Permafrost Creep and Rock Glacier Dynamics

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ABSTRACT

This review paper examines thermal conditions (active layer and permafrost), internal composition (rock and ice components), kinematics and rheology of creeping perennially frozen slopes in cold mountain areas. The aim is to assemble current information about creep in permafrost and rock glaciers from diverse published sources into a single paper that will be useful in studies of the flow and deformation of subsurface ice and their surface manifestations not only on Earth, but also on Mars. Emphasis is placed on quantitative information from drilling, borehole measurements, geophysical soundings, photogrammetry, laboratory experiments, etc. It is evident that quantitative holistic treatment of permafrost creep and rock glaciers requires consideration of: (a) rock weathering, snow avalanches and rockfall, with grain-size sorting on scree slopes; (b) freezing processes and ice formation in scree at sub-zero temperatures containing abundant fine material as well as coarse-grained blocks; (c) coupled thermohydro-mechanical aspects of creep and failure processes in frozen rock debris; (d) kinematics of non-isotropic, heterogeneous and layered, ice-rich permafrost on slopes with long transport paths for coarse surface material from the headwall to the front and, in some cases, subsequent re-incorporation into an advancing rock glacier causing corresponding age inversion at...
depth; and (e) the dynamics of rock glaciers, which include spatial and temporal variations in velocity that are manifested in the ridges, furrows and other surface structures typical of rock glaciers, as well as their down-valley motion. Copyright © 2006 John Wiley & Sons, Ltd.

KEY WORDS: permafrost; rock glaciers; cold mountains; periglacial slope movement; geomorphic landforms; Martian ice flow features

INTRODUCTION

The growing interest in mountain permafrost has led to rapid progress in recent years in understanding both the gravitationally-induced creep of permafrost on slopes in cold mountain areas and the motion of rock glaciers (Figure 1). The planetary science community has also become increasingly interested in diverse features visible on Mars that resemble rock glaciers or are suggestive of deformation of subsurface ice (Head et al., 2005). The widespread occurrence of ice just below the surface is consistent with measurements from the gamma-ray and neutron spectrometers on board Mars Odyssey (Feldman et al., 2004) and theoretical expectations (Mellon and Jakosky, 1995). In addition, conservative estimates suggest that Mars may possess a planetary inventory of water equivalent to a global ocean 0.5–1 km deep of which as much as 90–95% is believed to reside as ground ice and groundwater within the planet’s crust (Clifford, 2003). Hence considerable attention is focused on Martian permafrost, especially because the continuing interest in extraterrestrial life is concentrated on the surface where the common ingredients to sustain life—photons and water—occur together.

In this paper, available knowledge is reviewed that could eventually: (1) help advance our understanding of rock glaciers and related features on Earth and Mars; and (2) form the basis for numerically modelling the thermo-mechanically coupled dynamics of creeping ice/rock mixtures. It summarises recent sophisticated field and laboratory experiments (drilling, borehole observations, geophysics, photogrammetry and remote sensing, creep tests, etc.) that have improved our understanding of the principal processes underlying the motion of deep-seated, large-scale creep of perennially frozen debris in alpine settings. The report also refers to the large set of well-established methods, from simple to sophisticated, that are currently available for probing, mapping and monitoring permafrost, and for modelling its spatial distribution in complex high-mountain topography (e.g. Etzelmüller et al., 2001; Harris et al., 2001; Hoelzle et al., 2001; Vonder Mühll et al., 2002).

The term ‘rock glacier’ has been used for various complex landforms of cold mountain areas. Two extremes represent the ends of a geomorphic continuum: (a) steadily creeping perennially frozen and ice-rich debris on non-glacierised mountain slopes; and (b) debris-covered glaciers in permafrost-free areas. Semantically, they can be differentiated as follows: (a) is a ‘rock glacier’ whereas (b) is commonly called a ‘debris-covered glacier’. Terminological
problems arise in areas with continental climates where surface ice interacts with periglacial and subglacial permafrost (Haeberli, 1983). There, complex mixtures of buried surface ice and ice formed in the ground can occur and all transitions from debris-covered polythermal or cold glaciers to ice-cored moraines and deep-seated creep of perennially frozen sediments are possible (cf. Haeberli, 2000; Kneisel et al., 2000). Narrow definitions of landform terms are confusing in such cases and should be replaced by a flexible concept considering, for each individual case, the materials involved, thermal conditions, processes, time periods and spatial scales. The following sections concentrate predominantly to exclusively on rock glaciers on non-glacierised mountain slopes.

A clear distinction must be made between two fundamental but complementary characteristics (Haeberli, 2000):

- ground thermal conditions (permafrost) that allow for the formation and the long-term existence of subsurface ice; and
- composition of the ice/rock mixtures, that defines the amount and distribution of ice existing below the surface and enabling the steady-state creep of the ice/rock mass considered.

Permafrost is used here to describe a specific ground thermal condition (temperature at or below 0°C for a minimum of one—or even better two—years), irrespective of ice content or lithology. Such thermal conditions not only enable the formation and preservation of a variety of ground ice but also determine the spatio-temporal scales of permafrost creep within ice-containing materials.

THERMAL CONDITIONS

Surface and Active Layer

The seasonally frozen active layer above the perennially frozen material is commonly a few decimetres to a few metres thick (Humlum, 1997). Energy fluxes at the surface and thermal characteristics of the active layer control the initiation, growth and maintenance of rock glaciers. The coupling between atmospheric and ground thermal processes due to the non-conductive heat transfer processes is highly complex. Increased knowledge of this coupling and of rock glacier active-layer thermal characteristics is, therefore, essential for understanding the age, climatic sensitivity and palaeoclimatic significance of rock glaciers (Humlum, 1996).

Surface Conditions and Energy Fluxes.

In mountain areas outside the High Arctic, many rock glacier fronts closely approach limits of local permafrost occurrence. As a consequence, mean annual surface temperatures are often not far from melting conditions. Rock glacier surfaces are often characterised by a system of ridges and depressions, orientated parallel or transverse to the overall flow direction (Figure 2). Numerical modelling of energy fluxes through such surfaces requires parameterisation of solar radiation, sensible heat, surface albedo, heat conduction and advection, latent heat transfer, etc., for complex topography and extremely heterogeneous surface characteristics (air, rock); it needs a correspondingly large amount of precisely measured or computed data. Models of energy-balance and thermal-offset (accounting for the difference between the temperature at the surface and at the permafrost table, respectively (Mittaz et al., 2000; Hoelzle et al., 2001)) must be combined. Modelling the evolution of the snow cover (Mittaz et al., 2002) is one of the principal components because snow cover not only insulates the ground from the cold air during winter and from warm air in summer but also critically influences the surface albedo and, hence, net radiation. Initial results of detailed modelling (Stocker-Mittaz et al., 2002) and measurements in coarse debris mantling a glacier (e.g. Conway and Rasmussen, 2000) clearly show that advective subsurface fluxes are important. Special conditions are encountered at the head of many rock glaciers under near-perennial to perennial patches of surface ice—often deposited by...
spring avalanches. Such perennial snowbanks or glacierets are generally thin (metres), cold and frozen to the bed, because ice temperature cannot rise above zero during the summer while temperatures drop far below zero during the winter (Haeberli, 2000). As a consequence, they are conducive to generating negative ground temperatures and participate in the build-up of frozen material, which tends to creep downslope.

Active Layer Characteristics and Processes.

