{(0.4 \cdot \text{MAAT})}$. The dotted curves are also of the form $c \cdot e^{(0.4 \cdot \text{MAAT})}$ but with different c values. (For factors a and c , and exponent b see text).

maximum of the specific rockglacier. At sections with maximum surface speed an “optimal” factor combination acts in terms of speed. Thus, we preferred to choose maximum speeds rather than average values (or any other statistical measures).

- (ii) Similarly, due to the above factors an individual rockglacier might not reach its potential maximum speed given by its ground temperature. Insufficient slope, thickness, ice content, etc., decrease its actual speed as compared to the theoretical maximum speed. The vertical scatter (i.e. scatter in y -direction) observed in Fig. 3 is a consequence of these individual combinations of rockglacier speed factors. Most of these factors are unknown for almost all rockglaciers considered in Fig. 3 (exceptions see Potter et al., 1998; Arenson et al., 2002). As a consequence, we performed also a statistical analysis considering only the maximum rockglacier speed for a certain MAAT section (i.e. the upper boundary of the scatter; see rectangle points in Fig. 3; i.e. the “maximum range of the maximum speeds”).

- (iii) The MAAT is by far not equivalent to the mean annual ground temperature (MAGT) or mean annual ground surface temperature (MAGST) which are more directly related to the actual ground temperature at a certain depth, which in turn influences the actual rockglacier deformation. Using the MAAT as a proxy for ground temperature neglects important factors of the surface energy balance such as snow cover or radiation (e.g. Stocker-Mittaz et al., 2002). However, for most rockglaciers of our sample MAGT or MAGST data are not available. A horizontal scatter (i.e. uncertainty in x -direction) for individual points in Fig. 3 on the order of 1 °C or more can thus be well expected from the differences between MAAT and MAGT for individual rockglaciers. Thereby, a general (or average) offset between MAAT and MAGT affecting all rockglaciers of our population does not so much influence our statistical analysis, but rather do the second-order differences of this temperature difference for individual rockglaciers, which result in increased statistical scatter.

Following the above considerations we statistically estimate several exponential curve fits for the points in Fig. 3. While such an exponential function makes much sense towards cold ground temperatures, it does certainly not apply for temperatures around 0 °C e.g. due to the phase transition from ice to water (cf. Section 2). Two curve fits were computed, each for the entire

sample, and for the upper-boundary points (rectangles in Fig. 3):

$$v_{x \max}^s = a \cdot e^{b \cdot \text{MAAT}} [m \ a^{-1}] \quad (6)$$

$$v_{x \max}^s = c_2 \cdot e^{0.4 \cdot \text{MAAT}} [m a^{-1}] \quad (7)$$

In Eq. (7) an exponent of 0.4 was chosen in order to be able to compare the curve fit to Eq. (5). For the entire sample the curve parameters are: $a=2.0$, $b=0.8$, and $c_2=1.5$. For the upper-boundary points the parameters are: $a=2.8$, $b=0.6$, and $c_2=2.5$. The statistical error of the fits is in the range of 10%. As mentioned above, the values for c_2 can be compared to the according values for c_1 in Eq. (5), falling within the same range of values (c_1 is between 0.04 and 4.5 in the above assessment for Eq. (5); see also dotted curves in Fig. 3).

Eqs. (6) and (7) can be used to estimate the maximum speed of an unmeasured rockglacier from its MAAT, or to compare its measured maximum speed to the statistical sample compiled here.

From Fig. 3 and Eqs. (6) and (7) it becomes clear that a certain increase in MAAT (or T) leads to a much larger increase in potential rockglacier speed for warmer rockglaciers than for colder ones. Differentiation of Eqs. (6) and (7) gives:

$$\frac{\partial v}{\partial T} = ab \cdot e^{b \cdot T}, \text{ or } \frac{\partial v}{\partial T} = 0.4c \cdot e^{0.4 \cdot T} [m a^{-1} C^{-1}] \quad (8)$$

where v is the surface speed and T is the ice temperature, or v is the maximum surface speed and T is the MAAT, depending if the set of parameter of Eq. (5), or Eqs. (6) and (7) are used.

4. Specific observations

In this chapter we summarize four types of actual observations of rockglacier behaviour that might indeed be the consequence of a recent increase in surface and ground temperatures. The first sub-section presents a regional-scale compilation of velocity observations, the other three sub-sections describe local-scale observations for individual rockglaciers. All examples are from the Swiss Alps (Fig. 4).

