EARTH SURFACE PROCESSES AND LANDFORMS
Earth Surf. Process. Landforms **37**, 77–91 (2012)

Published online 28 September 2011 in Wiley Online Library
(wileyonlinelibrary.com) DOI: 10.1002/esp.2223

State of Science

Is climate change responsible for changing landslide activity in high mountains?

Christian Huggel,[1]* John J. Clague[2] and Oliver Korup[3]
[1] Glaciology, Geomorphodynamics & Geochronology, Department of Geography, University of Zurich, Zurich, Switzerland
[2] Department of Earth Sciences, Simon Fraser University, Vancouver, BC, Canada
[3] Earth and Environmental Sciences, Potsdam University, Germany

Received 8 March 2011; Revised 29 July 2011; Accepted 6 August 2011

*Correspondence to: Christian Huggel, Glaciology, Geomorphodynamics & Geochronology, Department of Geography, University of Zurich, Zurich, Switzerland. E-mail: christian.huggel@geo.uzh.ch

ESPL
Earth Surface Processes and Landforms

ABSTRACT: Climate change, manifested by an increase in mean, minimum, and maximum temperatures and by more intense rainstorms, is becoming more evident in many regions. An important consequence of these changes may be an increase in landslides in high mountains. More research, however, is necessary to detect changes in landslide magnitude and frequency related to contemporary climate, particularly in alpine regions hosting glaciers, permafrost, and snow. These regions not only are sensitive to changes in both temperature and precipitation, but are also areas in which landslides are ubiquitous even under a stable climate. We analyze a series of catastrophic slope failures that occurred in the mountains of Europe, the Americas, and the Caucasus since the end of the 1990s. We distinguish between rock and ice avalanches, debris flows from de-glaciated areas, and landslides that involve dynamic interactions with glacial and river processes. Analysis of these events indicates several important controls on slope stability in high mountains, including: the non-linear response of firn and ice to warming; three-dimensional warming of subsurface bedrock and its relation to site geology; de-glaciation accompanied by exposure of new sediment; and combined short-term effects of precipitation and temperature. Based on several case studies, we propose that the following mechanisms can significantly alter landslide magnitude and frequency, and thus hazard, under warming conditions: (1) positive feedbacks acting on mass movement processes that after an initial climatic stimulus may evolve independently of climate change; (2) threshold behavior and tipping points in geomorphic systems; (3) storage of sediment and ice involving important lag-time effects. Copyright © 2011 John Wiley & Sons, Ltd.

KEYWORDS: climate change; landslides; glaciers; permafrost

Introduction

Many regions of the world are experiencing increases in mean, maximum, and minimum air temperatures and more frequent heavy precipitation (IPCC, 2007). Of particular concern are extreme events such as heat waves, droughts, exceptional rainfall, and floods (Füssel, 2009; Smith *et al.*, 2009), that are assessed in detail in the Intergovernmental Panel on Climate Change (IPCC) Special Report on Managing Risks from Extreme Events, published in late 2011. Landslides are another process with causal links to climate change, primarily through precipitation, but in some cases also through temperature (Figure 1A; Sidle and Ochiai, 2006; Crosta and Clague, 2009). A large body of literature is concerned with rainfall as a trigger for shallow landslides (Corominas, 2000; Wieczorek and Glade, 2005), and empirical intensity-duration thresholds have been established for rain-induced landslides in various regions throughout the world (Caine, 1980; Larsen and Simon, 1993; Guzzetti *et al.*, 2008). How contemporary climate change could affect landslide activity, however, remains largely unresolved. One key area of uncertainty is how climate change will alter the probability of damaging slope failures during a specified period (Cruden and Varnes, 1996; Lateltin *et al.*, 2005). This probabilistic measure has direct application in quantitative risk assessments and appears to be a useful metric for quantifying effects of contemporary climate change on landslide occurrence. Observations of recent large landslides in some mountain regions suggest that climate change is having an effect on slope stability (Evans and Clague, 1994; Geertsema *et al.*, 2006; Jakob and Lambert, 2009), but the evidence is typically ambiguous, and physical cause–effect relationships remain speculative or conceptual (Huggel, 2009).

The attribution of a climatic effect to a given sample of landslides requires that non-climatic and anthropogenic causes and triggers be eliminated. Inconsistency or bias in landslide reports, however, makes statistically rigorous analysis challenging. Here we focus on high-mountain areas where anthropogenic effects are absent or minimal, and where snow, glaciers, and permafrost sensitive to temperature changes may compromise slope stability (Evans and Clague, 1994; Haeberli *et al.*, 1997). Climate change may alter rates of physical and chemical weathering or degrade permafrost, all of which may