The thickness of the active layer on rock glaciers is a function of local meteorological characteristics, thickness and duration of the winter snow cover, and the active layer’s textural characteristics. The grain size usually decreases with depth in the active layer. With increasing depth, fine-grained material gradually becomes abundant, and close to the permafrost table, fines are often predominant (Figure 3). On pebbly rock glaciers (Matsuoka and Ikeda, 2001), fines may be significant through most of the active layer. Especially large boulders may penetrate the general terrain surface to form ‘natural cairns’ (Humlum, 1988). Active-layer thicknesses from about 0.5 m to 3 m are typically reported from excavations, drill-holes or geophysical soundings (e.g. Barsch, 1996; Vonder Mühll, 1996; Wagner, 1996; Humlum, 1997; Isaksen et al., 2000). Depending upon deformation of the rock glacier, the active layer may extend across zones of longitudinal extension and compression, giving rise to local thermo-mechanical interactions (Haeberli and Vonder Mühll, 1996), introducing at a local scale either degradation or aggradation of the permafrost below due to mechanical thickening (compression) and thinning (extension) of the active layer.

Automatic dataloggers can readily be used to monitor ground temperatures, snow cover duration, zero-curtain effects, etc. During winter periods with low wind speed and thin snow cover, radiative heat losses may produce very cold air, which tends to penetrate into the active layer. For example, rock glacier active layer temperatures in Greenland in early October follow the surface temperature with little lag (Figure 4). A thin snow cover during autumn and winter contributes to rapid cooling of the active layer, partly due to surface radiation effects (Keller and Gubler, 1993), but primarily by not blocking surface openings in the active layer surface (Bernhard et al., 1998).

Cold air penetrating somewhere upslope slowly flows downslope within the active layer, to resurface at lower elevations. In contrast, when a thick snow cover exists on rock glaciers, active layer temperatures are mainly controlled by conduction (Harris and Pedersen, 1998; Humlum, 1997; Hoelzle et al., 1999). In the latter situation, temperature variations are smooth and follow the surface temperature with considerable lag. In Figure 4, this characterises the period from late November to late April. Non-conductive processes may, however, be important if vertical holes (funnels) exist in the winter snow, connecting the outside air with the air beneath the snow cover. As described by Keller and Gubler (1993) and Bernhard et al. (1998), such funnels are mainly found along surface depressions in the rock glacier surface and provide easy access for cold air to reach the interior of the active layer, displacing warmer air upwards.

The presence of snow cover in late winter and early spring introduces non-conductive heat transfer processes during periods of snowmelt, as surface meltwater percolates into the active layer and refreezes (Humlum, 1997). Similarly, if warm air masses deliver rain on cold snow in the winter or spring, the refreezing of percolating water can warm the snow and the upper part of the active layer appreciably (Putkonen and Roe, 2002).

Such an event is illustrated in Figure 4 at the end of May 2001, when temperatures rise rapidly at all levels in the active layer. Radiation effects during summer produce very warm air masses at the surface. However, due to density differences, such warm air rises and does not penetrate into the active layer. Corresponding heat transfer is conductive, while cooling during clear nights may take place much more rapidly due to density-induced non-conductive effects (Humlum, 1997). This is essentially what has been called Balch ventilation, defined as the insulating effect of air-filled
voids in coarse deposits (Thompson, 1962; Barsch, 1996). Wind pumping generated by high wind speeds may, however, upset the above general pattern and drive warm surface air masses into the active layer by way of forced ventilation, dynamically displacing colder and denser air masses (Humlum, 1997), as is well known from similar processes in snow (Clarke et al., 1987; Colbeck, 1989).

Permafrost Conditions

Boreholes.

Relatively few boreholes have been drilled into and through rock glaciers to permit investigation of the thermal characteristics of the deeper subsurface. In the 1970s, two boreholes were completed in the Swiss Alps (10 m at Murtèl: Barsch, 1977a; 7 m at Gruben:

Figure 4 Surface and active layer temperatures (°C) and photographs illustrating conditions from October 2000 to June 2001 at Mellemfjord, Disko Island, central West Greenland. Upper left photograph: 4 October 2000; upper right: 14 May 2001; lower left: 24 May 2001; lower right: 26 June 2001. The photographs were obtained by an automatic digital camera looking SW towards two cirques with rock glaciers. The mountains on the opposite side of the fjord reach about 1000 m a.s.l.
Barsch et al., 1979) and one in Canada (Kluane Range: Johnson and Nickling, 1979). In the latter, temperatures remeasured some six years after drilling were no longer negative indicating a very rapid thaw of the permafrost (probably assisted by water advection). In 1987, a 58 m deep borehole drilled through the Murtel-Corvatsch rock glacier initiated one of the longest measurement time series of Alpine permafrost temperatures (Vonder Mühll and Haeberli, 1990; Haeberli et al., 1998; Vonder Mühll et al., 1998). At Pontresina-Schaflberg, two boreholes were drilled in 1990 (37 m and 65 m deep) as part of a project concerning snow avalanches and debris flows (Vonder Mühll and Holub, 1992). A borehole was cored to 10 m depth in 1995 at Galena Creek, Wyoming, USA (Ackert, 1998). Within a joint project of three institutes of the Swiss Federal Institute of Technology (ETH) Zurich, several boreholes were cored in the Muragl rock glacier (four boreholes, ca. 70 m deep, 1999) and in the Murtel-Corvatsch rock glacier (two boreholes, 51 m and 63 m, 2000: Arenson and Springman, 2000; Arenson, 2002; Vonder Mühll et al., 2003). The main results obtained concerning thermal conditions are summarised in Table 1.

The rock glaciers investigated are all located near the local boundary of permafrost distribution and correspondingly contain warm permafrost typically some tens of metres thick. Kudrayahov et al. (1991) discuss technical aspects of drilling in permafrost. Details about drilling in mountain permafrost and parameters to be investigated are given in Vonder Mühll (1996) and Arenson (2002).

**Data Analysis and Results.**

Heat transfer in permafrost largely occurs through conduction (Lachenbruch et al., 1988), hence permafrost temperatures can be analysed using classical heat conduction theory (Carslaw and Jaeger, 1959). Borehole temperatures have been measured in rock glaciers with an absolute and relative accuracy of \(< \pm 0.1^\circ\text{C}\) and \(< \pm 0.05^\circ\text{C}\), respectively (Vonder Mühll, 1992), and borehole measurements in other permafrost regions and ice sheets indicate that modern instrumentation could provide considerably more precise definition of the thermal regime of rock glaciers (e.g. Cuffey and Clow, 1997; Dahl-Jensen et al., 1998).

The most systematic analysis of temperatures observed in a rock glacier borehole was made for the 1987 borehole at Murtel 2 (Vonder Mühll and Haeberli, 1990; Vonder Mühll, 1992; Vonder Mühll et al., 1998). Murtel rock glacier is located at the lower limit of permafrost distribution and its tongue is advancing onto permafrost-free terrain. Heat flow was determined as about 150 mW m\(^{-2}\), with thermal conductivity of the frozen material being determined (a) experimentally on core samples in the laboratory and (b) using observed amplitude reductions and phase lags at depths. The exceptionally high temperature gradient and heat flow are caused by the existence of a talik at a depth of about 50 m (Vonder Mühll, 1992), probably maintained by groundwater flow in summer. At the same site, the time series of borehole data collected since 1987 (Figure 5) clearly documents some thermal characteristics of permafrost, in particular:

- below the active layer, heat is transported almost exclusively by conduction resulting in an exponential decrease in amplitude and a linear increase in phase lag with depth;
- the subsurface acts as a low-pass filter in terms of the temperature signal introduced at the surface, resulting in a low-amplitude sinusoidal annual signal at about 10 m depth and a temperature that is close to the annual average;
- snow strongly influences permafrost temperatures, and
- since the beginning of the measurements, temperatures ranged from −2.6°C to −1.2°C at ≈10 m