4.1. Recent acceleration of rockglacier creep

In Fig. 5 rockglacier speeds in the Swiss Alps since the 1970s are compiled according to different measurement periods. All data represent average speeds from the central parts of the rockglaciers. By that procedure, slow or even inactive lateral rockglacier sections were excluded (cf. Section 3 on rockglacier speed versus MAAT). For each rockglacier, exactly the same sample of points was taken for each period in order to avoid biases due to sample differences. However, since the delimitation of the “central parts” was defined manually, the absolute numbers for the average speeds are somewhat arbitrary and will change with a different definition of the area where rockglacier speed is tracked over time. Thus, the proportional change in speed should be interpreted rather than the absolute speed of the rockglaciers investigated. Furthermore, the time periods compared to each other are in parts very different. It should, therefore, be kept in mind that potential speed variations within a (longer) measurement period are not captured by the comparison. Caution is also needed when short periods are analysed, because the

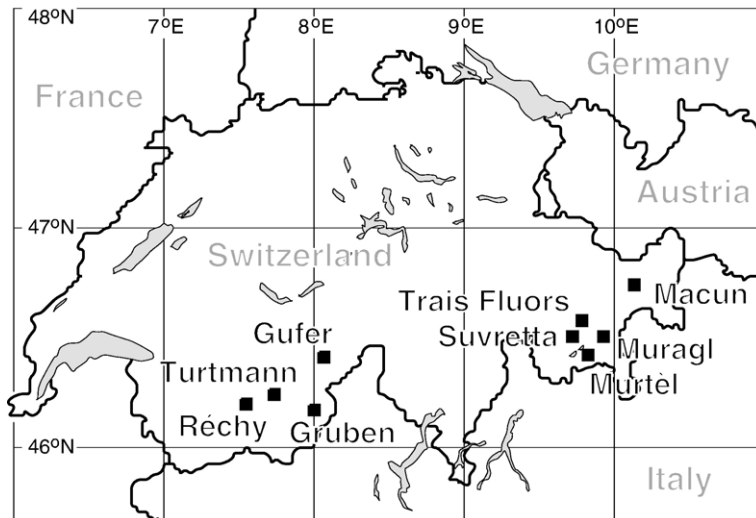


Fig. 4. Location of rockglacier sites in the Swiss Alps, which are referred to in this contribution.

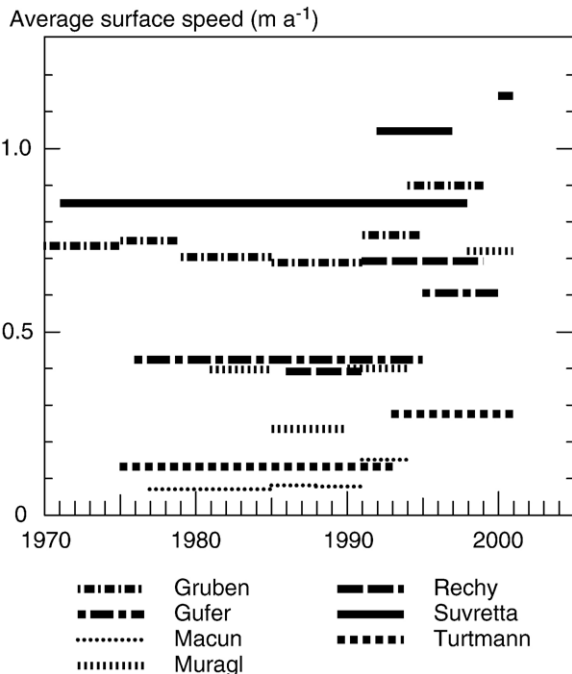


Fig. 5. Average rockglacier surface speeds in the Swiss Alps as measured for different periods since 1970. Most of the data are from photogrammetry, some from geodesy. The Macun data are from Zick (1996), all other data are from the authors of this contribution.

measurements then just depict a “snapshot” of speed, which is not necessarily representative for the long-term behaviour of the individual rockglacier.

The data for Gufer rockglacier (Eggishorn, Aletsch region, Valais) are derived from photogrammetric analyses based on airphotos of 1976, 1995 and 2000 (Kääb, 2005). An increase in average speed of about 40% was observed between 1976–1995 and 1995–2000. (For the photogrammetric technique applied for all rockglaciers see (Kääb and Vollmer, 2000).

The data for Gruben rockglacier (Saas valley, Valais) were compiled from airphotos of 1970, 1975, 1979, 1985, 1991, 1994, 1995, and 1999 (Kääb et al., 1997; Strozzini et al., 2004). Thereby, only the fast-creeping lower part of the rockglacier was analysed, not the entire rockglacier (cf. Kääb et al., 1997). Between 1970 and 1995 speed variations in a range of 10% can be found, with a speed increase between 1985–1995 and 1994–1999. An increase of about 20% or more seems to have occurred after 1994 compared to the 1970–1995 average. (Measurements were done 1985–1995 and 1994–1999).

