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Effects of climate change on mass movements in mountain environments

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Abstract

Changes in temperature and precipitation have a range of impacts, including change of glacier extent, extent and duration of snow cover, and distribution and thermal properties of permafrost. Similarly, it is likely that climatic changes affect frequency and magnitude of mass movements, such as shallow landslides, debris flows, rock slope failures, or ice avalanches. However, so far changes in mass-movement activity can hardly be detected in observational records. In this progress report we document the role of climate variability and change on mass-movement processes in mountains through the description and analysis of selected, recent mass movements where effects of global warming and the occurrence of heavy precipitation are thought to have contributed to, or triggered, events. In addition, we assess possible effects of future climatic changes on the incidence of mass-movement processes. The report concentrates on high-mountain systems, including processes such as glacier downwasting and the formation of new ice-marginal lakes, glacier debuttressing and the occurrence of rock slope instability, temperature increase and permafrost degradation, as well as on changing sediment reservoirs and sediment supply, with a clear focus on studies from the European Alps.

Keywords

Alps, climate change, glacier, hazard, landslide, mass movement, permafrost

I Introduction

At present, the level of confidence is high that the global average temperature of the past few decades was warmer than any comparable period during the last 2000 years (Mann et al., 2008). Current evidence also suggests that temperatures during the past 25 years were higher than any period of comparable length since AD 900 for many, but not all, regions (Borgatti and Soldati, 2012; IPCC, 2007). The Fourth Assessment Report (AR4) of the Intergovernmental Panel on Climate Change (IPCC) also reports that most land regions of the world show an increase of high temperatures for the past 50–100 years, usually expressed as the 90th or 95th percentile of the long-term record (Trenberth et al., 2007). In Europe, the frequency of hot days has almost tripled during the period 1880–2005 (Della-Marta et al., 2007), whereas Ding et al. (2010) and Kunkel et al. (2010) found a strong increase in the number of heat waves since the

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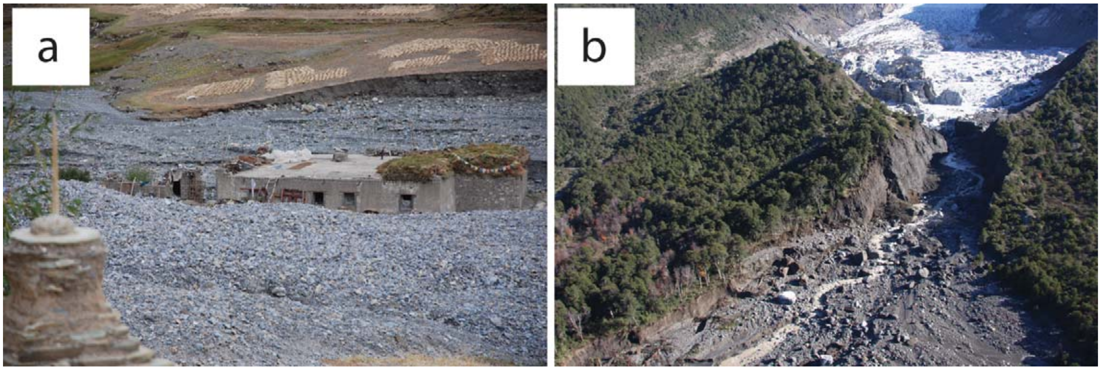

Figure 1. Global warming is likely to lead to a greater frequency and magnitude of heavy precipitation events. More and more intense rainfalls may trigger – among other processes – (a) flash floods and/or debris flows in mountain areas such as in Ladakh (India; photo: Raphael Worni) or (b) overflow and breaching processes in moraine-dammed lakes (Ventisquero Negro glacier, Bariloche, Argentina; Worni et al., 2012).

1960s for China and the USA. Based on simulation runs forced with different greenhouse gas emission scenarios, the IPCC (2007) concludes that the rate of warming until the end of the 21st century is likely to be faster than ever recorded from historical or proxy records. Although changes in average conditions may have serious consequences by themselves, the key impacts of climate change will be felt due to changes in interannual climate variability and weather extremes (see Borgatti and Soldati, 2012; IPCC, 2007). In high-mountain regions, the evolution of mean and extreme temperatures will likely be comparable to what has been described above; however, studies specifically focusing on trends at high elevations have not been published so far.

The capacity of air to hold moisture is a function of temperature. As a consequence, global warming is likely to lead to an overall greater frequency and magnitude of heavy precipitation events (Fowler and Hennessy, 1995). Despite the fact that the projection of the occurrence of atmospheric phenomena of relatively small extent, such as storms, tend to suffer from greater uncertainty than is the case for regional atmospheric patterns, an increase in the frequency and intensity of extreme precipitation events has been identified in different sets of observational data from several regions of the World (IPCC, 2007; Figure 1a). For the future, projections likewise suggest decreasing return periods of extreme rainfall events (Christensen and Christensen, 2003; Kharin et al., 2007; Kyselỳ and Beranová, 2009; Orlowsky and Seneviratne, 2012).

Changes in temperature and precipitation are considered likely to have a range of secondary effects, including on the extent of glaciers, the distribution and duration of the snow cover, and on the temperature and three-dimensional distribution of permafrost. However, while there is theoretical understanding for increased mass-movement activity as a result of predicted climate change in mountain environments, changes in activity can hardly be detected in observational records. In addition, uncertainty remains considerable as a result of error margins inherent in scenario-driven global predictions, and due to the lack of spatial resolution of downscaled projections (Crozier, 2010). At lower elevations, it has also been reported that both the frequency and magnitude of landslides could decrease as a result of climate change (e.g. SE England, Collison et al., 2000; Dehn et al., 2000).