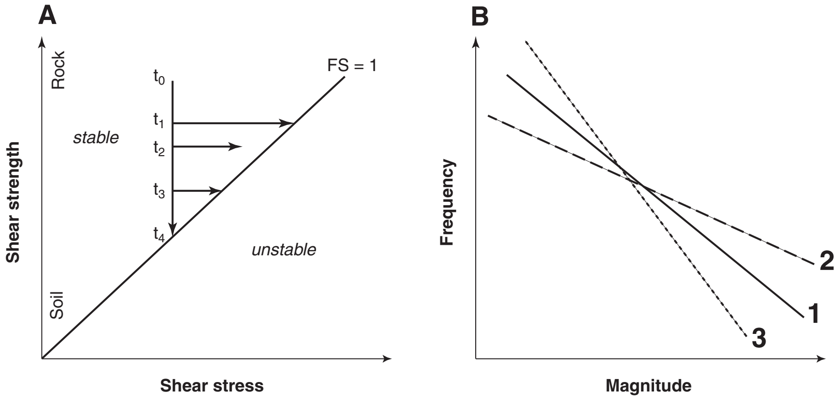

Figure 1. (A) Simplified relationship between shear strength and shear stress for a hillslope. Factor of Safety (FS) = 1 delineates a force equilibrium where the hillslope is at the threshold of failure. Vertical arrow is a time trajectory starting at t_0 and ending at t_4, during which shear strength gradually decreases to the point of failure through, for example, weathering, static fatigue, or permafrost degradation. Horizontal arrows mark dynamic loading events (i.e. potential triggers) such as earthquakes or rainstorms. Event at t_1 produces sufficient additional shear stress to trigger slope failure, whereas the event at t_2 does not. However, a subsequent event comparable to that at t_2 may trigger failure in a sufficiently weakened hillslope (t_3). (B) Schematic diagram highlighting inverse scaling relationship between landslide frequency and magnitude, such as area or volume (both axes are log-scaled). For simplicity, a simple power-law scaling is assumed. Line 1 is a reference state prior to climate change; line 2 is a possible scenario related to climate change, where larger landslides begin to dominate at the expense of smaller ones; line 3 is a scenario where smaller landslides begin to dominate at the expense of larger ones.

change the bulk strength properties of slope-forming materials (Figure 1A; Davies *et al*., 2001; Arenson and Springman, 2005; Harris *et al*., 2009). A change in the frequency or magnitude of precipitation affects infiltration rates and pore-water pressures, as well as the erosional efficacy of fluvial processes. An increase in rainfall may thus affect hillslope stability through dynamic loads during high-intensity rainstorms, slope undercutting, or redistributions of topographic-induced stresses in rock slopes (debuttressing effects) (Augustinus, 1995; Ballantyne, 2002). Clearly, the timescale on which climate change affects slope stability and landslide hazards is important (Figure 2). For instance, debuttressing effects can have lag times of the order of millennia (post-glacial), but can also act on timescales as short as decades (see section on 'Lower Grindelwald Glacier, Swiss Alps'). Similarly, ice unloading due to widespread glacier shrinkage is suggested to have effects on seismicity on timescales of millennia, but also over periods of decades, as recent studies in Alaska have shown (Sauber and Ruppert, 2008).

The approach we use here is a hybrid one, integrating review elements and case studies. In the second section we distinguish between large landslides in rock and ice, in the third section alpine debris flows, and in the fourth section coupled processes among landslides, glaciers, and rivers. Each section is introduced by a review and theory, followed by a limited number of case studies that highlight specific aspects of climate change and landslides.

We focus on identifying potential characteristic signals of climate change in a number of large ($>10^5\,m^3$), catastrophic

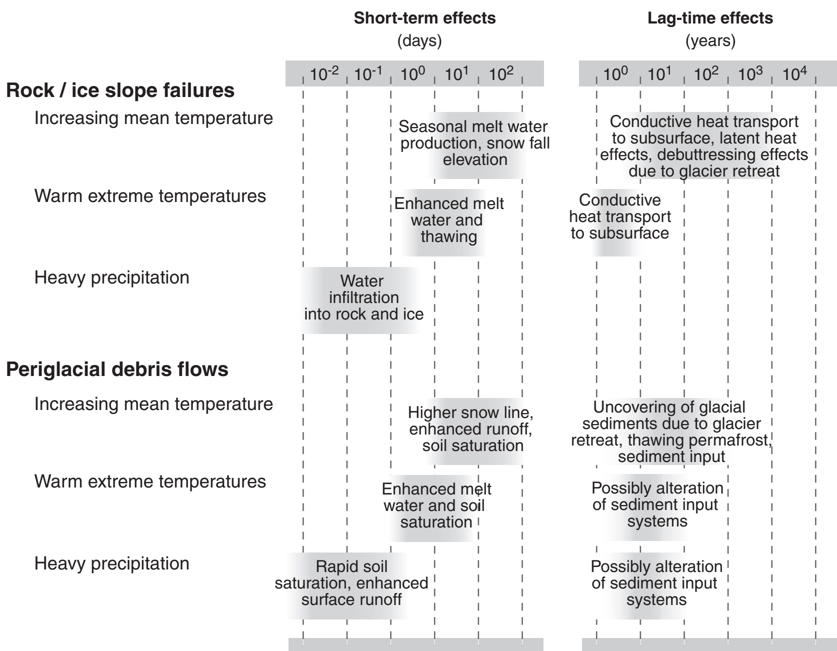

Figure 2. Temporal relationships between processes that affect slope stability and climate drivers. Short-term effects, ranging from minutes to months are distinguished from lag effects on the order of years to millennia. The timescales indicated by the grey bars are only approximate, but highlight the variability in the different processes. The spatial scale, although not shown here for the sake of simplicity, is closely related to the temporal scale. For example, the effect of colder late Pleistocene climate may still exist at depths of several hundred metres in rock (Noetzli and Gruber, 2009).