### Table 1  Key temperature data from boreholes through rock glaciers.

<table>
<thead>
<tr>
<th>Site</th>
<th>Number of boreholes</th>
<th>Maximum drilling depth (m)</th>
<th>Mean temperature at the permafrost table (°C)</th>
<th>Permafrost thickness (estimated) (m)</th>
<th>Depth of shear zone (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Murtel-Corvatsch (Switz.)</td>
<td>5</td>
<td>63</td>
<td>−2.0</td>
<td>&gt;50, talik</td>
<td>22–30</td>
</tr>
<tr>
<td>Gruben</td>
<td>1</td>
<td>7</td>
<td>−1.0</td>
<td>80</td>
<td></td>
</tr>
<tr>
<td>RG II, Kluane Range</td>
<td>1</td>
<td>17</td>
<td>−0.6</td>
<td>&gt;20</td>
<td></td>
</tr>
<tr>
<td>Pontresina-Schafberg 1/1990</td>
<td>1</td>
<td>65</td>
<td>−1.5</td>
<td>&gt;60</td>
<td>16</td>
</tr>
<tr>
<td>Pontresina-Schafberg 2/1990</td>
<td>2</td>
<td>37</td>
<td>−0.5</td>
<td>30</td>
<td>ca. 30</td>
</tr>
<tr>
<td>Galena Creek</td>
<td>1</td>
<td>10</td>
<td>−1?</td>
<td>?</td>
<td></td>
</tr>
<tr>
<td>Muragl</td>
<td>4</td>
<td>72</td>
<td>&gt;−0.5</td>
<td>20</td>
<td>17–20</td>
</tr>
</tbody>
</table>

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DOI: 10.1002/ppp
depth, and from $-2.1^\circ$C to $-1.4^\circ$C at $\sim$20 m with a warming trend of a few tenths of degrees celsius per decade.

The temperature at the base of the active layer is the upper boundary condition for the permafrost temperatures at depth. From borehole temperature data above the depth of zero annual amplitude, the date of the first major snowfall, the duration of positive and negative temperatures and the occurrence of zero-curtain effects—especially during snowmelt—appear to be the most important factors influencing overall ground temperatures. Zones of circulating water or air can exist within or below the permafrost body of a rock glacier. Besides the talik already mentioned at Murtèl rock glacier, water was observed in several boreholes in connection with pressuremeter tests in two boreholes (1/2000 (Arenson, 2002) and 2/2000), and a sudden warming took place between 3 and 20 m depth on 25 April 2001 (Vonder Mühll et al., 2003) without affecting the top 3 m of the borehole. In the latter case, warming must have happened from the bottom upwards, and subsequent cooling took place from the top downwards. Indications of temperature variations being closely related to atmospheric temperatures and layers of frozen ground below the main portion of the permafrost were observed in all three boreholes from the Muragl rock glacier. Such short-term changes at depth are interpreted to arise from heat advection due to air, rather than water, movements (Vonder Mühll et al., 2003).

**COMPOSITION**

**Rock Component**

The characteristics and origin of the rock component of rock glaciers can be examined in the light of lithological influences, debris supply processes and subsurface structure. The initial formation of a rock glacier requires thick debris accumulation under a periglacial climate conducive to permafrost. Typical origins of such debris accumulation include debris-laden snow avalanches, episodic rock avalanches and long-lasting rockfall activity, which are triggered by unloading during deglaciation, earthquakes, heavy rains or weathering processes. Further rock glacier development requires continuous debris input from rock walls, which balances the removal of debris as it is advected downslope by the motion of the rock glacier (e.g. Burger et al., 1999; Haeberli et al., 1999; Humlum, 2000). Modelling of rock glacier dynamics should incorporate debris budgets in the rock wall-rock glacier system. This downslope transfer of debris has its own significance as a slope process that can play an important role in the geomorphic evolution of cold-climate, high-relief areas (Barsch, 1977b, 1977c, 1996; Frauenfelder et al., 2003; Humlum, 2000).
Debris input mainly results from physical weathering in rock walls, which periodically releases blocks of rock that fragment into coarse materials as they fall onto talus slopes and rock glaciers. In an early model of rock glacier evolution, Olyphant (1983, 1987) simply assumed the rate of debris input to be a function of a topographic factor, the area of the rock wall above the rock glacier. This simplification is probably appropriate at a local scale where climate and geology can be assumed to be similar. A universal model of the rate of debris input, however, should encompass three kinds of source rock wall characteristics: topographic (height, width, gradient and exposure of the source rock wall), geological (grain sizes and pore structures that affect weathering, joint spacing, continuity and orientation of joints) and climatic factors (thermal and hydrological regimes).

Lithological Controls.

Table 2 highlights geological effects on rock glacier distribution. Four major rock types, granite, gneiss, sandstone and limestone, are commonly found in rock glaciers in the form of boulders or blocky debris. Lithologies that mainly produce finer or platy debris, including schist and shale, are more rarely sources of material for rock glaciers (e.g. Wahrhaftig and Cox, 1959; Evin, 1987). Rock walls composed of the former group of lithologies often feed relatively large rock glaciers that have a blocky outer layer lacking interstitial fine debris, which can be termed ‘bouldery’ rock glaciers. In contrast, the latter group produces much smaller forms (mostly less than 300 m in length) termed ‘pebbly’ rock glaciers, which consist mainly of pebbles and cobbles filled with fine debris (Matsuoka et al., 2005; Figure 6A). Despite their size, the latter group is more susceptible to weathering than the former, so that the debris input to rock glaciers per unit area of rock wall may actually be more rapid. At least two conditions are responsible for small forms. First, mechanically weak rocks can only support low rock walls, which restrict the total debris input despite the higher weathering susceptibility. Where such a lithology constitutes an exceptionally high rock wall, however, large pebbly rock glaciers can develop (Figure 6B). Second, the fine-grained debris that is relatively abundant in the active layer of pebbly rock glaciers is frost-susceptible, and hence is prone to downslope motion by solifluction or debris flows. The resulting rapid downslope transport of debris inhibits debris accumulation. The rounded frontal slope of pebbly rock glaciers may also reflect such surficial soil movements.

Environmental Controls.

Debris production processes can be classified in terms of temporal and dimensional scales (Whalley, 1984; Matsuoka et al., 1998). Weathering associated with diurnal freeze-thaw and thermal fatigue can only be effective on the outermost rock a few centimetres or decimetres thick and causes small rock falls. Weathering associated with seasonal temperature variations, including annual freeze-thaw cycling, can be effective to a depth of a few metres, possibly triggering larger rock falls and potentially removing the entire active layer. Major rock avalanches and rock wall failures can occur progressively over longer time scales as a result of: (1) increasing gravitational stresses by oversteepening the rock wall or decreasing the support provided by permafrost due to lowering of the rock glacier surface; and (2) weakening the rock mass either by increasing stresses inside rock fractures, due to increasing water or ice pressure, or by localised chemical weathering that slowly lowers the fracture toughness of rock (Hallet et al., 1991). Rock avalanches are also commonly observed in association with major earthquakes in mountainous terrain (e.g. Bull et al., 1994; Keefer, 1994; Bull and Brandon, 1998). A question is then raised: which process is most responsible for rock glacier development?