Airphotos from 1981, 1985, 1990, and 1994 were compiled for the central part of Muragl rockglacier (Upper Engadine, Grisons) revealing a drop in average speed from 1981–1985 to 1985–1990 by over 30%, and

then again a rise for 1990–1995 by 70% to approximately the same speed as 1981–1985 (Kääb, 1998). The data are also compared to terrestrial measurements at the same location from 1998–2001 revealing another rise of about 80% (see following section on seasonal velocity variations). However, the fact that Muragl rockglacier shows drastic changes in speed within very short time intervals calls for special care when interpreting the speed variations of this rockglacier. A small shift in the measurement period, e.g. by some weeks, is potentially able to significantly alter the speed variations obtained photogrammetrically (and geodetically).

For Becs-de-Bosson rockglacier (Val Réchy, Valais) airphotos of 1986, 1991, and 1999 were analysed. An increase in speed of about 70% was found between 1986–1991 and 1991–1999 (Kääb, 2005). (See also last paragraph of this section).

Airphotos of 1971, 1992, 1997, and 1998 were analysed for Suvretta rockglacier (Piz Julier, Upper Engadine, Grisons) (Kääb, 2000; Frauenfelder et al., 2005; Kääb, 2005). For 2000–2001 the speed of selected surface blocks was surveyed terrestrially (Kääb, 2005; Kääb and Reichmuth, 2005). The results of the latter one-year measurements might be affected by seasonal or pluriannual speed variations. The photogrammetric measurements reveal an increase in speed by about 20–30% between the periods 1971–1992 and 1992–1997. The speeds on Suvretta rockglacier are the highest considered in this section. (Note that overlapping periods 1971–1998 and 1992–1997 were compiled rather than subsequent ones due to different airphoto scales for the two periods).

For the orographic right side of the Turtmann valley (Valais) the speed of 14 rockglaciers was analysed from airphotos and digital image data of 1975, 1993, and 2001 (Roer, 2005; Roer et al., 2005b,c). Here, only the average speeds of all 14 rockglaciers are given. All rockglaciers showed acceleration in the second measurement period, with speed increases for individual rockglaciers ranging from 20–350%, on average 125%. (See also following section on rockglacier instability).

The measurements for Macun rockglacier (Lower Engadine, Grisons) are taken from Zick (1996). The values indicated in Fig. 5 represent the average speeds for six rockglacier zones, where a total of about 100 surface points were surveyed repeatedly. The increase in speed for 1991–1994 compared to the periods 1977–1985–1988–1991 amounts to about 50%. It should be noted, however, that for 1965–1967 (not depicted) some of the zones showed already speeds nearly as high as 1991–1994 (Barsch and Zick, 1991; Zick, 1996). Macun rockglacier speeds are quite low compared to the other rockglaciers considered in this section.

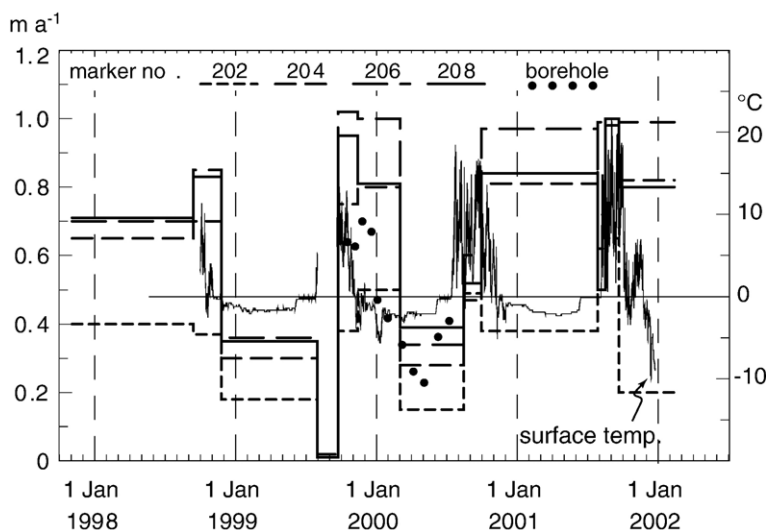


Fig. 6. Surface speed variations during 1998–2001 on Muragl rockglacier as measured from repeated terrestrial surveying of markers. The ground temperature (thin solid line) was obtained from three miniature data loggers at a depth of several decimetres. The bold dots indicate borehole deformation measurements by [Arenson et al. \(2002\)](#) close to marker 204. Marker no. 202 is the lowest one observed and is situated at the front of the rockglacier lobe ([Kääb, 2005](#)). This might explain the different magnitude of speed and speed variations compared to the other markers depicted.