This progress report therefore aims at documenting the role of climate variability and change on mass-movement processes in mountainous regions through (1) the description and analysis of selected, recent mass movements where effects of global warming and the occurrence of heavy precipitation are thought to have contributed to, or triggered, events. We then address (2) possible effects of future climatic changes – as projected by Global Circulation Models (GCM) and Regional Climate Model (RCM) runs – on the occurrence of future mass-movement processes, and (3) speculate about possible consequences of climate and mass movements on hazards and risks. Most examples illustrated in this report are from the European Alps with a clear focus on case studies from high-elevation sites in Switzerland.

II Glacier downwasting and the formation of new ice-marginal lakes

One of the most obvious consequences of climate change at high-elevation sites is the widespread retreat and disintegration of glaciers (e.g. Diolaiuti et al., 2011; Zemp et al., 2007). The consequences for natural hazards following increasingly rapid changes in glacier geometry are multiple and include the formation of ice-marginal lakes, ice avalanches and mass movements originating from the recent debuttressing of previously glacierized walls and hillslopes.

A prominent phenomenon associated with glacier retreat and changes in glacier geometry is the formation and growth of ice-marginal lakes. Glacial lakes have been classified into several types according to their position relative to the glacier and the damming mechanism (Clague and Evans, 2000; Richardson and Reynolds, 2000; Roberts, 2005). The different lake types are more or less frequent in different regions of the world, depending on climatic, glaciologic, topographic, geologic and other factors. Hazards related to glacier retreat and the formation of glacial lakes have been recognized for several decades (Evans and Clague, 1994). Indeed, severe disasters have occurred in the past as a result of outburst floods from glacial lakes in various high-mountain regions of the world, including the Andes (Carey, 2005; Hegglin and Huggel, 2008; Reynolds et al., 1998; Worni et al., 2012; Figure 1b), the Caucasus and Central Asia (Aizen et al., 2007; Narama et al., 2010), Hindukush-Himalayas (Richardson and Reynolds, 2000; Vuichard and Zimmermann, 1987; Xin et al., 2008), North America (Clague and Evans, 2000; Kershaw et al., 2005) and the European Alps (Haeberli, 1983; Haeberli et al., 2001).

Rapid lake formation and growth that has been accelerated in recent years is generally a global phenomenon but has not been observed everywhere with the same level of detail. Some of the best documented recent developments are from the Swiss Alps (Künzler et al., 2010; Werder et al., 2010), such as the Trift, Lower Grindelwald, Chüeboden and Plaine Morte (Bernese Alps), Rhone (Valais Alps) and Palü glaciers (Grisons Alps). These lakes have formed within the past decade and are all located at the terminus of glaciers where subglacial topography has been overdeepened by the glacier. At Trift glacier (Figure 2), positive feedback processes, mainly related to the thermal energy of water, accelerated glacier melt and resulted in the formation and extensive growth of the proglacial lake in only three years (Kääb and Haeberli, 2001). Although the Trift lake has become a major tourist attraction, there is considerable concern about potential hazards in case of a lake outburst that could be triggered by ice avalanches from the ice fall zone of Trift glacier or rockfalls following debuttressing of the steep lateral slopes (Dalban Canassy et al., 2011).

An even more dramatic situation than at Trift has developed at Lower Grindelwald glacier (Figure 3a). This glacier has been in strong

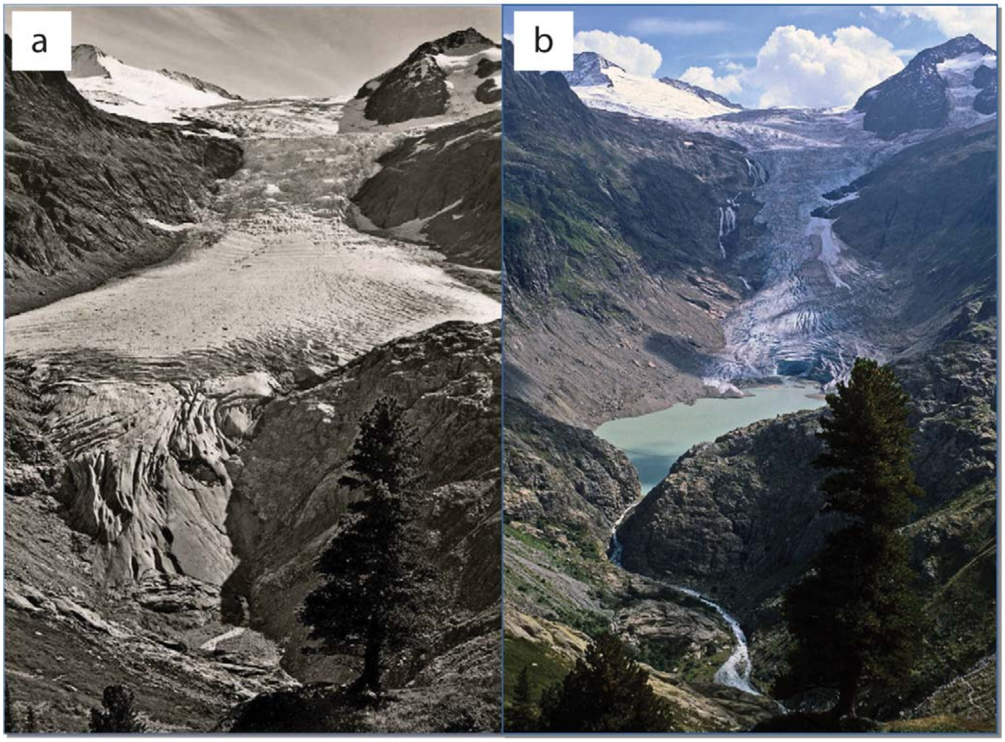

Figure 2. Comparison and evolution of the terminal part of Trift glacier in the Central Swiss Alps. (a) In 1948 massive glacier ice covered the glacially overdeepened trough (photo: Gesellschaft für ökologische Forschung). (b) By 2006 the ice of the glacier terminus part had almost completely gone, leaving behind a large lake. Due to potential flood hazards, the lake is regularly monitored.