Field observations suggest that seasonal events mainly contribute to talus accumulation, at least in areas that are not subject to frequent and large earthquakes. Boulder falls associated with annual freeze-thaw action are considered to prevail for many rock glaciers, because they are volumetrically

<table>
<thead>
<tr>
<th>Lithology</th>
<th>Major type of surface material</th>
<th>Number of regions</th>
</tr>
</thead>
<tbody>
<tr>
<td>Granite (incl. granodiorite)</td>
<td>Bouldery</td>
<td>16</td>
</tr>
<tr>
<td>Gneiss</td>
<td>Bouldery</td>
<td>14</td>
</tr>
<tr>
<td>Sandstone</td>
<td>Bouldery</td>
<td>13</td>
</tr>
<tr>
<td>Limestone (incl. dolomite)</td>
<td>Bouldery</td>
<td>12</td>
</tr>
<tr>
<td>Schist</td>
<td>Pebbly</td>
<td>10</td>
</tr>
<tr>
<td>Basalt (incl. andesite)</td>
<td>Bouldery</td>
<td>8</td>
</tr>
<tr>
<td>Shale (incl. mudstone, phyllite)</td>
<td>Pebbly</td>
<td>8</td>
</tr>
<tr>
<td>Quarzite</td>
<td>Bouldery</td>
<td>6</td>
</tr>
<tr>
<td>Rhyolite</td>
<td>Pebbly</td>
<td>3</td>
</tr>
<tr>
<td>Conglomerate</td>
<td>Bouldery</td>
<td>3</td>
</tr>
<tr>
<td>Porphyry</td>
<td>Bouldery</td>
<td>2</td>
</tr>
</tbody>
</table>
dominant and most common during seasonal thawing periods (e.g., Church et al., 1979; Matsuoka et al., 1998). Figure 7 shows a large boulder fall that occurred above the Murteil rock glacier in June 1997. The debris volume is equivalent to \(1 \text{ mm} \) of retreat of the whole rock wall, which is comparable with the Late Holocene mean annual retreat rate for this site (Haeberli et al., 1999). Other studies also estimate that rock walls above active rock glaciers have been retreating at rates in the order of millimetres per year, based on estimates of what is needed to sustain the flux of debris in rock glaciers (Barsch, 1977c, 1996; Frich and Brandt, 1985; Humlum, 2000). Moisture and rock joint distribution in the bedrock also affect the production rate of rock fall debris, but little is known about how these factors vary in space and time in any area. Freeze-thaw weathering may be most intensive in a humid zone near the base of the active layer (Hallet et al., 1991), which suggests that the thaw depth affects the maximum size of removable rock. The size of the rock fragments in rock falls depends also on the joint spacing in the bedrock (Matsuoka, 2001). Because joints in schist and shale are closely spaced, for example, short-term temperature cycling (e.g., Matsuoka, 1991), which affects only the outer few centimetres of rock walls, can be important in delivering debris to pebbly rock glaciers where these lithologies occur. Debris is also transferred by secondary processes that remove material already loosened by weathering processes and temporarily accumulated on ledges or in couloirs. These processes include snow avalanches and debris flows, which are of particular significance in transporting fine materials onto rock glaciers.

**Development of the Two-layer Structure.**

Bouldery rock glaciers are composed of outer blocky sediments representing the active layer and an inner matrix-supported frozen core (e.g., Wahrhaftig and Cox, 1959). The outer blocky layer lacks interstitial fine materials and the blocks are generally much larger than those comprising the talus above the rock glacier. The inner frozen core is a mixture of clasts, fine-grained lithic fragments and various types of ground ice (Fisch et al., 1977; Haeberli and Vonder Muhll, 1996; Elconin and LaChapelle, 1997). The
wide range of the grain size distribution reflects the diverse processes by which material is transferred to the rock glacier surface, including rockfalls, debris flows and snow avalanches. Several factors probably contribute to the formation of the outer blocky layer, including fall sorting, by which—during a rockfall event—the larger blocks have sufficient kinetic energy to reach the rock glacier surface while finer clasts are trapped on the talus (e.g. Haeberli et al., 1998). The debris layer eventually covers the larger blocks due to creep, and subsequently new boulders fall and roll onto the finer frozen core. This process develops the two-layer structure and can explain the difference in the size of blocks. Other plausible factors include ‘kinetic sieving’, a phenomenon in which coarse particles in granular mixtures gradually move upward as the mixture is shaken (which for our purposes corresponds to disturbing the outer blocky layer due to deformation of the rock glacier surface) because of the likelihood that small grains can settle further than larger ones (Rosato et al., 1987) and washing away of fine-grained material when permafrost has thawed (e.g. Giardino and Vick, 1987). However, neither of these processes accounts for blocks in the outer layer tending to be larger than those in the talus above the rock glacier and washing appears ineffective on pebbly rock glaciers.

Understanding the genesis of the internal structure is relevant when modelling the debris budget, because it affects the debris flux in the rock glacier.

**Ice Component**

Information about the internal structure of rock glaciers is important but very difficult to obtain. Hence, few direct observations have been reported on ice distribution or ice physical and chemical characteristics, including organic impurities.

**Occurrences, Volumes and Structures.**

The structure of the permafrost core of rock glaciers, as determined by direct observation of borehole cores, large outcrops and tunnels is summarised in Table 3. These observations reveal three general types of ice occurrences: (1) massive ice with dispersed debris and intercalated debris-laden ice, overlying a basal layer of rock; (2) a few metres to tens of metres of debris-laden ice overlying massive ice; and (3) ice/rock mélange throughout (Figure 8).

Ice crystals in massive ice in thin section and in natural, sun-etched exposures consist of fine-to-coarse interlocking grains with diameters of 5 mm (Barsch et al., 1979), 5–10 mm (Wagner, 1990) and 10–30 mm (Elconin and LaChapelle, 1997). Wagner (1990) measured C-axis orientations at a depth of 9.5 m in Murteł rock glacier; the mean dip of the basal planes was 11.5° from horizontal, roughly parallel to the surface slope. Spherical and elongated air bubbles are common in rock glacier ice. Air bubbles 1–2 mm in diameter and elongated up to 20 mm in the direction of flow were reported by Elconin and LaChapelle (1997). The air content in the upper 10 m of a core retrieved from Murteł rock glacier averaged 3% by volume (Wagner, 1990). Alternating layers of bubbly and non-bubbly ice can also be prevalent (e.g. Barsch et al., 1979); Elconin and LaChapelle (1997) reported foliation, defined by layers of bubbly and bubble-free ice, which was strongly associated with other flow-generated structures.

**Chemical and Organic Constituents for Determining Ice Origin and Age.**

18Oxygen and Deuterium: The relationship between these stable isotopes in rock glacier ice has been subject to in-depth analysis in two studies (Stauffer and Wagenbach, 1990; Steig et al., 1998). Results from both suggest that insignificant isotopic fractionation occurred before ice formation, indicating a meteoric or short-residence groundwater source.

Nuclear Bomb Era Radionuclides: Diverse radionuclides in rock glacier ice and meltwater have been analysed, including 210Pb and 238U (Gaggieler et al., 1990), 3H (Wagenbach, 1990), 14C, 37Cl, and 35S (Cecil et al., 1998). These studies suggest that modern precipitation does not infiltrate more than 0.5 m into rock glacier ice, the ice is more than a century old, and that meltwater is probably derived from snow and ice that fell after the end of atmospheric atomic bomb testing.

Chemical Impurities: Chemical analyses of solute and insoluble material from 17 m of core taken from the Murtel rock glacier were performed by Baltsengner et al. (1990). Analysis of ion constituents and concentrations in solution show little if any anthropogenic impact. Extracted insoluble crystalline matter was derived from local bedrock.