landslides that have occurred over the past 15 years in glacierized or formerly glaciated areas in the European Alps, the Americas, and the Caucasus. Table I provides a summary of the described case studies plus a few additional events. Our sample includes rock slides, rock avalanches, debris flows, and ice–rock avalanches (*sensu* Cruden and Varnes, 1996) that were not triggered by earthquakes. Although relatively rare during the historic period, such landslides have caused much destruction and loss of live (Eisbacher and Clague, 1984; Haeberli *et al.*, 1989; Huggel *et al.*, 2005; Pralong and Funk, 2006). Large and catastrophic bedrock landslides also produce more persistent geomorphic evidence than smaller slope failures and thus offer

Table I. Summary of selected recent slope failure events and their relation to climate and climate change

Location	Date of occurrence	Approximate mass failure volume (10^6 m^3)	Maximum failure elevation (m a.s.l.)	Relation to climate or climate change
Large rock slope failures				
Monte Rosa, Italy	21 April 2007	0·3	4000	Massive ice loss and bedrock failures in adjacent areas during previous years and decades; permafrost occurrence; very warm temperatures days-weeks before failure
Kolka, Caucasus, Russia	22 September 2002	130	4300	Probably warming permafrost and warming steep glacier ice
Mount Steele, Yukon	24 July 2007	27–80	4640	Permafrost occurrence; warm, melting temperatures days-weeks before failure
Mount Steller, Alaska	14 September 2005	50	3100	Probably warming permafrost and steep glacier ice; warm, melting temperatures days-weeks before failure, ridge situation
Mount Miller, Alaska	6 August 2008	22	2200	Probably warming permafrost and steep glacier ice; warm, melting temperatures days-weeks before failure
Mount Cook, New Zealand	14 December 1991	60	3755	Probably warming permafrost and steep glacier ice; very warm temperatures days-weeks before failure, ridge situation
Thurwieser, Italy	18 September 2004	2·5	3725	Probably warming permafrost at several meters bedrock depth, ridge situation
Mount Munday, BC	May–July 2007	3	3100	Probably warming permafrost; warm temperatures days or weeks before failure
Alpine debris flows				
Rotlaui, Switzerland	22 August 2005	0·5	2400	Abundant glacial sediment in source zone due to recent glacier retreat; permafrost occurrence in sediment body; extremely high rainfall intensity
Salcantay, Peru	27 February 1998	25	4200	Abundant glacial sediment in source zone due to glacier retreat; high accumulated rainfall amounts, warm temperatures implying high snowfall line and further liquid water from ice and snowmelt
Tambo de Viso, Peru	16 January 1998	0·4	2400–5600	Abundant unconsolidated sediment; intensive rainfall during the 1997/1998 El Niño event
Gerkhozhansu River / Tyrnyauz	18 July 2000	10	3500	Abundant glacial sediment in source zone due to recent glacier retreat; likely permafrost occurrence in sediment body and remnants of dead ice
Process coupling				
Ritzlihorn, Switzerland	2009–2010	>0·3 (total)	3260	Warming permafrost in rock fall source zone, days of high temperatures precede failures and debris flows
Lower Grindelwald glacier, Switzerland	2006–2010		3000	Long-term glacier retreat leads to glacier lake formation, debuttressing and moraine and bedrock failures, probably warming permafrost in rock fall source zone
Mount Meager, BC	6 August 2010	48	2400	Long-term glacier retreat leads to debuttressing effects; probably warming permafrost in rock slope failure source zone
Mount Harold Price, BC	22–24 June 2002	1	1720	Warming of steep rock slope; rockslide transformed into a debris flow

the possibility of recording past climate change (Dapples *et al.*, 2002; Korup and Clague, 2009; Borgatti and Soldati, 2010).

Catastrophic Landslides Involving Bedrock and Ice

A recent spate of large slope failures in rock and ice, resulting in rock and ice avalanches with volumes of the order of 10^5 to $10^8\,m^3$ have raised questions about the extent to which they are related to climate change (Gruber and Haeberli, 2007; Huggel *et al.*, 2008b). The best documented events are in the European Alps and include, for example, the $2{\cdot}5 \times 10^6\,m^3$ rock avalanche from Punta Thurwieser, Italy, in 2004 (Sosio *et al.*, 2008); the $2–3 \times 10^6\,m^3$ Brenva rock avalanche, Mont Blanc, Italy, in 1997 (Barla *et al.*, 2000); the $\sim 10^6\,m^3$ rock slides from Dents du Midi and Dents Blanches, Switzerland, in 2006; and the rock avalanche from the east face of Monte Rosa, Italy, in 2007. Other high-mountain regions with large numbers of recent large rock and ice avalanches include the Southern Alps of New Zealand (Allen *et al.*, 2009), British Columbia, Canada (Evans and Clague, 1988; Geertsema *et al.*, 2006), Yukon and Alaska (Lipovsky *et al.*, 2008; Huggel *et al.*, 2010), and the Caucasus (Haeberli *et al.*, 2004). Most of these rock slope failures have sources in areas of warm permafrost.