In addition to the studies included here, a number of other authors have also observed recent increases in rockglacier speed in the European Alps. For instance, [Lambiel and Delaloye \(2004\)](#) and [Lambiel \(2005\)](#) found an increase of about 8–10% or 38–50%, respectively, between 2000–2001 and 2001–2003 for two rockglaciers in the Yettes Condjâ valley (Valais, Swiss Alps), and of about 80% or 22%, respectively, between 2001–2003 and 2003–2004. On Bets-de-Bosson rockglacier (Réchy valley, Valais, Swiss Alps; see also above) [Perruchoud and Delaloye \(2005\)](#) reported speeds for 2001–2003 that were 50% higher than the above photogrammetric measurements for 1986–1999. For 2003–2004 they found an additional increase of about 10–15%. For frozen material at Aget–Rogneux these authors mention an increase in speed of about 60% between 2001–2003 and 2003–2004. Presently, an overview of recent rockglacier acceleration in the European Alps is being compiled ([Roer et al., 2005a](#)).

Some further long-term monitoring series of rockglacier speed are available from the Austrian Alps (e.g. [Kaufmann, 1998](#); [Schneider and Schneider, 2001](#); [Kaufmann and Ladstädter, 2002](#)). [Schneider and Schneider \(2001\)](#), for instance, found inter-annual speed variations for Äusseres Hohebenkar rockglacier in the range of about 100%, with significant correlation to air temperature variations. For Reichenkar and Ölgruben rockglaciers a marked increase in surface speed was observed since about 1990. Thereby, Reichenkar rockglacier showed no seasonal speed variations, whereas Ölgruben

rockglacier did in a pronounced way (K. Krainer, personal communication, 22 September 2005).

4.2. Seasonal speed variations

[Fig. 6](#) shows the horizontal speed of selected markers on Muragl rockglacier (Upper Engadine, Swiss Alps) as measured during repeated terrestrial polar surveys using a high-precision total station TCA1102 by Leica Geosystems (motorised, combined theodolite and laser ranger). In total, measurements were performed for 20 markers on two flow lobes ([Kääb et al., 2003](#); [Kääb, 2005](#)). The selection shown here is representative for the entire sample.

Surprisingly, the speed of the frozen debris changes from magnitudes smaller than 0.2 m a^{-1} to up to 1 m a^{-1} within a few months. These “stop-and-go” events happen in a fairly synchronous way over the study area. The average speed over the observed period of 1998 to 2001 is higher than the 1981 to 1994 average speed obtained from photogrammetry ([Kääb and Vollmer, 2000](#)) (see Section 4.1).

In a 72 m deep borehole close to point no. 204, short-term deformation variations were observed by [Arenson et al. \(2002\)](#) for winter 1999/2000 (points in [Fig. 6](#)). Projecting the borehole-derived variations into adjacent years clearly shows that the magnitude, average speed, wavelength and/or onset of the changes vary significantly from year to year. For logistical reasons, the repeated surveying of the surface markers could not be performed in equal time intervals.

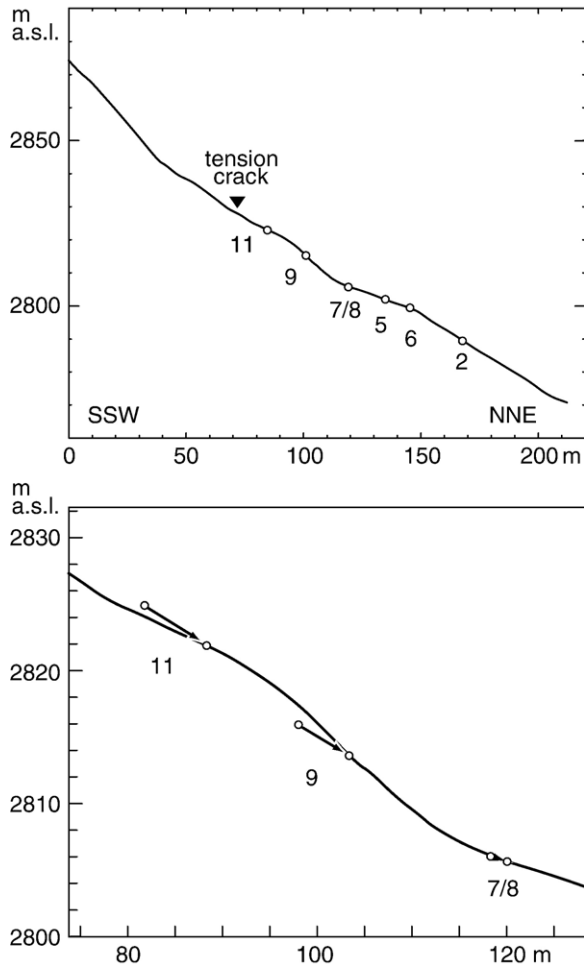


Fig. 7. Topographic down-slope profile along a frozen slope at Trais Fluors (Upper Engadine, Swiss Alps) with selected surface markers for geodetic displacement measurement. The lower panel shows an enlargement for a section of the upper panel, with displacement vectors 1998–2005 indicated in the same scale as the topography. Points no. 7 and 8 are adjacent markers with similar elevation and thus only one position on the profile projection is shown.