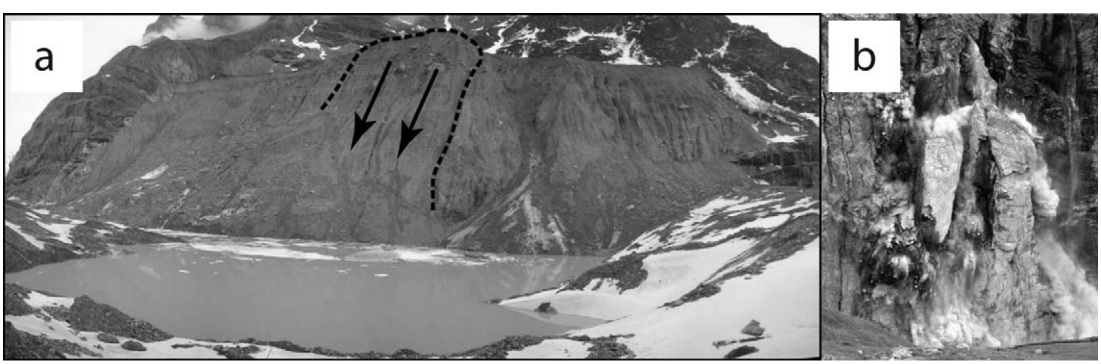

Figure 3. (a) The large moraine at the glacial lake of Lower Grindelwald glacier that partly failed on 22 May 2009. The dashed line indicates the failed mass. The volume of the landslide was about 300,000 m^3, with 100,000 m^3 reaching the lake and generating an impact wave which, however, did not cause any dam overtopping or flood downstream. The volume of the lake reached >2.5 $\times$ 10^6 m^3 water in 2009 (image: gletschersee.ch). (b) Rockslide from the Eiger resulting from the debuttressing after the retreat of the Lower Grindelwald glacier.

retreat since its maximum glacier extension during the Little Ice Age around 1860, with an accelerated retreat since the 1980s (Zumbühl et al., 2008). The glacier terminus is currently located at the upper end of a gorge in a glacially overdeepened trough which is constrained at the downstream end by a rock slope. Downwasting of the glacier in its terminal part resulted in a loss of between 60 and more than 80 m of ice thickness between 1985 and 2000 (Paul and Haeberli, 2008). In recent years a glacial lake started to form in the terminus area of the glacier (Figure 3a). In 2004 and 2005, the lake had a limited volume but has subsequently continuously grown in the spring and early summer seasons, resulting in lake volumes of 250,000 m^3 in 2006, 1.3 million m^3 in 2008 and 2.5 million m^3 in May 2009 and the occurrence of a glacier-lake outburst flood (GLOF) in 2008 (Werder et al., 2010).

III Glacier debuttressing and the occurrence of rock slope instability

The concentration of pronounced effects of glacier downwasting and debuttressing on rock and moraine slopes, permafrost degradation, rockfalls and debris-flow activity, all interacting with the formation and growth of glacier lakes and further glacier decay, is often remarkable. At Lower Grindelwald glacier, the rock slope failure above the glacier terminus (Figure 3b) is a textbook example of the effects of glacier retreat, downwasting and associated debuttressing effects on rock slope stability, and could in fact serve as a model case for increasingly destabilized future high-mountain environments. The response of a rock slope to glacier downwasting has been reported to result in (1) large rock avalanches, (2) large-scale, progressive and slow rock mass deformation, and (3) frequent rockfall events (Ballantyne, 2002). The three modes of response are all consequences of stress redistribution and release and may act in a combined way. Rock slope failure thereby often represents the result of slope steepening by glacial erosion and unloading or debuttressing due to glacier retreat (Augustinus, 1995; Holm et al., 2004).

Examples that may be related to glacial oversteepening or debuttressing include the Brenva and Triolet rock avalanches in the Mont Blanc massif in the 18th and 20th centuries (Deline, 2009), the Sherman glacier-rock avalanche in 1964 in Alaska (although earthquake-triggered; Shreve, 1966), a significant number of rock avalanches in the Karakorum (Hewitt, 1988, 2006, 2009) and several rock avalanches in British Columbia, Canada (Geertsema et al., 2006), to name just a few. The nature, timing and scaling of rock slope failures due to glacial debuttressing is strongly conditioned by geology, in particular by lithology and structure, i.e. rock mass strength, orientation and inclination of discontinuities, and density and depth of joint networks. The timescale of failure and its delayed reponse in relation to glacier retreat has been much debated (Ballantyne, 2002). Abele (1994), for instance, noted that almost all large rockslides in the European Alps have been favoured by glacial oversteepening and subsequent debuttressing, with some failures during the Lateglacial but others occurring much later during the Holocene (e.g. the 1991 Randa rockslide, Swiss Alps; Figure 4).