In spite of such evidence, the role of permafrost thaw in rock-slope failure is difficult to demonstrate conclusively. Research in this field is formative, but important progress has been made during the past several years. It is now understood, for example, that heat conduction and advection of heat by water moving in fractures are important processes of permafrost degradation (Gruber and Haeberli, 2007). Ridges, spurs, and peaks experience more rapid permafrost degradation than flat and gentle surfaces, because atmospheric heat can enter the subsurface from several sides (Noetzli and Gruber, 2009). In this context, we observe that several recent large rock slope failures had sources in such ridge and spur locations. Thawing effects due to advection by running water are not yet sufficiently understood, but are subject to ongoing research. Recent findings suggest that this process can produce thaw corridors along fractures on annual timescales (Hasler *et al.*, 2011).

Moreover, warming or thawing of ice-filled clefts can reduce shear strength of the materials underlying the slope (Davies *et al.*, 2001; Günzel, 2008). Increases of hydrostatic pressure in previously ice-filled fractures can result from infiltration of water. Yet it remains difficult to attribute individual slope failures conclusively to permafrost degradation, given the impediments to measuring subsurface properties and thus distinguishing the effects of permafrost from other factors that control slope stability (Fischer and Huggel, 2008; Fischer *et al.*, 2010). It is possible, however, to determine the location and extent of permafrost on rock slopes and to identify where thaw is most likely to occur. Measurements and models of subsurface temperature fields have greatly helped shape our understanding of impacts of climate change on alpine permafrost and related slope stability (Wegmann *et al.*, 1998; Gruber *et al.*, 2004a; Noetzli *et al.*, 2007; Hasler, 2011).

Climate change affects permafrost in rock slopes on different spatial and temporal scales. Due to the large time lag of heat diffusion, permafrost may persist at depth, even where surface temperatures are no longer favorable (Wegmann *et al.*, 1998; Noetzli *et al.*, 2007). The recent warming signal has penetrated into bedrock to depths of decametres, and will continue to reach further depth in coming decades.

The concept of the para-glacial cycle (Ballantyne, 2002) has been applied to hillslope stability in formerly glaciated regions. Augustinus (1995), for example, argues that slopes gradually adjust to altered stress regimes following glacier downwasting and retreat. The idea that glacial debuttressing decreases the stability of slopes has been influential and has led many authors to conclude that some large bedrock landslides are a lagged response to major climate change (Abele, 1974; Cossart *et al.*, 2008). Proposed lag times following de-glaciation are of the order of millennia (Ivy-Ochs *et al.*, 2009; Prager *et al.*, 2009), but may also be as short as decades (see section on 'Lower Grindelwald Glacier, Swiss Alps').

Here we highlight two case studies that further support the notion of a causal relationship between glacier retreat, permafrost degradation, and slope failure. A third example demonstrates potential ambiguities of attributing large failures in ice and bedrock to climate change.

Monte Rosa, Italy

The Monte Rosa east face, Italian Alps, extends from 2200 m to > 4500 m above sea level (a.s.l.) and is one of the largest rock walls in the Alps. It is formed of gneissic rocks and is partly covered by steep glaciers that have thinned and retreated considerably over the past 30 years (Figure 3; Haeberli *et al.*, 2002; Fischer *et al.*, 2006). Frequent slope instability involving both ice and rock has been documented since about 1990 (Fischer *et al.*, 2006). The largest slope failure was a $1{\cdot}1 \times 10^6\,m^3$ avalanche of ice and snow that detached from the middle section of the east face at ~3600–3800 m a.s.l. in August 2005. The avalanche reached the foot of the rock face, where a large supra-glacial lake had formed in 2002 but had drained in 2003. Had the lake still existed in 2005, the avalanche would have generated a displacement wave with likely catastrophic consequences for the downstream community of Macugnaga. The avalanche, which occurred at night, buried much of a pasture near an alpine hut frequented by tourists during daytime. Less than two years later, in April 2007, $\sim 0.3 \times 10^6\,m^3$ of rock detached from the east face at ~4000 m a.s.l. and fell to its base at ~2200 m a.s.l. The rock mass detached from a dip slope on which a large amount of ice had been lost in recent years, likely altering the local temperature and stress fields (Fischer *et al.*, 2011). High-resolution photogrammetry and LiDAR-based topographic studies revealed that smaller, more frequent rock and ice avalanches during the last two decades of the twentieth century were responsible for $\sim 20 \times 10^6\,m^3$ of mass loss from the east face of Monte Rosa (Fischer *et al.*, 2011). This loss predominantly occurred from about 3300 to 4100 m a.s.l., corresponding to areas of warm to cold permafrost.