Organic Impurities: Stem remains from seven different moss species were extracted from 3 m below the active layer in the Murtel rock glacier (Haeberli et al., 1999). The 14C age of the moss was 2250 ±100 years BP. Twenty-three taxa of pollen and spores extracted from the moss were present in the region between 2000–8000 years BP. These data and other collateral evidence indicate ice aggradation spans the greater part of the Holocene. Konrad et al. (1999) reported the following 14C ages: (1) 200 ±40 years BP of a pine needle ~100 m from the headwall at an undisclosed depth; (2) 2250 ±35 and 2040 ±35
Table 3  Summary of direct observations of active rock glacier internal structures.

<table>
<thead>
<tr>
<th>Exposure type</th>
<th>Internal structure—generalised$^1$</th>
<th>Reference</th>
<th>Rock glacier name and location</th>
</tr>
</thead>
<tbody>
<tr>
<td>Three tunnel levels with several</td>
<td>admixed ice and rock throughout, estimated ice volume 40–70%</td>
<td>Bateman and McLaughlin (1920)</td>
<td>‘Glacier Mine’ (informal), south-central Alaska, USA</td>
</tr>
<tr>
<td>cross-cuts</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>horizontal tunnel</td>
<td>admixed ice and rock throughout, with massive ice near lateral margin</td>
<td>Brown (1925)</td>
<td>‘Hurricane Basin’ (informal), Colorado, USA</td>
</tr>
<tr>
<td>~100 m long and ~30 m below surface</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>borehole, 10.4 m</td>
<td>stratified debris-laden ice and massive ice, estimated average ice volume 50–60%</td>
<td>Barsch (1977a)</td>
<td>‘Murtèl I’, Swiss Alps</td>
</tr>
<tr>
<td>outcrop, 10 m deep</td>
<td>Ice/rock mélange with deformed ice lenses, high ice content</td>
<td>Fisch et al. (1977)</td>
<td>‘Comb de Prafleuri’ (informal), Swiss Alps</td>
</tr>
<tr>
<td>borehole, 7 m deep</td>
<td>stratified debris-laden ice and massive ice, estimated average ice volume 50–70%</td>
<td>Barsch et al. (1979)</td>
<td>‘Gruben’, Swiss Alps</td>
</tr>
<tr>
<td>Three boreholes to bedrock, 50–60 m</td>
<td>massive ice in upper 25 m, with ice volumes of 90 to 100%, debris-laden ice and massive ice layers at intermediate depth (~7 m thick), and rock with interstitial ice to bedrock</td>
<td>Haeberli et al. (1988); Haeberli (1990); Arenson et al. (2002)</td>
<td>‘Murtèl I’, Swiss Alps</td>
</tr>
<tr>
<td>deep</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>outcrop, longitudinal section</td>
<td>debris-laden ice upper 2–3 m overlying massive ice 3–11 m thick with scattered rock fragments</td>
<td>Moore and Friedman (1991)</td>
<td>‘Berthoud Pass’ (informal), Colorado, USA</td>
</tr>
<tr>
<td>&gt;80 m long, max. thickness 14 m</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Four boreholes</td>
<td>admixed rock and ice up to 24 m thick; drilling stopped in ice laden with fine grained debris</td>
<td>Hamre and McCarty, 1996; Hamre et al. (2000)</td>
<td>‘East Lone Mountain’, Montana, USA</td>
</tr>
<tr>
<td>outcrop, complete transverse section</td>
<td>Ice/rock mélange throughout with folded strata, ice volume &gt;50%</td>
<td>Elconin and LaChapelle (1997)</td>
<td>‘Fireweed’ (informal), south-central Alaska, USA</td>
</tr>
<tr>
<td>&gt;90 m wide, max. thickness 60 m</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>several boreholes</td>
<td>massive ice up to 25 m thick with dispersed debris overlying unknown thickness of rock and ice</td>
<td>Konrad et al. (1999)</td>
<td>‘Galena Creek’, Wyoming, USA</td>
</tr>
<tr>
<td>Four boreholes to bedrock, 64–72 m</td>
<td>layers of ice and rock, estimated ice volume 40–70%</td>
<td>Arenson et al. (2002)</td>
<td>‘Muragl’, Swiss Alps</td>
</tr>
<tr>
<td>deep</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

$^1$ The term ‘admixed ice and rock’ is used when field evidence of a sheared matrix is not given; ‘melange’ is used when field evidence of a sheared matrix is given.
years BP from unspecified organic material in ice at the surface ~400 m from the accumulation zone; and (3) 1680 ± 70 years BP from leaf fragments in the upper 0.2 m of ice halfway down the rock glacier.

Ice Origin: Diagnostics.
Differentiation between ice of glacial or non-glacial origin in permafrost is usually difficult because the two types share petrographic, stratigraphic, physical and chemical qualities (cf. Mackay, 1989; French and Harry, 1990). Transformation of original structures due to flow (i.e. flow metamorphism) in rock glacier permafrost and the close proximity of some rock glaciers to glacierets and small glaciers compound this problem.

Petrofabrics: Glacier-like ice crystal morphology, orientation and air bubble content observed in rock glaciers do not indicate glaciogenic ice, as is sometimes assumed. Post-depositional processes, most notably flow metamorphism, lead to glacier-like crystal morphology and orientation but are independent of ice origin. Spherical and elongate air bubbles, in ordered or dispersed arrangements, are common in both types of ice. Bubble foliation in glacier ice is caused primarily by flow metamorphism (e.g. Hambrey, 1979), and by analogy, so is bubble foliation in rock glaciers (Elconin and LaChapelle, 1997).

Stratigraphy: Small- to large-scale depositional and deformational structures composed of massive and debris-laden ice, formed in glacial and non-glacial environments, can appear very similar.

Water Quality: A common assumption is that ice formed in the non-glacial environment is composed of ion-rich groundwater, while glacier ice is composed of relatively pure meteoric water. The ionic content of glacial ice and of short-residence groundwater can be quite similar, however. In addition, rock glacier ice is likely to be polygenetic and may include significant meteoric contributions. Moreover, debris-rich ice typically from the basal portion of glaciers shares many characteristics with diverse forms of ground ice.

Ice Origin: Accumulation Area (‘Rooting Zone’ or ‘Source Area’).
Typically, the head of a rock glacier is located at the base of a bedrock headwall and is the primary accumulation site for both rock debris and ice. Due to dynamic and diverse atmospheric, hydrologic and geologic conditions in alpine environments, a wide range of ice formation processes can exist in close proximity in space and time, particularly in the accumulation zone. Observable types of ice formation in the accumulation zone at the surface are related to snow metamorphism and surface icings. In the subsurface, negative mean annual ground temperatures and abundant liquid water assure ice growth in rock debris in the form of interstitial, segregation, injection or vein ice (Wahrhaftig and Cox, 1959; Wayne, 1981; Haeberli and Vonder Mühll, 1996). Sources for the liquid water include groundwater, rain and meltwater from snow, avalanche runout, firn and ice.

GEOMETRY AND KINEMATICS

There is a long tradition of geodetic and photogrammetric surveying of the effects of permafrost creep within active rock glaciers (Kääb, 2005). Available measurements indicate that the surface movements of rock glaciers are characteristically slow (typically decimetres per year) and that they form a continuous and internally coherent flow field as expected with thermally controlled, macroscopically homogeneous permafrost. It is also evident from the surface flow fields that the debris transported on the surface of rock glaciers originates from talus and other accumulations, such as moraines or other heaps of loose debris, and most likely over time scales of millennia (Figure 9). The precise determination of surface flow fields in rock glaciers also provides a context for subsurface information obtained from boreholes and geophysical soundings, and helps to determine the degree to which this subsurface information is representative spatially.