Recent advances in geochronology have helped to better constrain the ages of many rock slope failures in alpine environments (Ivy-Ochs et al., 2009; Prager et al., 2008). The picture that evolves from the different studies and approaches is one of a varying response of rock slope failures to glacial retreat and downwasting. Failures can occur on timescales of 10^1 to 10^4 years, depending on the glaciation history, topography or geology. Cruden and Hu (1993) proposed an exhaustion model of temporal distribution of rock slope failures which in essence suggests that the number of failures exponentially decreases following deglaciation. Although the implicit timescale is of the order of 10^3 rather than 10^1

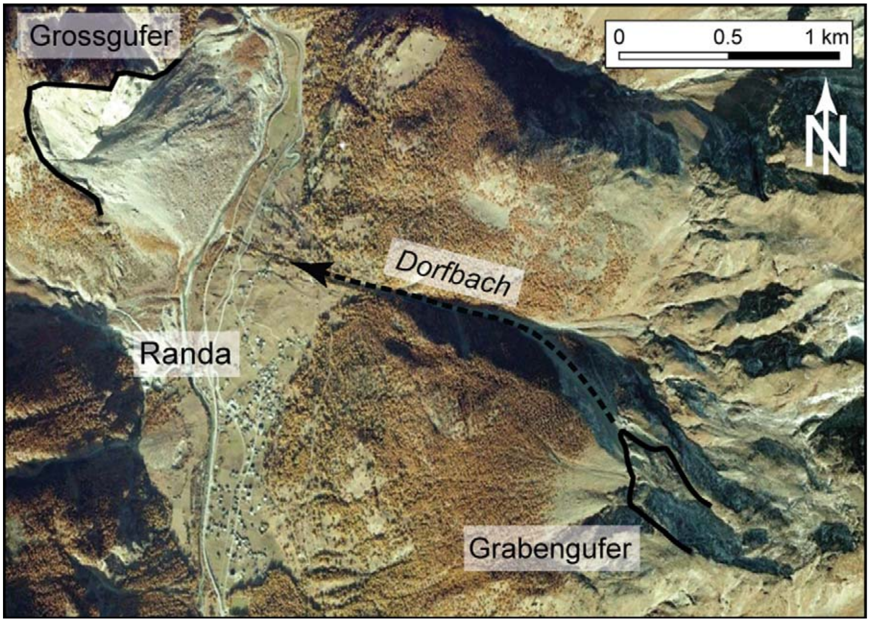

Figure 4. Image of the Grabengufer rock glacier and Dorfbach debris flow areas and the Grossgufer rock slide at Randa. Solid lines indicate the rock glacier and rock slide scar, the dashed line the trajectory of the Dorfbach debris flows, fuelled by the debris from the instabilities of the rapidly moving Grabengufer rock glacier *Source*: Base image from Google Earth, October 2009

years, the model is of importance in view of hazards related to such rock slope failures. It can be suggested that the current rapid glacier downwasting is likely to promote many rock slope failures at rather short timescales, i.e. probably on the order of decades. Such a timescale was also observed during the recent rock slope failure at Lower Grindelwald glacier where the failure occurred as a response of glacier downwasting in the past several decades. Although most such failures may not reach inhabited areas, they are of importance because of tourism and mountaineering activities (especially in the Alps), and especially in consideration of impacts on existing or newly forming natural or artificial lakes. The Lower Grindelwald glacier is just one example, but similar processes and potential hazards have been documented in other mountain ranges as well (e.g. New Zealand; Kirkbride and Warren, 1999).

IV Temperature increase and permafrost degradation

Important effects of climate change on mountain slope stability are furthermore related to warming and thawing of permafrost. Permafrost exists in many steep rock slopes in high-mountain environments and its degradation due to global warming can affect slope stability. Although this link might be intuitively clear, the mechanisms of permafrost degradation and related slope stability are rather complex, and the corresponding research field is relatively young (Gruber and Haeberli, 2007). As a result, many aspects and links remain uncertain to date because of the complexity of interacting processes.

Evidence comes from a number of recent slope failures in permafrost areas (Figure 4), including mass movements at scales that range over several orders of magnitude from block

Figure 5. Block and rockfall from the terminus of the Grabengufer rock glacier into the debris flow initiation zone of the Dorfbach torrent (see also Figure 4) *Source*: Photo by Florian Frank, August 2010

and rockfall (Figure 5) to rock avalanches (volumes of $\sim 10^2$ to 10^7 m^3), observed predominantly in the Alps but also in other mountain regions. Several studies have demonstrated that the heat wave in summer 2003 and the related excessive thawing of the active layer of permafrost bodies have resulted in an unusually high number of rockfalls at high-elevation sites in the European Alps (e.g. Gruber et al., 2004). In their reconstruction of rockfall activity since AD 1600, Stoffel et al. (2005) showed that (1) the temperature increase of the past ~ 30 years has resulted in increased rockfall activity, and (2) the warm summers around AD 1720 created conditions favourable for the release of large rockfalls, comparable to those of 2003, at a case-study site in the Valais Alps, possibly as a result of unusually large active layer thawing. Fischer et al. (2011) observed an increase of large rockslide failures in the Swiss Alps and neighbouring areas for the past two decades as compared to the 20th century. Ravanel and Deline (2011) corroborate these findings in a more detailed, local study in the Mont Blanc area.

Examples of large rock failures in the Alps include the 2004 2.5×10^6 m^3 rock avalanche from Punta Thurwieser, Italy (Sosio et al., 2008); the 1997 2–3×10^6 m^3 Brenva rock avalanche, Mont Blanc region, Italy (Barla et al., 2000; Deline, 2009), the 2006 $\sim 10^6$ m^3 rockslides from Dents du Midi and Dent Blanche, Switzerland, the 2007 rock avalanche from Monte Rosa east face, Italy (Fischer and Huggel, 2008), and the December 2011 2–3×10^6 m^3 rock avalanche at Piz Cengalo, Val Bregaglia, in the southern Swiss Alps. Many other regions have also experienced major rock failures. In the Chugach Mountains, Alaska, a $\sim 50 \times 10^6$ m^3 large rock and ice avalanche was released from Mt Steller in 2005. Some 10–20×10^6 m^3 of rock and ice released from the NNE face of Dzhimaraih-Khokh (Caucasus) entrained large parts of Kolka glacier in 2002, resulting in a devastating