Meteorological observations at high elevations on Monte Rosa and other sites in the Alps have shown that atmospheric warming can cause strong, non-linear increases in firn and ice temperature. Melting initiates at temperatures of greater then $-10\,°C$, and subsequent refreezing can strongly affect warm firn and ice due to latent heat production (Suter and Hoelzle, 2002; Hoelzle *et al.*, 2010). In this manner cold ice can transform into polythermal or temperate ice within several years (Haeberli and Alean, 1985). This effect, together with enhanced infiltration of melt water to the base of steep glaciers, could change the potential source zones of ice avalanches.

The spatio-temporal analysis of landslide activity at Monte Rosa indicates that the primary onset of mass loss at the end of the 1980s and early 1990s occurred in tandem with a strong increase of local air temperatures (Fischer *et al.*, In press). The large mass losses imply changes in mechanical stress and thermal surface and subsurface fields that influenced subsequent landslide activity, including the 2007 rock avalanche. The lesson learnt from Monte Rosa thus is that slope instability initially may be triggered by climate change, but then can develop further independently of additional climatic stimuli.

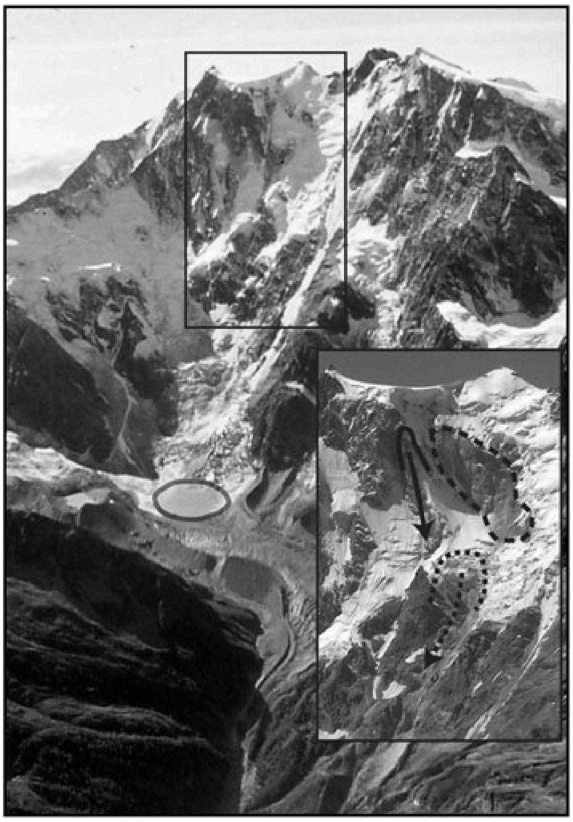

Figure 3. Significant changes in the Monte Rosa east face between 1985 (background) and 2004 (photo insets). Areas demarcated by the dotted and full line arrows indicate, respectively, the failure zones of the 2005 and 2007 avalanches. The area enclosed by the dashed line is a zone that became completely de-glaciated between 1985 and 2004. The ellipse at the toe of Monte Rosa on Belvedere Glacier is the location of a supra-glacial lake that formed in 2002 and generated a non-catastrophic outburst in 2003 (as a reference for scale, the diameter of the lake is approximately 500 m). (Photographs courtesy of J. Alean and L. Fischer.)

Kolka-Karmadon, Caucasus

The 2002 ice–rock avalanche in the Kasbek region of the Russian Caucasus is emblematic of a complex long-runout landslide. Approximately $10–20 \times 10^6$ m^3 of rock and glacier ice detached from a steep slope of intensely fractured rock below the summit of Dzhimarai-khokh (4780 m a.s.l.; Haeberli *et al.*, 2004; Kotlyakov *et al.*, 2004; Huggel *et al.*, 2005; Evans *et al.*, 2009). The failure zone extended from about 3500 to 4300 m a.s.l. The rock mass fell onto Kolka Glacier and sheared off most of it. More than 100×10^6 m^3 of rock and ice avalanched down the Genaldon valley for 19 km before coming to rest above the narrow Karmadon gorge. A debris flow continued downvalley another 15 km before stopping 4 km above the town of Gisel. The avalanche and the debris flow devastated the valley and killed 120 people. The ice-rich debris formed a dam at Karmadon, impounding several lakes in tributary valleys with up to 5×10^6 m^3 of water. Potential floods from these lakes were a threat to downstream areas (Haeberli *et al.*, 2004), but the lakes emptied non-catastrophically.

The thermal state of the source area was studied by Haeberli *et al.* (2003) and Huggel (2009), who concluded that mean annual air temperatures in the detachment zone were about −5 to −10°C, suggesting the presence of cold permafrost. Ten temperature loggers were installed on rock faces around Kolka Glacier to monitor seasonal ground temperatures at 10 cm depth. Preliminary results indicate that the lower limit of permafrost on the north-northeast face of Dzhimarai-khokh, where the failure occurred, is ~3000 m a.s.l. and that the temperature range in the detachment zone is between −3 and −8°C (Huggel, 2009).