Ground surface temperatures at five locations throughout the study area were measured from miniature temperature loggers placed under large blocks. The temperature series obtained do not show marked differences among each other. The onset of rockglacier speed-up is delayed by about 2 mos compared to the surface melt-out of snow, which is indicated by a ground temperature rise above zero (Fig. 6). The delay of rockglacier speed-up after penetration of meltwater into the ground amounts to about 3 to 4 mos. The meltwater penetration is indicated by the rise of the basal temperature of snow (BTS) from permafrost conditions (approximately -3°C) to the so-called stable “zero-curtain” (0°C). In winter 2002, insufficient snow cover and thus lack of insulation led to extraordinary ground cooling.

Observations of seasonal variations in rockglacier speed are rare. Perruchoud and Delaloye (2005) found a speed maximum in late summer for Becs-de-Bosson rockglacier (Réchy valley, Valais, Swiss Alps; see also Section 4.1). For 1971 to 1973, Barsch and Hell (1975) reported slight seasonal velocity variations for Muragl rockglacier (summer speed about 130% of winter speed). For Murtèl rockglacier they found seasonal variations of up to one order of magnitude during the same time period. In contrast to their findings, the borehole studies since 1987 at Murtèl rockglacier and accompanying terrestrial surveys revealed little to no speed variations (Wagner, 1996). A clear seasonal cycle in movement was detected by Haeberli (1985) for Gruben rockglacier. Seasonal speed maxima were found for autumn 1981 and spring 1982.

4.3. Activation of a small rockglacier

At Trais Fluors (Upper Engadine, Grisons; altitude about 2800 m a.s.l.) the movement of about 30 selected blocks on west and north facing slopes covered by solifluction lobes is monitored since 1998 (Ikeda et al., 2003). (For the technique applied see section on seasonal velocity variations on Muragl rockglacier). At the beginning of the surveys, no pronounced topography was noticed on the slopes. However, it soon turned out that the blocks on the northern slope showed displacement rates of up to 1 m a^{-1} and more (Figs. 7, 8), whereas the blocks on the western slope displaced on the order of a few cm a^{-1} as expected from solifluction features in the area (not depicted). Using borehole inclinometry, Ikeda et al. (2003) found that the rapid surface movements are not

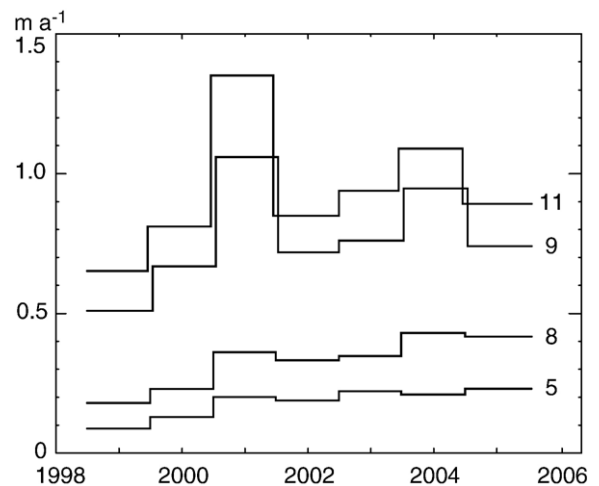


Fig. 8. Inter-annual speed variations for markers from Fig. 7, Trais Fluors. Speed of marker no. 7 is not depicted due to an interruption of the observation series at this point caused by snow remains.

restricted to the surface but that 90% of the slope deformation occurs at depths larger than 5 m. Temperature measurements and DC resistivity surveys on the northern slope suggest a debris-rich frozen layer at temperatures close to the melting point and down to a depth of 10–20 m (Ikeda et al., 2003).

Up to now (2005) the surprisingly high speed of the slope under study continued (Fig. 8). Meanwhile, the slope topography clearly reflects the high deformation and the strong negative down-slope gradient of speed. An initial, protalus-rampart-like ramp has developed around points no. 9 and 11 (Figs. 7, 8). The fact that the slope of the surface particle displacement (i.e. the relation between the vertical and the horizontal vector component; see vectors in Fig. 7) is similar to the overall surface inclination of the slope points to mass advection playing a further important role in the development of the small rockglacier- or protalus-rampart-like feature.

Over longer time periods, dynamic conditions as they are currently found on the slope under study would have led to a pronounced rockglacier topography through mass conservation (Frauenfelder, 2005). However, such a landform is not visible. On the other hand, the mass currently transported down-slope due to the high speed is

very unlikely to be in equilibrium with the material supplied from above by rockwall retreat (Ikeda et al., 2003). Both considerations together let us conclude that the deformation feature must be transient. Though hard to prove in detail, ground temperatures might have exceeded a certain threshold (e.g. allowing melt water to percolate) at a time before the measurements were started. This might then have caused the deformation to start, or to accelerate by some orders of magnitude. In fact, the inter-annual monitoring series of block displacements (Fig. 8) shows that in particular points no. 9 and 11 (and other points not depicted here) may be highly sensitive to variations in boundary conditions. The peak speed between summer 2000 and summer 2001 has, for instance, been attributed to snow conditions, which prevented ground cooling during winter and thus led to increased ground temperatures (Ikeda et al., 2003). The 2004 speed peak has not yet been analysed using the snow and ground temperature measurements in the area.