Optimal investigation of permafrost creep requires:

• spatially distributed information on kinematics to account for three-dimensional effects;

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DOI: 10.1002/ppp
Monitoring Techniques

Ground-based approaches as well as air- and space-borne ones have been applied for monitoring rock glacier kinematics with high resolution. Ground-based surveys utilise triangulation and laser ranging (e.g. Haeberli, 1985; Zick, 1996; Sloan and Dyke, 1998; Koning and Smith, 1999; Konrad et al., 1999; Krainer and Mostler, 2000), or satellite geodesy (e.g. GPS; Berthling et al., 1998). Other methods for deformation measurements, such as steel tapes or strain wires, are seldom used (e.g. Haeberli, 1985; White, 1987). Terrestrial laser scanning provides local digital terrain models (DTM) with very high resolution and accuracy (Bauer et al., 2003), and airplane-based laser scanning systems have recently been used to generate a DTM with a resolution approaching 1 m for rock glaciers in the Dry Valleys of Antarctica and on Mt St Helens, USA (http://www.nasa.gov/vision/earth/lookingatearth/mshelenslidar.html).

Photogrammetry is the remote sensing technology most commonly used for rock glacier monitoring. Rock glacier thickness changes and surface displacements can be measured from repeated stereo-imagery (e.g. Haebelri and Schmid, 1988; Kääb et al., 1997, 1998; Kaufmann, 1998a, 1998b; Kääb, 2002, 2005). Recent advances, especially in image processing, permit the measurement of surface motion on rock glaciers with unprecedented resolution, accuracy and spatial coverage, which has led to a number of new insights into rock glacier development and behaviour (Kääb and Volme, 2000; Kaufmann and Ladstaedter, 2002; Kääb et al., 2003; Figure 10). Space-borne differential synthetic aperture radar interferometry (DInSAR) is also able to measure rock glacier surface displacements, especially those in the vertical plane, with an accuracy of a few millimetres to centimetres (Rott and Siegel, 1999; Kenyi and Kaufmann, 2003; Nagler et al., 2001; Rignot et al., 2002; Strozzi et al., 2004).

Dynamic Processes

The main processes related to the evolution of a rock glacier can be identified from high-resolution

measurements of three-dimensional surface velocity fields. The fundamental process is creep within the permafrost. Nearly all high-resolution studies on the deformation of (visually) active rock glaciers have been able to detect creep. Whereas maximum speeds are in the order of several metres per year, the detection of minimum speeds seems mostly restricted by the available measurement accuracy. From comparison of available studies on rock glacier speed, it is clear that differences in slope, thickness, temperature, or internal composition are not able to explain differences in speed at individual sites in any straightforward way. On the other hand, within a single rock glacier, the spatial pattern of surface speeds often seems simply related to the pattern of surface slopes (Konrad et al., 1999). ‘Cold’ polar rock glaciers, in general, creep more slowly than ‘warm’ Alpine ones (Kääb et al., 2002). Relatively low speeds at the rock-glacier root zones and margins probably reflect thinner deforming layers (Kaufmann, 1998a; Kaufmann and Ladstaedter, 2002). Separation of individual influences on surface speeds is difficult due to the usual lack of knowledge about the internal structure (thickness and composition).

The advection of surface micro-topography by creep may result in a pattern of local positive and negative thickness changes. A pattern of heaving in front of individual transverse ridges and corresponding lowering behind them has been observed from high-resolution studies (Kääb et al., 1998; Kääb and Vollmer, 2000; Kääb and Weber, 2004). Comparing the observed vertical rates with ones calculated from creep speed and surface slope clearly confirms these local patterns of surface elevation change. Three-dimensional straining may lead to local heaving or thinning. For an incompressible medium such as pure ice, compression and extension should equal out. Horizontal compression of the frozen debris is accompanied by vertical extension, that is by surface heaving and vice-versa. Such relationships between horizontal and vertical strain rates can be applied for rock glaciers where ice contents exceed saturation (Kääb et al., 1998; Kääb and Vollmer, 2000; Rignot et al., 2002) and where ground ice is not highly aerated or crevassed.

General thaw settlement and frost heave as an expression of thermal disequilibrium (active-layer thickness > debris cover thickness) rather than a climate signal (Haeberli and Vonder Mühll, 1996; Kääb and Haeberli, 2001; Kääb and Reichmuth, 2005). According to the available monitoring series, such small-scale differential melting or frost heave is clearly an exception and usually represents rather a disturbance of the general thermal equilibrium. Widespread differential melting likely occurs only at the front where the ice content melts out (Kääb et al., 1997) or in areas of preferential water flow. In fact, for most monitored rock-glacier mass changes, no clear signal of overall mass gain or loss has been observed. Only the study of the Murtel rock glacier clearly showed an overall mass loss of a few centimetres per year (Kääb et al., 1998).

Little is known about temporal changes in rock glacier creep rates. For some rock glaciers or parts of them, current surface velocity fields clearly differ from those of the past (Frauenfelder and Kääb, 2000). Over shorter time scales, surface speeds vary both cyclically and non-cyclically. Clear seasonal variations in speed have been observed (Haeberli, 1985; Arenson et al., 2002; Kääb et al., 2003). Non-cyclic pluriannual changes in speed might result from external (climate?) forcing (Zick, 1996; Kääb et al., 1997; Kääb and Frauenfelder, 2001; Schneider and Schneider, 2001).
Monitoring of rock glacier velocities to date has revealed several possible causes, mostly connected to impacts of climate variations at the scale of decades and longer time periods.

**Perspectives**

The technology for monitoring rock glacier dynamics is comparably well developed. Research limitations primarily result from the limited availability of advanced techniques, the difficulty of developing and implementing systematic monitoring strategies, and the inherent difficulty of making subsurface measurements of rock glacier motion. Current research tendencies point towards an increase in accuracy and temporal and spatial resolution, with an improvement in measurement automation, and towards enhanced application of space-borne techniques in order to cover remote areas (Kääb, 2005). The knowledge of geometry and surface kinematics of rock glaciers is quite detailed, and the primary process responsible for the creeping motion of rock glaciers is relatively well understood to be due to creep deformation in the ice, whereas analyses of the dynamics of rock glaciers remain elusive. Displacement rates cannot be quantified because, contrary to ordinary glaciers that are comprised of relatively homogeneous ice, rock glaciers tend to consist of inter-layered debris-rich and ice-rich domains, and in general, information about the spatial extent and rheological behaviour of these domains is insufficient to permit detailed flow calculations. Moreover boundary conditions equivalent to specifying accumulation and ablation rates are difficult to define, especially in view of the multiplicity of inputs of ice and debris in rock glaciers. The general lack of information on the internal composition and architecture of rock glacier together with the complicated and poorly-defined rheological behaviour of icy debris and debris-rich ice limit attempts to quantitatively understand velocity variations, sensitivity to external forcing, rock glacier advance mechanisms, and the development and dynamic interpretation of microtopography. A large number of globally distributed monitoring series could substantially improve our understanding of the basic processes of rock glacier creep with their local and temporal variabilities.

**CREEP OF ROCK GLACIERS**

The creation of a unified constitutive model for permafrost found in alpine rock glaciers, to represent creep and failure in variable temperature regimes, remains a distant goal. The only reliable way at present is to carry out analyses of individual components of rock glacier with simplified geometry and assumptions. Examples include separate consideration of: (1) deformation within the active layer which is very strongly linked to pore pressures (e.g. Arnold et al., 2005); (2) creeping of the main body of the rock glacier, which can be approximated using appropriate creep parameters for the relevant stress state and temperature (Arenson and Springman, 2005b); or (3) rheological response to time-dependent changes in temperature and hence state. All of these interact and suitable parameters must be selected accordingly. We may also assume that the slope is infinite, or focus on more local viscous features, for example on the surface of the rock glacier, depending on our specific area of interest.