Historic climate trends for the region are poorly known, but the observed rapid retreat of glaciers, which has accelerated since the mid-1990s, attests to recent warming (Stokes *et al.*, 2006). Changes in glacier cover on the north-northeast face of Dzhimarai-khokh are not nearly as well documented as for Monte Rosa. However, the two large mountain walls are strikingly similar, and the assumption that hanging glaciers in the Caucasus have undergone comparable changes to those in the Alps is not unreasonable. A large loss of glacier ice would have changed temperature and stress regimes in the underlying bedrock. Thermal modeling has shown that ice may have a deep warming effect on bedrock through latent heat dissipation due to freezing of meltwater, reaching instability when temperatures approach the melting point (Huggel *et al.*, 2008a; Huggel, 2009). The currently still-limited understanding of the mechanics of failure and entrainment of Kolka Glacier warrant future research (Haeberli *et al.*, 2004; Kotlyakov *et al.*, 2004; Evans *et al.*, 2009).

Mount Steele, Canada

A large rock-ice avalanche occurred within a steep gully system on the precipitous north face of Mount Steele (5067 m a.s.l.), southwest Yukon Territory, Canada, on July 24, 2007 (Figure 4; Lipovsky *et al.*, 2008). Two days before this event a large (~ 3 million m^3) ice avalanche occurred on the same slope. The leading edge of the ice avalanche traveled across Steele Glacier below the gully, up a 275-m-high rock ridge, and then down onto another glacier below. The total horizontal travel distance is about 8 km. The July 24 event involved several tens of millions of cubic meters of ice and rock. The headscarp is about 540 m wide and exposes a wall of glacier ice ~70 m thick. The failed material accelerated down a 44° slope from about 4600 to 2800 m a.s.l., before running onto Steele Glacier. The debris descended up to 2160 m and traveled a maximum horizontal distance of almost 6 km, leaving a deposit about 3·7 km^2 in area on the glacier. In the distal part of its travel path, the landslide reached, but did not overtop, the ridge crested by the earlier ice avalanche. Some of the debris at the northwest edge of the deposit and much of the debris that climbed the ridge slid backwards on reverse slopes after reaching its limits of travel.

The seismic magnitude estimated from long-period surface waves (M_s) is 5·2. Modeling of the waveforms suggests an estimated duration of approximately 100 seconds and an average velocity of between 35 and 65 m s^{-1} (Lipovsky *et al.*, 2008). This landslide is one of 18 large rock avalanches known to have occurred since 1899 on slopes adjacent to glaciers in western Canada.

Based on long-term (1971–2000) temperature data from Burwash Landing, located at 807 m a.s.l. 76 km northeast of Mount Steele, and a lapse rate of 0·65°C/100 m, mean annual air temperatures at the Mount Steele detachment zone (~ 3000–4650 m a.s.l.) are estimated to range between −18 to −29°C. Such temperatures indicate cold ice, possibly with a zone of recrystallization and infiltration during summer, especially in the lower part.

The Burwash Landing temperature record furthermore indicates an increase of mean temperature of 0·65°C per decade over the past 40 years (1966–2007), amounting to as much as 2·6°C in total (Figure 5). For 10 days prior to the landslide, daily maximum temperatures were about 3°C above the 1971–2000 July monthly average maximum temperature. The average daily temperature for the month of July was ~2°C higher than

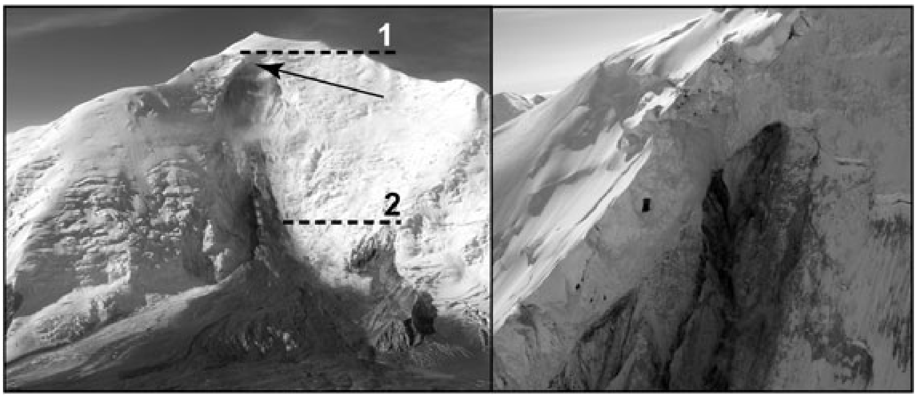

Figure 4. The 2007 ice and rock avalanche, Mount Steele, Yukon, Canada. Left: Overview of the landslide scar and landslide material deposited on Steele Glacier. Dashed lines indicate the upper and lower end of the detachment zones at 4650 m and 3000 m a.s.l., respectively. The arrow points to the location of the enlarged image at the right hand. Right: Close-up of the failure zone in bedrock and glacier ice. There is evidence of seepage surfacing at the ice–bedrock interface of the failure zone. (Photographs taken on August 2, 2007, by P. Lipovsky.)

normal, suggesting that maximum daily air temperatures were between −3°C and 7·5°C, respectively, for the upper and lower ends of the detachment zone (Figure 6). We find that, although the upper part of the source zone remained a few degrees below freezing, the lower part was above freezing for several days before failure. Photographs taken shortly after the event appear to show frozen water on the rupture surface (Figure 4), raising the possibility that sensible heat, solar radiation, or geothermal heat was sufficient to produce free water. However, the possible role of long-term climate change and short-term air temperature variations on the slope failure remains to be studied in more detail.