Mean annual speed of the surveyed blocks seem to have increased by roughly 50–100% during the monitoring period. However, for the lower points this could also be a pushing effect from the protalus feature further up.

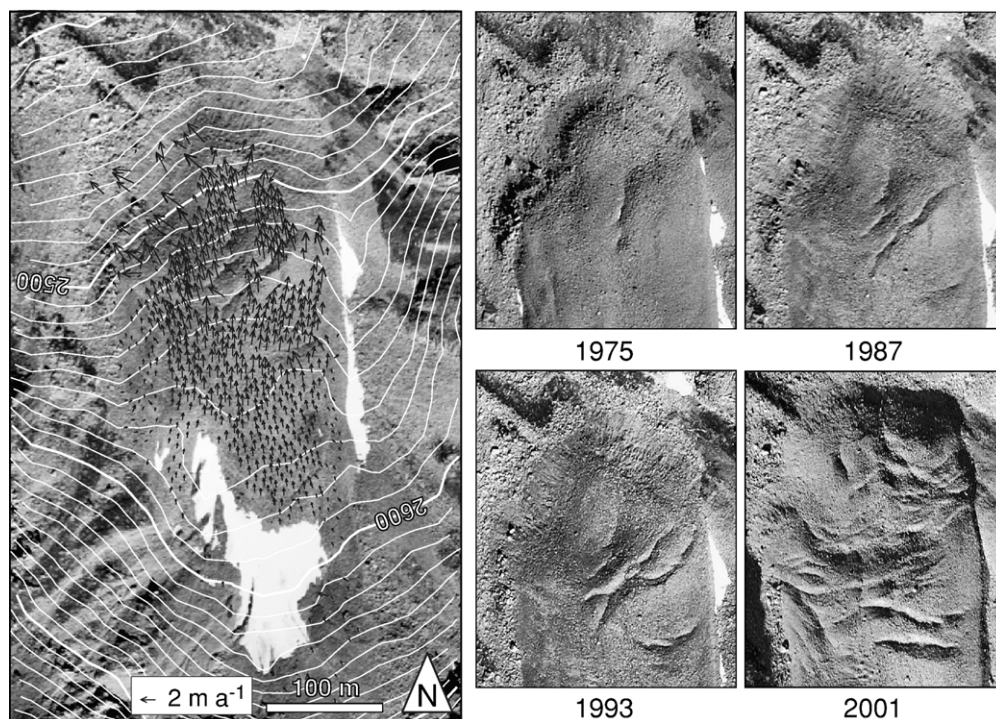


Fig. 9. Instability of a rockglacier snout (Turtmanntal, Swiss Alps). Left panel: velocity vectors measured from airphotos of 1987 and 1993; right panels: orthophotos of the terminus section of the rockglacier. Images of 1975, 1987, and 1993 are courtesy of swisstopo; the 2001 image is a linear array CCD (pushbroom) image taken by the HRSC-A camera, courtesy of the Department of Geography, University of Bonn. The rockglacier instability led to enhanced rockfall and required construction of a protection dam (not depicted). (See also Roer et al., 2005b).

Worth to be mentioned in this context are the findings by [Lugon et al. \(2005\)](#) who report about the possible reactivation of an inactive rockglacier at Arolla (Valais, Swiss Alps), a process which might, to some extent, be comparable to that found at Trais Fluors.

4.4. Instability of rockglacier sections

For two rockglaciers in the European Alps instabilities of large terminal sections were observed recently. For a rockglacier in the Grüob valley, lateral to the Turtmann valley (Valais, Swiss Alps), first transverse crevasses could be seen on airphotos from 1987 ([Fig. 9](#)). Airphotos from 1993 show enhanced crevasses, and in 2001 a total collapse of the terminal section of the rockglacier was observed. Between the periods 1975–1981, 1981–1987, and 1987–1993 photogrammetric measurements revealed a steady increase of surface speeds ([Roer, 2005; Roer et al., 2005b,c](#)). Between 1993 and 2001 photogrammetric measurements were impossible for most sections due to the surface destruction accompanying the rockglacier collapse. In particular during the observational period 1993–2001 the rockglacier advanced significantly ([Fig. 9](#)).