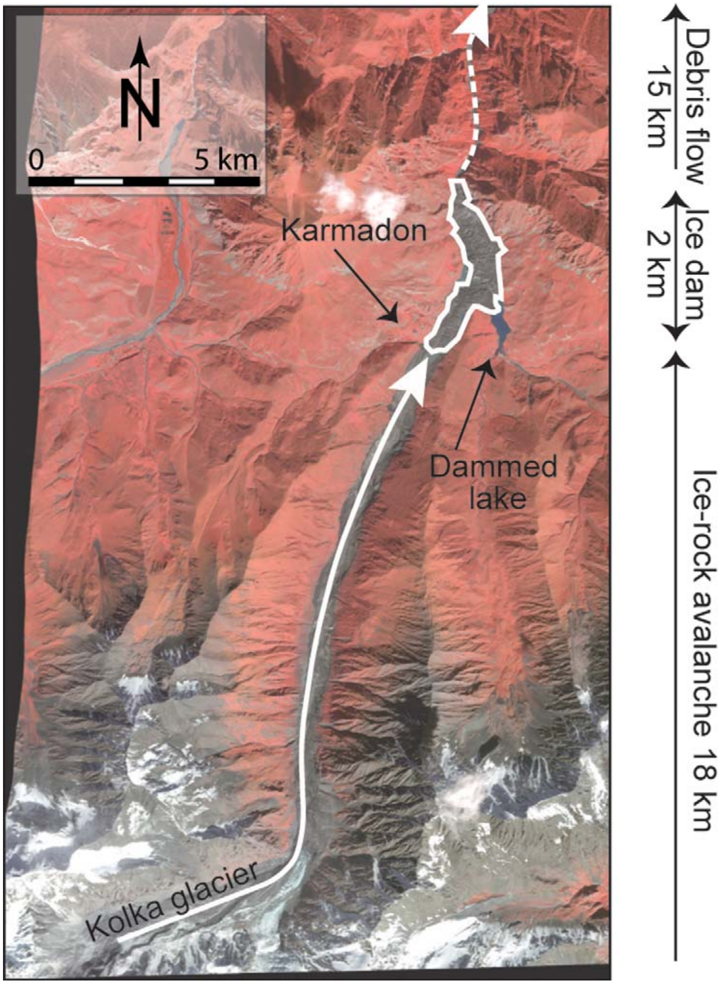

Figure 6. Reconstruction of the trajectory of the 2002 Kolka ice-rock avalanche in the Caucasus (Russia) overlain on a QuickBird false colour infrared image acquired on 25 September 2002, five days after the avalanche. A massive slope failure in glacier ice and bedrock in the northeast face of Dzhimarai-Khokh at about 4300 m asl impacted Kolka glacier. A large portion of Kolka glacier was then destabilized to form a high-speed avalanche that travelled at maximum speeds of >300 km/h down the Genaldon valley. The avalanche was dammed at the entrance of a gorge and formed a massive dam of about 130×10^6 m^3 ice and rock debris. Liquid parts of the avalanche travelled further downstream for about 15 km, devastated the valley and caused a total of about 120 fatalities.

$>100 \times 10^6$ m^3 ice-rock avalanche (Evans et al., 2009; Haeberli et al., 2004; Huggel et al., 2005; Kotlyakov et al., 2004; Figure 6).

Notably, climate change affects permafrost in rock slopes on different spatial and temporal scales. Knowledge of the temperature distribution and dynamics at depth, and related 3D effects are in fact important in order to improve our understanding on how climate change affects slope stability. Noetzli et al. (2007) modelled temperature and distribution of permafrost in idealized 3D topography and demonstrated that contemporary permafrost temperatures at depth are significantly influenced by the climate of the past millennia, including the last Ice Age. In the perspective of such timescales of heat diffusion, the 20th-century warming may only have reached a depth of some tens of metres on steep slopes at high elevations (Haeberli et al., 1997). Due to the large timelag of heat diffusion, permafrost at greater depth may be present where surface temperatures no longer favour its occurrence (e.g. Noetzli et al., 2007; Wegmann et al., 1998). As a consequence, potential effects on slope stability by recent warming may currently have penetrated to depths of tens of metres but will continue to reach increasingly greater depths with future warming. An exception may be convective and advective heat diffusion processes which can penetrate much faster and may be particularly favoured by discontinuity systems in bedrock (Gruber and Haeberli, 2007). In an attempt to assess climate change impacts on bedrock permafrost for the next few decades, Salzmann et al. (2007) coupled downscaled future climate projections from RCMs with the thermal model developed by Noetzli et al. (2007).

Although 3D thermal modelling of climate change effects on mountain permafrost is an important step forward, the processes through which air temperatures influence slope stability are not understood in sufficient detail. Davies et al. (2001) demonstrated with laboratory tests that thawing permafrost is accompanied by a reduction of shear strength in ice-filled clefts. The resulting variations and the increase of hydrostatic pressure in previously ice-filled fractures might therefore result in reduced slope stability. Furthermore, slow growth of segregation ice over long periods of time during permafrost aggradation has been suggested to widen joints in their frozen state (Murton et al., 2001; Sass, 2005), and thus contribute to a reduction in stability when permafrost thaws under conditions of warming (Harris et al., 2009).

According to Gruber and Haeberli (2007), permafrost degradation in steep bedrock would be effective through heat conduction, and by advection of heat through water percolating in fractures. As shown by Noetzli and Gruber (2009) and Noetzli et al. (2007), degradation by conduction is enhanced in convex topography (i.e. ridges, spurs and peaks) due to warming from several sides. Several of the aforementioned large rock slope failures occurred from such ridge and spur situations. However, this understanding based on theoretical considerations and modelling assumes idealized homogeneous rock conditions without discontinuity systems. In nature, rock joints are likely to have a major influence on the rate and extension of permafrost degradation and slope stability. Recent measurements in rock slopes of the Matterhorn (Switzerland), for instance, revealed the importance of melt water penetration into cleft systems as a driver of rock deformation (Hasler, 2011).