Debris Flows in Peri-glacial Environments

Debris flows are common processes and hazards in mountains (Chiarle *et al.*, 2007). Most debris flows are triggered by prolonged or intense rainfall, in some cases with snowmelt (Jakob and Hungr, 2005), or by catastrophic drainage of naturally dammed lakes (Clague and Evans, 2000; Korup and Tweed, 2007). Other triggers include alluvial bulking of debris along steep stream courses and erosion of debris cones on steep rock slopes (Rickenmann and Zimmermann, 1993). In recent decades, glacier retreat due to atmospheric warming has exposed large amounts of poorly consolidated sediment prone to slope failure and debris flows. Some of the largest debris flows in the Alps in recent years have had sources in newly de-glaciated areas (Zimmermann and Haeberli, 1992; Chiarle *et al.*, 2007; Scheuner *et al.*, 2009). There and elsewhere, large debris flows have occurred when glacier- and moraine-dammed lakes suddenly drained (Haeberli, 1983, 2008; Grove, 1987; Evans and Clague, 1994; Clague and Evans, 2000; Richardson and Reynolds, 2000; Korup and Tweed, 2007).

There are two basic approaches to predicting rainfall-triggered landslides. The first involves physically based models that simulate the physical processes that are relevant to landslide initiation in order to determine landslide occurrence in space and time (Montgomery and Dietrich, 1994; Iverson, 2000; Crosta and Frattini, 2003). The second approach uses empirical models that statistically relate rainfall measurements, such as intensity and duration, to documented landslide events (Wieczorek and Glade, 2005; Guzzetti *et al.*, 2007). Several landslide-triggering precipitation thresholds have been developed for mountain ranges in Europe and North America (Zimmermann *et al.*, 1997; Marchi *et al.*, 2002; Jakob and Weatherly, 2003).

There is not yet clear evidence for an increase in the frequency or magnitude of debris flows in peri-glacial areas during recent decades, which precludes an unambiguous attribution to climate change. However, extreme rainfall events, which are common

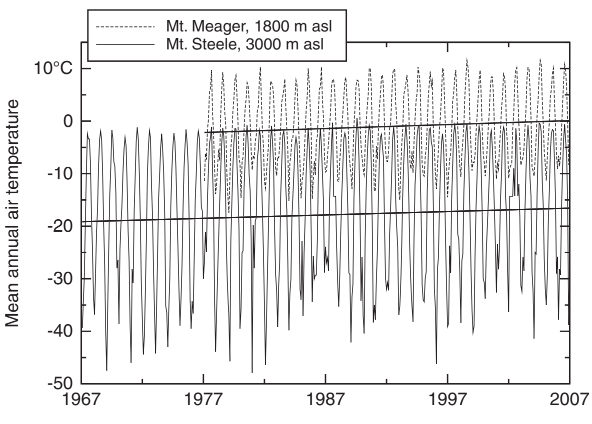

Figure 5. Mean annual air temperatures extrapolated to the lower end of the landslide detachment zones at Mount Meager and Mount Steele. Data were extrapolated from the Whistler meteorological station at 657 m a.s.l., 70 km southeast of Mount Meager, and the Burwash Landing meteorological station at 807 m a.s.l., 76 km northeast of Mount Steele, using a lapse rate of 0·65°C/100 m. Linear trends show warming of 2·25°C (1977–2007) and 2·6°C (1967–2007) for Mount Meager and Mount Steele, respectively.

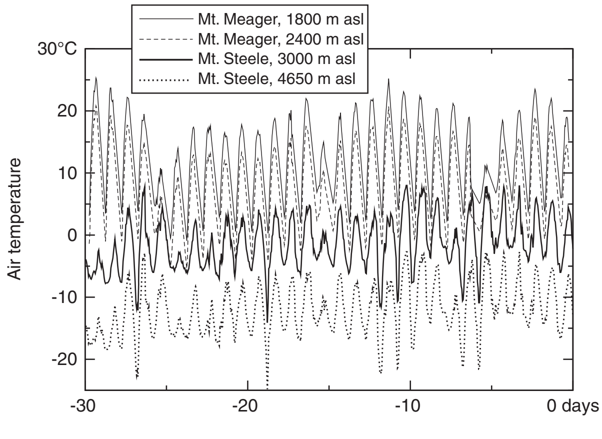

Figure 6. Extrapolated air temperature records for the 30 days prior to the Mount Meager and Mount Steele landslides, for the upper and lower ends of the detachment zones. Temperatures were extrapolated as described for Figure 5.