A behaviour similar to the Grüob valley rockglacier was also reported since the 1970s for the Äusseres Hochebenkar rockglacier in the Ötztal (Tirol, Austrian Alps) ([Kaufmann and Ladstädter, 2002](#)). In both cases, Grüob and Äusseres Hochebenkar, the destabilized zones are located in steep rockglacier sections just below a significant longitudinal increase in surface slope. For the rockglacier in the Grüob valley, photogrammetric stereo-interpretation suggests that the crevasses developed are several metres deep, presumably exceeding the active layer thickness. In addition, the longitudinal extension of the crevasse pattern over time points to stress transfer, which can only be due to ground ice. Both findings suggest that the destabilization is not restricted to the active layer alone. Due to the lack of in-situ analyses for the rockglacier in the Grüob valley, no more detailed process descriptions can be given.

5. Discussion

5.1. Model

The numerical model for simulating surface speed variations, which are driven by surface temperature variations, reveals for a pure ice body:

- (1) Seasonal surface temperature variations are expected to cause speed variations up to about 15%, significantly depending on the average ice temperature. The latter effect is a consequence of the

assumed exponential increase of the rate factor with temperature.

- (2) Speed variations in the order of 10–50% or more may be expected for pluriannual temperature changes. The higher sensitivity compared to seasonal temperature changes is due to the enhanced heat diffusion with lower frequency of temperature forcing, i.e. lower thermal attenuation.
- (3) The warmer the ice temperatures are, the larger are the variations of surface speed due to a certain change in surface temperature.
- (4) A soft layer within the ice body does not necessarily enhance the speed variability. Depending on its vertical position, it is able either to amplify or to compensate the effect of changes in rate factor with temperature. This is due to the fact that the temperature of a soft layer – and with it the deformation – might reach its maximum at a time when the other layers are comparably warm and fast, or comparably cold and slow – depending on the depth of the soft layer.
- (5) The less deep a soft layer is located, the higher is the magnitude of surface temperature variations that reach it. Since the rate factor is assumed to change with temperature it follows, in principle, that the less deep a soft layer is located the higher are the variations in surface speed of the creeping ice body. Note that for shorter frequencies of temperature variations the compensation effect discussed in the paragraph above overlays this effect.
- (6) For steeper surface slopes (we tested 12° against 20°) the percentage speed variability due to surface temperature variations remains similar. However, because the overall speed (and with it the volume advected in the vertical column) is significantly higher for steeper conditions compared to less inclined situations, the same percentage variation means a significant difference in spatial mass flux variation for different slopes. As a consequence, surface temperature variations seem to be able – according to our model – to cause significant spatio-temporal variations in rockglacier mass flux at changes in slope. In other words, temperature variations might cause spatio-temporal variations in horizontal speed gradients. The development of rockglacier surface micro-topography, such as transverse ridges, might be connected to such processes (cf. [Kääb and Weber, 2004](#)).

A number of parameters in the model may be (and were) altered. Such a procedure changes the numbers

given above. The basic conclusions given here, however, do not change substantially.

5.2. Comparison between field observations and model

As already shown in Section 3, the increase of the rate factor with ground temperature as assumed in the model is well confirmed by analysing the relation between MAAT and surface speed for a large number of rockglaciers. This suggests that differences in ground temperature indeed are a fundamental parameter of differences in rockglacier dynamics. The closer the ground temperature of a rockglacier approaches 0 °C the larger is the increase in speed for a given increase in surface or ground temperature (Eq. (8)). Rockglaciers in warmer climatic conditions or closer to the lower limit of local permafrost distribution (i.e. rockglaciers with temperatures close to 0 °C) are thus expected to react more sensitively to ground temperature changes from air-temperature or precipitation changes (in particular snow) than colder ones.

There is no doubt that the inclusion of processes other than one-dimensional heat diffusion and viscous temperature-dependent deformation within the model would have a significant effect. The interactions between debris and ice in the material, and especially the action of water, can be of major importance (Arenson, 2002; Ladanyi, 2003). In fact, the model results cannot explain the large variability range of 100% and more from the average speed found for Muragl rockglacier (Section 4.2). From borehole studies, this rockglacier is known to have a low ice-content with a temperature slightly below or close to 0 °C and a water table was encountered below the permafrost base at a depth of 30–36 m (Arenson et al., 2002; Vonder Mühll et al., 2003). For other rockglaciers, observed speed variations are in the range modelled here (Barsch and Hell, 1975; Haeberli, 1985; Kääb et al., 1997; Arenson et al., 2002).

In most of the investigated cases the overall increase in rockglacier speed observed for the last decades in the Swiss Alps (Section 4.1) is in the same order as the speed increase modelled for a surface temperature increase of about 1 °C. Though a number of processes at the atmosphere–ground interface and within the ground act apart from the pure heat diffusion considered here (e.g. Delaloye, 2004), this agreement suggests that surface and ground temperature warming might indeed be responsible to a large extent for the recently observed rockglacier acceleration. However, some cases with significantly larger accelerations than modelled suggest that other processes may be important as well. As mentioned before, the presence of water in rockglaciers close to the melting point might thereby play a prominent role.