The rheological behaviour of rock glaciers is highly sensitive to the character and temperature of their deforming layers, which usually comprise mixtures of ice and rock elements, the latter ranging from silty sand up to large boulders. In addition, unfrozen water may be present and influence responses, at permafrost temperatures close to the melting point of the ice, in particular at the base of the rock glacier. As in the mechanics of rock masses, large-scale behaviour of such mixtures is difficult to evaluate either by direct in situ measurements or by taking samples, but it can be estimated by a constitutive modelling approach.

Modelling the entire creep process, following instantaneous elastic and plastic deformation due to initial load application, including primary (e.g. see Vyalov et al., 1962; Domaschuk et al., 1991; Wijeweera and Joshi, 1991, for specific models), secondary (e.g. Andersland and AlNouri, 1970) and tertiary creep, is now possible for ice or related homogeneous mixtures according to developments by Goughnour and Andersland (1968), Fish (1983, 1984), Ting (1983), Zhu and Carbee (1983), or Gardner et al. (1984). However these models are less suitable for rock glaciers because of their heterogeneity. Moreover, some models only acknowledge tertiary creep and for others, the secondary (steady state) creep phase is reduced to an inflection point, that is the decrease of the strain rate with time is directly followed by acceleration, without an interval at a constant minimum strain rate. While all these deformation phases are important for certain practical problems, such as foundations in permafrost, and are well covered in the literature (e.g. Andersland and Ladanyi, 2004), they are of less importance in connection with rock glaciers, where secondary creep (at a minimum strain rate) should dominate.
Glen’s (1955) original equation for secondary creep of polycrystalline ice is often used for the creep of permafrost and other polycrystalline solids:

\[ \dot{\varepsilon} = B\sigma^n \]  

(1)

where \( \dot{\varepsilon} \) is strain rate, \( \sigma \) is a corresponding stress, and the coefficients \( B \) and \( n \) are material constants. \( B \) contains the effect of temperature, grain size, fabric and impurity content; it can be written explicitly as (e.g. Andersland and Ladanyi, 2004):

\[ B = \frac{\dot{\varepsilon}_c}{\sigma_{c\theta}} \]  

(2)

where:

\[ \sigma_{c\theta} = \sigma_{co}\left(1 + \frac{\theta}{\theta_c}\right)^w \]  

(3)

Here, \( \dot{\varepsilon}_c \) denotes the reference strain rate, \( \sigma_{c\theta} \) is the reference stress or creep modulus, for temperature \( \theta = -T \), and \( \theta_c \) is an arbitrary reference temperature (see data for polycrystalline ice from Morgenstern et al. (1980) \( \dot{\varepsilon}_c = 10^{-5} h^{-1}, \sigma_{co} = 0.103 \text{ MPa}, \) and \( w = 0.37 \) valid for \( T = -1^\circ \text{C}, \) and \( 10^{-7} \leq \dot{\varepsilon} \leq 10^{-2} h^{-1} \)). Many authors (Vyalov et al., 1962; Fish, 1983, 1984; Ting, 1983; Zhu and Carbee, 1983; Gardner et al., 1984; Wijeweera and Joshi, 1991) have reported results consistent with Equation (1) or similar expressions from uniaxial creep tests on cylindrical samples. Goughnour and Andersland (1968), Andersland and AlNouri (1970) and Domaschuk et al. (1991) have done so based on triaxial test data, which have the potential to represent the in situ stress state more faithfully, although size effects (particle sizes related to sample size) and sample disturbance must be considered and minimised where possible, by increasing sample size, using appropriate cooling fluids, and cooling the drill bit during drilling and extraction.

Figure 12 shows two different samples post-creep (volumetric ice content: 55% (no. 27) and 74% (no. 43)). The initial sample diameter was 74 mm and the initial sample height was approximately 150 mm. Significant compression, that is volume decrease, was recorded for sample no. 43, which also had a higher volumetric air content (26%, Arenson, 2002).

The homogeneous approach presented above, with averaged creep parameters derived from small-scale tests, works very well for uniform ground conditions found for some arctic permafrost (in certain areas—e.g. Morgenstern et al., 1980; Ladanyi, 2002; Sego et al., 2003). Recent geotechnical and geophysical investigations and monitoring at Muragl, Murtel-Corvatsch and Pontresina Schaflberg based on earlier work (e.g. Haeberli et al., 1998; Kääb and Vollmer, 2000) have confirmed, however, that the internal structure of a rock glacier is extremely heterogeneous (Arenson, 2002; Musil, 2002; Maurer et al., 2003), leading to significant variations in the parameters back-fitted to standard creep expressions (Arenson and Springman, 2005a, 2005b), or requiring explicit theoretical simulation. For example, Ladanyi (2002, 2003) and Ladanyi and Archambault (2003) attempted to simulate the creep behaviour of ice/rock mixtures and ice-filled irregular rock joints under variable stress and temperature conditions.
Pressuremeter tests allow creep parameters to be obtained in situ (e.g. in arctic permafrost or ice; Ladanyi and Johnston, 1973; Kjartanson et al., 1988; Ladanyi and Melouki, 1993; Ladanyi, 1996), as well as data on load controlled (relaxation) tests. They are promising, even for alpine permafrost (Arenson et al., 2003b), because the internal conditions (in situ stress and temperature) can approach reality if the pressuremeter probe is pre-cooled prior to insertion in the borehole shortly after completion of drilling to test depth using pre-cooled rotary boring techniques rather than percussion methods. Nonetheless, the volume of permafrost investigated by this technique in the annulus around the probe is still very limited when compared to the volume of a rock glacier. In addition, temperature disturbances due to the drilling have to be considered and the creep parameters derived can vary significantly.

Temperature probably has the most significant influence on creep behaviour. The minimum strain rate approached during the secondary creep phase increases with increasing temperature. The Arrhenius equation may represent this behaviour acceptably at temperatures colder than about $-10^\circ$C (e.g. Mellor and Testa, 1969; Hooke et al., 1980). At temperatures close to the melting point of ice, which are important in terms of degrading permafrost, the unfrozen water changes the behaviour (Barnes et al., 1971). Voytkovsky (1960) presents an empirical relationship that was successfully applied (e.g. Hooke et al., 1980; Arenson and Springman, 2005b).

The rheological behaviour of sand/ice mixtures also depends on the applied normal stress. The effect can be estimated using the methods presented, for example, in Andersland and Ladanyi (2004). Arenson and Springman (2005a, 2005b) report results from triaxial constant strain rate and constant deviatoric stress tests on ‘undisturbed’ samples from Muragl and Murtel-Corvatsch (Figure 12) dependent upon a constant confining cell pressure ($\sigma_3$), deviatoric stress ($\sigma_1 - \sigma_3$), temperature and volumetric solids content (Figure 13). The data confirm that warm, ice-rich permafrost under higher normal stresses will exhibit greater steady-state strain rates.

Within many of the rock glaciers examined to date, significantly higher local strain rates occur within a thin ‘shear’ zone, which may be 1–3 m or more in thickness. The depth of this zone and the magnitude of strain rate are strongly dependent on the internal structure (% ice, % soil, including typical particle sizes) and local temperature, which is often very close to $0^\circ$C (e.g. Haebelri, 1985; Wagner, 1992; Haebelri et al., 1998; Arenson et al., 2002). Usually there is virtually no deformation below this shear zone and relatively little creep deformation above it.