V Changing sediment reservoirs and sediment supply

In addition to future changes in temperature and rainfall intensity, changes in sediment supply and land use are important determinants for mass-movement frequency and magnitude. Recent observations at several sites in the Swiss Alps indicate that sediment supply can in fact change significantly as a result of permafrost degradation of rock and scree slopes or mass movements related to other processes (Huggel et al., 2012).

While the average flow speed of rock glaciers in the Valais Alps (Switzerland) was usually below 1 m yr^{-1}, ground and remote sensing based monitoring has revealed acceleration of rock glaciers surface flow-speed up to 4 m yr^{-1}, and occasionally up to 15 m yr^{-1}, in recent years (Delaloye et al., 2008; Roer et al., 2008; Figure 4). This phenomenon has been reported over wide regions of the Alps, with morphological features similar to those observed with landslides, such as transversal cracks, surface subsidence in the upper part and rapid advance of the frontal part (Kääb et al., 2007; Roer et al., 2008). Further field studies thereby indicate that rock glacier surface speed is increasing with increasing mean annual air temperature (Kääb et al., 2007) and, as such, warming can exert indirect control on debris-flow magnitude and frequency (Stoffel et al., 2011).

At Ritigraben (Valais Alps), permafrost thawing and accelerating flow speeds of rock glaciers have been demonstrated to deliver more sediment into the debris-flow channels under current conditions than in the past (Lugon and Stoffel, 2010). As a consequence, the volume of the largest debris flows has risen by one order of magnitude since the 1920s (Stoffel, 2010) and is likely to further increase with ongoing permafrost degradation (Stoffel and Beniston, 2006). The frequency of debris-flow events was not, in contrast, directly affected by these changes, as their release depends on meteorological triggers rather than sediment availability. Triggering meteorological conditions have been shown to occur less frequently under current climatic conditions as compared to those of the late 19th and early 20th centuries (Stoffel et al., 2011).

Debris fans are a widespread landscape element across the Alps formed by repeated debris flows that were often up to an order-of-magnitude larger than those observed today. Present-day debris flows typically deposit material on the fan rather than eroding any significant amount of sediment of the fan. However, recent debris-flow events have been observed in the Swiss Alps that developed sufficient erosive power to remobilize large amounts of sediment on Holocene fans. A model case is the August 2005 Rotlaui debris flow in Guttannen, central Swiss Alps, which was the largest event in Switzerland for the past 20 years. From a total debris-flow volume of 500,000 m^3, $\sim$300,000 m^3 was entrained on the fan, increasing debris-flow magnitude by a factor >2 due to erosion on the fan (Figure 7). The Rotlaui debris flow initiated in extensive sediment that was uncovered by the recession of the local glacier during the past decades. Permafrost contributed to hydrological conditions favouring rapid surface runoff. As observed with previous debris flows, glacier retreat since the Little Ice Age and associated uncovering of large amounts of sediment was an important factor for particularly large debris flows in the Alps. It can therefore be hypothesized that poorly consolidated, exposed sediment (1) will become increasingly available with future glacier recession (and possibly from permafrost degradation as well), and (2) could facilitate the generation of large debris flows with little or no historical precedence. As shown by the 2005 Rotlaui event, such large debris flows could exceed a critical threshold of shear stress when entering Holocene fans, and transform deposition-dominated into erosion-dominated flows.

VI Climate projections, mass-movement perspectives

It has become clear that slope stability and mass movements in high-mountain regions are complex and highly interlinked systems. Sediment supply, as described above, can also be controlled by climatic changes through processes such as the increase of the active layer of permafrost, increase of rock glacier flow speed, or aggradation of rock debris from destabilized rock flanks. At the same time, with the ongoing and eventually complete downwasting of glaciers, certain sources of sediment may become depleted or disconnected from downslope systems; an increase of vegetation cover at higher altitudes could reduce erosion rates and create more stable conditions in the long term. In this case, conditions might improve and imminent hazards could decrease at certain locations on a multi-decadal timescale.

Over decadal timescales, topography can be affected significantly as a result of shrinking glaciers and slope failures (Geertsema et al., 2006; Holm et al., 2004; Paul and Haeberli,