Similarly, the process of rockglacier activation found at Trais Fluors (Section 4.3) cannot be explained by a function of ice deformation versus temperature of the form applied here (e.g. Fig. 3). A much steeper gradient in the relation is required, for instance through a larger exponent b in Eq. (6). In fact, the curve fits without fixed b reveal a larger exponent b (0.6–0.8) compared to the one derived for glaciers (0.4) and used in our model (Fig. 3; Eqs. (6) and (7)). Laboratory tests seem also to suggest a more pronounced increase of the rate factor above around –2 °C (Section 5.1.1 in Arenson, 2002). For special cases such as Trais Fluors even significantly larger exponents b might apply and lead to a behaviour, which resembles a strong increase in deformation above a certain ground temperature threshold. The possibility for such an increase could well be connected to certain geotechnical ground characteristics (Arenson, 2002). These results suggest to run the presented model also with different exponents for the functions $v=f(\text{MAAT})$ and $A=f(T)$, or different temperature-dependent creep laws (Arenson, 2002; Ladanyi, 2003).

The instabilities reported in Section 4.4 represent a further continuation of the sequence of observations discussed here in a sense that heat diffusion alone is by far not able to explain these phenomena. Rather, impact of meltwater, thermokarst processes and/or significant melt-out of the ice content (i.e. processes related to the ice–water phase transition) could play the crucial role, possibly in connection with a number of other processes such as spatio-temporal changes in shear stress (e.g. creep over a terrain step), active layer detachments, erosion of insulating blocks, etc.

6. Conclusions

The combined findings from modelling and field investigations presented here suggest that the creep of perennally frozen material close to 0 °C is more sensitive to climate forcing than the creep of colder material. Furthermore, a rise in rockglacier temperature leads to an exponential increase in average speed, accompanied by a marked increase in seasonal to pluriannual variability of velocities, i.e. a marked increase of sensitivity to temperature variations. Seasonal speed variations of rockglaciers of up to about 15% and decadal-scale variations of up to 100% in response to a 1 °C surface temperature increase can be explained by heat diffusion into the ground.

At a later stage of increasing rockglacier temperature, processes other than temperature-dependent deformation seem to play an important role (water, boundary effects between debris and ice, etc.), before a significant

loss of ice content by melt-out will be able to reduce the deformation rate of the frozen mass again towards its deactivation.

The model study stresses the importance of learning more about shear zones in rockglaciers. Such layers appear to be an essential part not only of rockglacier dynamics, but also of their response to temperature changes.

For applied geomorphological purposes such as landscape evolution studies or hazard management, the comparison between our model and the presented field observations let us expect the following, non-exclusive, set of potentially short- or medium-term reactions of rockglacier creep under conditions of present atmospheric warming:

- (1) Many rockglaciers might (or do already) show an increase in surface speed. The warmer the ground thermal conditions are, the higher is the expected percentage acceleration.
- (2) Such speed-up might in cases lead to an imbalance between mass transport due to creep and debris supply onto the rockglacier from its headwall. A long-term consequence might be the “dynamic inactivation” of such a rockglacier (Olyphant, 1987; Barsch, 1996).
- (3) The number of rocks that are transported over the rockglacier front and the total horizontal mass flux increase with increasing (surface) speed. As a consequence, local rockfall hazards and the build-up of potential debris-flow starting volumes might enhance in certain topographic conditions (e.g. Hoelzle et al., 1998).
- (4) With enlarging amplitude and increasing variability in timing for seasonal and inter-annual speed variations, special measures in the planning and analysis of rockglacier-velocity monitoring series might be needed.
- (5) Under certain topographic and geotechnical conditions, (transient) creep processes might be (re-) activated in frozen slopes, which have been (almost) stable so far.
- (6) Rockglaciers might show landslide-like instabilities, in particular at their snouts. Whereas an increase in ground temperature alone might only in rare cases be responsible for such destabilization, it might well trigger processes, which then lead to such ruptures.

As a consequence of the above potential (and in some cases indeed actual) reactions of rockglaciers to climatic change in cold mountains, strengthened rockglacier monitoring seems advisable, in particular in densely populated mountain regions such as the European Alps.

Acknowledgements

The geodetic measurements at Trais Fluors were conducted together with Norikazu Matsuoka and Atsushi Ikeda (University of Tsukuba, Japan), and a number of students of the Department of Geography, University of Zurich. The photogrammetric compilations for Réchy rockglacier were made in collaboration with Jean-Philippe Dousse and Reynald Delaloye, Department of Geosciences, University of Fribourg. Most of the airphotos used in this contribution were taken by the swisstopo flight-service, recently within the Swiss permafrost monitoring network PERMOS. Special thanks are due to Karl Krainer, Emmanuel Reynard, and Christoph Schneider for their thorough and helpful comments.

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