Strain rates in shear zones within 20 m of the surface can be influenced by seasonal temperatures, so that seasonal changes in the surface velocity have been detected for a rock glacier with changes exceeding 0.2 m/a, which reflect temperature variations over the seasonally affected zone (Arenson et al., 2002; Kääb et al., 2003). Seasonal changes in ground temperature are not only generated from changes in surface air temperature, but also from convective heat flow within permeable taliks or layers at the permafrost base. Shear deformation is generally accompanied by either dilation or contraction depending on the structure of the permafrost, but certainly does not occur at constant volume as is often assumed in rock glacier dynamics. Structural hindrance and hence dilation will occur if the debris content exceeds about 35% (e.g. Goughnour and Andersland, 1968; Ting et al., 1983) and the matrix is virtually air-free, whereas contraction will occur if there is a significant amount of air, for example > 5% present, which is not untypical for some rock glaciers (Wagner, 1990; Arenson et al., 2003a, 2004; Yasufuku et al., 2003). This will affect the volume of permafrost lying in and above the shear zone and the predictions of vertical and horizontal strain.

Numerical modelling of the four-phase interaction between solids, water, ice and air may be achieved using discrete elements (Arenson and Sego, 2005) in the future but the constitutive models will need to be developed at a particulate level while computer capacity must increase by another order of magnitude to make this feasible. Attempts to back-fit average creep parameters to a time series of deformation data, for example by treating the rock glacier as a homogeneous medium (e.g. Wagner, 1992), gave a useful perspective at the global scale but inevitably failed to represent the detailed deformation field that is necessary for in-depth understanding of rock glacier dynamics and for risk assessments in combination with permafrost degradation.

CONCLUSIONS AND RECOMMENDATIONS

Research on permafrost creep and the kinematics of active rock glaciers in cold mountainous areas has advanced rapidly during recent decades and continues to expand. This results from the application of sophisticated technology in comprehensive field and laboratory experiments, and especially from increasing cooperation among international specialists in geomorphology, microclimatology, glaciology,
geophysics, geodesy/photogrammetry and engineering. The resulting information provides a clearer picture concerning the basic materials, processes, phenomena and spatio-temporal scales involved in the motion of rock glaciers and associated features that also result from the gravitationally-induced deformation of subsurface ice.

Thermal conditions remain the fundamental boundary condition at the surface and define spatial scales (extent, distribution patterns and thickness of permafrost) as well as temporal dimensions (formation, preservation and melting of ground ice under conditions of Holocene climates and recent atmospheric warming). Surface temperatures affect the permafrost through a highly heterogeneous active layer commonly exhibiting large, wide-open voids, making conductive, convective and advective energy fluxes extremely complex and strongly dependent on snow-cover characteristics. The coarse blocks of this active layer, with its special thermal conditions, result

Figure 13  Triaxial creep tests at confining pressure $\sigma_3 = 200 \pm 0.4$ kPa for different volumetric ice contents, $w_i$, and temperatures, $T$; samples 1, 2, 4, 9 and 11 were artificially frozen; (A) $w_i = 15–33\%$; (B) $w_i = 61–78\%$; and (C) $w_i = 85–100\%$ (Arenson and Springman, 2005b).
from weathering, rockfall activity and grain-size sorting at the rock headwall of the creeping ice/rock mixtures. Ground ice is most likely to be polygenetic and generally far exceeds the pore space of the talus within which it forms. Massive subsurface ice sometimes occurs and may have formed below the surface or could have originated as surface ice from avalanche cones, perennial snowbanks, glacierets or small (mainly cirque) glaciers, and was subsequently buried at negative ground temperatures in debris either derived from valley walls or accumulating on the rock glacier surface from ablating ice. The coherent pattern of the surface motion evident in the continuous surface structures commonly seen on rock glaciers reflects high ground-ice contents. Such motion continues over millennia as indicated by high-precision geodetic/photogrammetric surveys and radiocarbon dating. Due to the striking heterogeneity of the material and differences in ice content and internal architecture between rock glaciers, however, the rheology generally remains difficult to define universally. In addition, a number of aspects of the flow and evolution of creeping permafrost and rock glaciers are poorly understood; these include short-term velocity fluctuations, the occurrence of thin but effective shear zones, extrusion flow and time-dependent volume changes.

There is a clear need for integrated numerical modelling of rock glacier development and evolution, as well as dynamics. However, this will require better characterisation of rock glacier composition and internal structure, and better understanding and more rigorous modelling of specific processes. Such quantitative holistic treatment is a difficult challenge and must be based on diverse data available from photogrammetry, data logging, borehole observations and laboratory experiments. It should include and combine partial models for:

(a) rock weathering, snow avalanches and rockfall events with grain-size sorting and mass wasting on scree slopes;
(b) freezing processes and ice formation in sub-zero scree, with abundant fine material as well as coarse blocks;
(c) coupled thermohydro-mechanical modelling of creep and failure processes in frozen rock debris, including the entire range of particle sizes from fine to coarse, and the entire range of ice contents from little or no ice in a dense interlocking matrix to massive subsurface ice with highly dispersed debris and significant air content;
(d) kinematics of non-isotropic, heterogeneous and layered, ice-rich permafrost on slopes with long transport paths for coarse surface material from the headwall to the front and, in some cases subsequent re-incorporation into the advancing rock glacier causing corresponding age inversion at depth; and
(e) dynamics of rock glaciers, which include spatial and temporal variations in velocity that are manifested in the ridges, furrows and other surface structures typical of rock glaciers, as well as their down-valley motion.

It was the clear goal of the IPA Task Force to concentrate on purely periglacial features and processes, which predominate in most regions. As a next step, however, interactions and feedbacks between snow, glaciers and frozen scree on mountain slopes should be considered in order to understand more complex landforms. In particular, the question of a gradation or continuum between rock glaciers and debris-covered glaciers should be examined in view of ground thermal conditions and the common contrast in surface character of these features: the transverse ridges, deep furrows and the steep, near-angle of repose terminus, typical of perennially frozen rock glaciers and the ablation pits and lakes typical of debris-covered glaciers.

Future studies of rock glaciers should be closely linked to other aspects of permafrost science, and should benefit from major recent advances in understanding the physical and chemical processes that operate in all permafrost regions (e.g. Hallet et al., 2004). These have resulted from diverse approaches that range from examining fundamental periglacial phenomena at the microscopic scale to recognising the importance of permafrost processes on climate change at the global scale. We firmly believe that significant advance in understanding most periglacial phenomena requires the close integration of field and theoretical studies, and will greatly benefit from ideas and techniques in other disciplines within the geosciences and beyond. Much is to be gained from cross-fertilisation between scientists and engineers as well. For example, the transfer of energy through a surface layer of coarse rock debris is central to understanding the thermal regime of rock glaciers. It is, however, poorly understood and little has yet been gained from the extensive studies of a very similar engineering problem associated with the use of artificial rock cover to avoid warming or melting permafrost. Another example is using what has been learned through extensive empiricism in engineering studies about frost resistance of concrete and other porous materials to provide a basis for further studies of the physics of frost weathering, which is one of the controls on the debris supply to rock glaciers.
Future studies should also appeal to the broad scientific audience that is now aware and interested in permafrost as a critical element of the Earth’s cryosphere particularly in light of its vulnerability to atmospheric warming and potential feedbacks to global climate. In addition, they should capitalise on the growing interest from the engineering and planetary communities, respectively, in geotechnical issues characteristic of cold regions and in diverse aspects of permafrost on Mars and other icy planets (e.g. Whalley and Azizi, 2003; Head et al., 2005).

REFERENCES