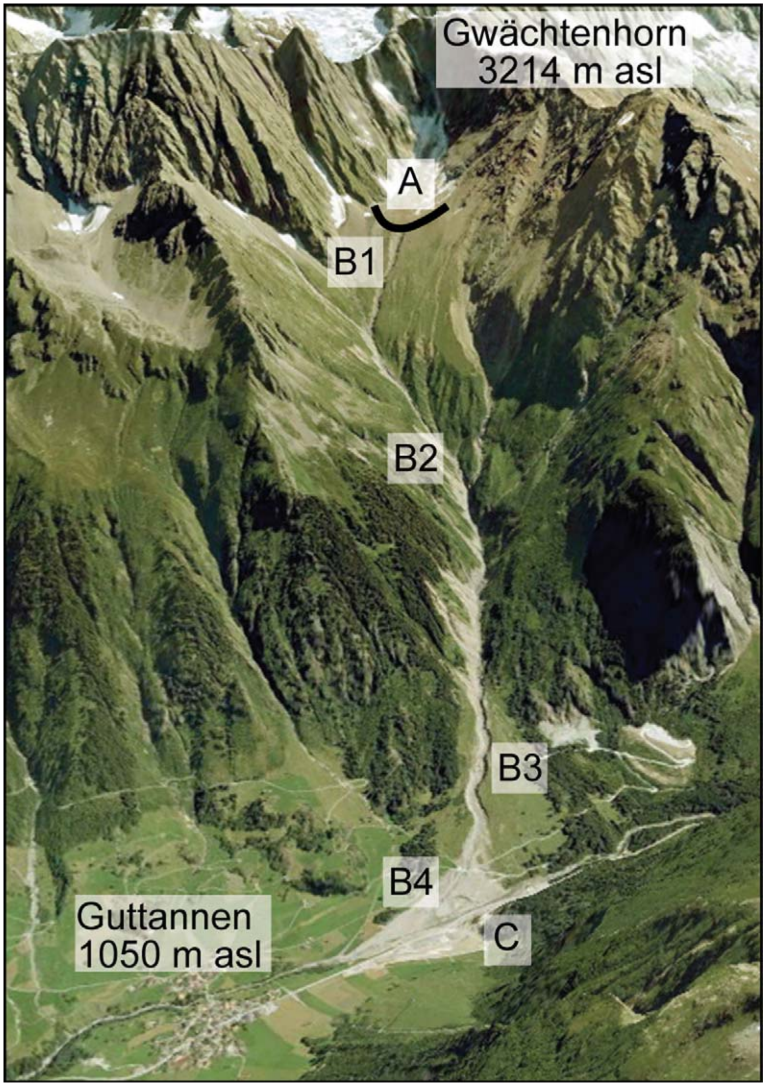

Figure 7. View of the catchment of the 2005 Rotlaui debris flow, at Guttannen, central Swiss Alps. (A) Zone of initiation consisting of an import body of partly frozen glacial sediment. (B1) Initial zone of debris flow erosion in glacial sediment. (B2) Debris flow transport zone with repeated channel erosion. (B3) Deep erosion on the Holocene fan, amounting to about 300,000 m^3 additional sediment. (B4) Deposition of debris flow material on the lower part of the fan, with runups on the opposite valley flank. (C) Destruction of the Grimsel highway and obstruction of the Aare river by debris flow material. *Source*: Base image from GoogleEarth, October 2009

2008). Temperature and precipitation are likely to change as well, but on much shorter timescales, presumably decadal. In its conclusions, the AR4 (IPCC, 2007) stated that both heavy precipitation and extreme temperature events (heat waves) will very likely increase in frequency in the course of the 21st century, although uncertainties are larger for precipitation than for temperature (Meehl et al., 2007). Post-AR4 studies confirm these statements, provide details on several regions of the world and increase the robustness of AR4 conclusions by using a larger suite of GCMs and RCMs (Hawkins and Sutton, 2010; Orlowsky and Seneviratne, 2012). In a global study, Kharin et al. (2007) conclude that the return times of extreme precipitation events will be reduced by a factor of two by the mid-21st century as compared to the late 20th century. Incorporating the effect of topography is absolutely crucial for realistic predictions of temporal and spatial changes of precipitation patterns in mountain regions, but clearly more work is needed to enhance climate models and to improve the spatial resolution of model runs using statistical and dynamical downscaling for extreme precipitation.

According to results from statistical downscaling of GCMs for North America, a strong increase of heavy precipitation events has been projected for the second half of the 21st century as compared to the second half of the 20th century over the south and central USA, but a decrease seems likely to occur over the Canadian prairies (Wang and Zhang, 2008). Precipitation extremes tend to increase also over northern and central Europe, yet with a possible decrease in southern Europe (Beniston et al., 2007; Schmidli et al., 2007). The trend of increasing precipitation extremes is also consistent with studies that focus on areas in Europe with complex topography (Kyselỳ and Beranová, 2009). Similarly, for most of southeastern South America and western Amazonia an increase of the intensity of extreme precipitation is projected for the period 2071–2100 (Marengo et al., 2009).

Uncertainties concerning seasonal, climate variability in the future are large but some tendencies and model consensus are nevertheless discernible: in northern and parts of central Europe an increase in extreme precipitation is projected in winter; whereas for summer the models give a less consistent picture of more frequent heavy rainfall (Christensen et al., 2007; Kyselỳ and Beranová, 2009; Orlowsky and Seneviratne, 2012). This is also in agreement with studies from the UK where a number of models indicate an increase of extreme precipitation in winter, spring and autumn/fall (Buonomo et al., 2007; Fowler and Ekström, 2009).

As far as future extreme temperatures are concerned, post-IPCC AR4 studies are mostly consistent with the conclusions of AR4: periods with temperatures in the uppermost percentiles are likely to be more intense, more frequent and longer lasting in a future warmer climate (Meehl et al., 2007). Daily minimum temperatures are projected to increase faster than daily maxima, which will supposedly lead to a decrease of the diurnal temperature range and could enhance weathering processes in mountain rock slopes. Considerable progress has been made in regional climate modelling for Europe since AR4 (IPCC, 2007), namely through the ENSEMBLES project (van der Linden and Mitchell, 2009) and the application of a large number of RCMs over Europe (25 or 50 km horizontal resolution; 1951–2050, some until 2100) with identical boundary conditions and the *Special Report on Emission Scenarios* (SRES) A1B (Nakicenovic and Swart, 2000). In one of the very few studies focusing on temperature extremes in high mountains, the ENSEMBLES climate model data was used to analyse the frequency of 5-day to 30-day events of very warm temperatures with clear melting conditions (above 5°C) at high-elevation sites (Huggel et al., 2010). It was found that such unusually warm events will be 1.5–4 times more frequent, in some models even up to 10 times more frequent, by 2050 as compared to 1951–2000. Rock slopes and steep glaciers in cold high-mountain regions are probably among the systems most sensitive to temperature changes, including extreme temperatures, with potential implications for slope stability, e.g. through the production of unusually large amounts of liquid melt water.

Projections of a gradual increase of (mean) temperature is typically more robust than that of temperature extremes, although the magnitude of change also bears a considerable range of uncertainty, depending on the climate models and emission scenarios that are applied. In Europe, mean temperatures are likely to increase more than on the global average, with warming likely to be strongest in winter over northern Europe and in summer over southern Europe (Christensen et al., 2007). Over the Alps, a primary region in terms of mass movements, warming may reach about $+2 \pm 1°C$ by 2050, and $+2.5–3 \pm 1.5°C$ by 2070 as compared to 1990 (OcCC, 2007). In the Andes, temperatures may rise by $3.5°C \pm 1.5°C$ by 2071–2100 with respect to the 1961–1990 reference period (Urrutia and Vuille, 2009). Similar projections were made for North America, with a warming of about 3–4°C of the annual mean temperature 2080–2099 versus 1980–1999, but with potential extreme warming over Alaska of up to 10°C in winter (Christensen et al., 2007). Changes in evaporation are also relevant for mass movement activity. A recent study analysing the complete IPCC AR4 (CMIP3) ensemble of GCM simulations for the end of the 21st century indicates generally increasing evaporation, whereas precipitation minus evaporation is found to be positive for the northern latitudes and negative for mid-latitude regions (Orlowsky and Seneviratne, 2012).

The effect of warming by several degrees Celsius on mass-movement activity is complex, involving multiple feedback processes that are difficult to predict. A limited number of studies made attempts to link different slope stability models (semi-empirical, physically based) with downscaled climate model output (Buma and Dehn, 2000; Collison et al., 2000; Schmidt and Glade, 2003). Uncertainties of climate models with respect to future rainfall intensities (Meehl et al., 2007), as well as site-specific conditions, appear to represent dominant aspects for the direction of change in mass-movement frequency. A reduction of shallow landslide activity is possible, as Bathurst et al. (2005) have found for a catchment in the southern Alps of Italy for scenarios of warmer and drier climate towards the end of the 21st century. Malet et al. (2007) assessed potential changes of slope stability for a site in the French Alps towards the end of the 21st century, using downscaled climate model output in conjunction with a hydrological and a slope stability model, and found to some extent an important reduction of slope stability, with effects from strongly reduced snow cover. Another approach was used by Jakob and Lambert (2009) in their assessment of changes in short- and long-term rainfall conditions towards the end of the 21st century from a number of climate models. The authors then applied a statistical relation of rainfall duration and intensity for landslide-triggering storms to analyse potential future changes in landslide frequency. A similar approach was used by Jomelli et al. (2009) in the Massif des Ecrins (French Alps) to analyse changes in the frequency of occurrence of debris flows by the end of the 21st century.

VII Future mass movements: implications for hazards and need for further research

The effects of changing mean and extreme temperature and precipitation are likely to be widespread and to influence both the occurrence (in terms of temporal frequency) and the magnitude of future mass movements in mountain regions around the globe. Despite uncertainties, slopes currently underlain by degrading permafrost will probably become less stable at progressively higher altitudes with ongoing climate change (Harris et al., 2001, 2009). One can also speculate that the probability of rock instability and the incidence of large ($>10^6$ m^3) rockfalls will increase in a warming climate (Holm et al., 2004; Huggel, 2009). Glacier downwasting will result in the formation of further ice-marginal lakes and subsequent problems of glacier lake outburst floods (GLOFs; Frey et al., 2010; Worni et al., 2012). On steep slopes, warming firn and ice temperatures may result in new sites of ice falls and ice avalanches (Harris et al., 2009). Provided that sediment supply is not a limiting factor, future debris flows have the potential to become larger in the future than they were in the past, but not necessarily more frequent and clearly conditioned by local site conditions (Bollschweiler and Stoffel, 2010; Jomelli et al., 2009; Stoffel et al., 2008; Schneuwly-Bollschweiler and Stoffel, 2012). The generation of cascading processes at high elevations might increase and result in chain reactions which are often difficult to predict (e.g. the impact of ice-rock avalanches into glacier lakes in the Cordillera Blanca, Peru, and other regions, with subsequent downstream flooding; Carey et al., 2012; Hubbard et al., 2005; Kershaw et al., 2005).

Recent developments at high-elevation sites have shown clearly that the sensitivity of mountain and hillslope systems to climate change is likely to be acute, and that events beyond historical experience will continue to occur as climate change continues (Goodfellow and Boelhouwers, 2012). Despite the lack of preserved analogues and although palaeo-evidence for the link between mass movements and climate may be fraught with many questions of interpretation (Lang et al., 1999; Stoffel et al., 2010), the inclusion of palaeo-records has the capacity to integrate signals over a wide area and to employ a broad range of supporting proxy information (Crozier, 2010). In this perspective, observable evidence from the far and near past points to an increase in mass-movement activity and a major mobilization of sediment under warming and/or wetter conditions (Sletten et al., 2003; Stoffel, 2010; Stoffel and Beniston, 2006).

Recent mass-movement processes originating from high-elevation sites have been observed at sites with no or little historical precedence – e.g. Rotlaui (Figure 7) and Spreitlaui debris flows, Switzerland – and have destroyed critical transport and energy infrastructure. The risk of damaging events is often particularly high in developing countries where both population and agriculture pressure on land resources lead to the exploitation of unstable slopes.

An improved understanding of the intrinsic complexity of mass-movement processes at high-elevation sites, cascades of processes and their relationships with, and dependency on, climate variability and change is crucially important in planning appropriate and prospective measures to reduce the negative impacts of future events. This progress report has tried to shed light on recent developments and state-of-the-art knowledge on possible effects of climate change on the occurrence and intensity of potential future mass movements. The contribution has also outlined the broad spectrum of recent research and the large suite of case-study results available in the research field, but also illustrated clearly that, despite all progress made in climate–mass movement studies, many questions yet need to be addressed in greater detail to overcome considerable gaps in knowledge which continue to exist in this challenging field of research.

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