debris flow triggers, have increased in many regions of the world over the past several decades (Trenberth *et al.*, 2007). Although projections differ depending on the climate model used and the specific region of interest (Kyselỳ and Beranová, 2009), heavy precipitation events are generally forecast to increase in the twenty-first century (Meehl *et al.*, 2007). Beniston (2006), for example, forecasts a shift of precipitation extremes to spring and autumn in the Swiss Alps by 2100. Other models, however, show an increase in extreme precipitation events in winter (Frei *et al.*, 2006). These projections carry implications for debris flow activity, but factors other than precipitation may also affect debris flow frequency and magnitude. Sediment supply, for example, may be equally important, as case studies in the Valais, Swiss Alps, have shown (Rebetez *et al.*, 1997; Lugon and Stoffel, 2010).

Few researchers have examined the effect of future changes in precipitation on landslide activity. Jakob and Lambert (2009) forecast an increase in total monthly rainfall in coastal British Columbia by the end of the twenty-first century, with a likely corresponding increase in landslides. Other researchers have linked downscaled climate model output to slope stability models [Buma (2000) and Malet *et al.* (2007) for the French Alps; Dehn *et al.* (2000) for the Italian Alps; Collison *et al.* (2000) for southeast England; and Crozier (2010) for New Zealand]. Bathurst *et al.* (2005) used scenario runs of the Hadley Center global circulation model HadRM3 for 2070–2099 as input to a physically based model to assess changes in landslide activity at a site in the Italian Alps. They found slightly reduced debris flow activity in a warmer drier climate.

Rotlaui-Guttannen, Swiss Alps

On August 28, 2005, following a period of intensive rainfall, a large debris flow occurred in the upper Aare valley and crossed the main Grimsel highway close to the village of Guttannen. Damming of the Aare River by debris flow deposits and subsequent overflow and flooding of the village caused major damage and closed the highway. The debris flow initiated at ~2400 m a.s.l. in an area of glacial sediment that had been exposed by recession of Homad Glacier during the past several decades (Figure 7).

The Rotlaui-Guttannen event is typical of large debris flows sourced in de-glaciated areas with large reservoirs of glacial sediment (Zimmermann and Haeberli, 1992; Rickenmann and Zimmermann, 1993). With a total volume of ~ 500,000 m^3, it is the largest debris flow in the Swiss Alps over the past two decades. Permafrost that existed in its source sediments promoted rapid surface runoff (Scheuner *et al.*, 2009). The snowline was high (over 3000 m a.s.l.) during the days of heaviest rainfall, thus all precipitation in the catchment was rain, of which about 170 mm were recorded in 48 hours between August 21 and 22, 2005 (Scheuner *et al.*, 2009). One remarkable aspect of the Rotlaui-Guttannen event is that it entrained more than half of the total flow volume (~ 300 000 m^3) along its path as it crossed and eroded a Holocene debris fan (Figure 8). Debris-flow bulking of a similar magnitude has rarely been documented. Yet, large debris flows that originate in glacier forefields during heavy rainfall may entrain large quantities of sediment along their paths, even on fans or other relatively low-gradient surfaces if critical erosion thresholds are exceeded. Debris flow volumes and hazard can be greatly increased under these circumstances (Huggel *et al.*, 2011).

Salcantay, Peru

Instability induced by recent glacier retreat and exposure of unstable glacial sediment is illustrated by a large landslide on Nevado Salcantay (6264 m a.s.l.), about 20 km south of Machu Picchu, Peru, in 1998. On February 27, during the rainy season, a slope with a large volume of unconsolidated sediment failed in the Quebrada Rayancancha, on the north slope of Nevado Salcantay, forming a debris flow with a volume of $\sim 25 \times 10^6$ m^3. The landslide scar was 1 km wide and 1 km long, and was located between 3950 and ~4200 m a.s.l. (Figure 9). The debris flow traveled along the Ahobamba valley, reaching up to 60 m on the valley sides and scouring the stream bed to a depth of ~30 m (Huggel *et al.*, 2004b). Upon reaching Vilcanota River, the debris flow dammed it to a height of 70 m. A hydroelectric power plant was flooded and covered with fine alluvium, causing damages of about 100 million US dollars. The Salcantay landslide is extremely large by Alpine standards – about two orders of magnitude larger than otherwise similar events in the European Alps (Rickenmann and Zimmermann, 1993; Zimmermann *et al.*, 1997; Scheuner *et al.*, 2009).

Extrapolation of mean annual air temperatures from regional climate stations at elevations of 2500 to 3500 m a.s.l. suggests that permafrost is absent at the elevation of the source area. Persistently high temperatures during several weeks prior to the event may have contributed large amounts of melt water from snow and ice on the surrounding mountain flanks, saturating soils in the source area (Figure 10). Likely more important, however, was the saturation of sediment by persistent antecedent rainfall. We used ground- and satellite-based rainfall measurements to reconstruct the precipitation record in the remote area where the landslide initiated (Figure 10), based on the Tropical Rainfall Measurement Mission (TRMM) Multi-satellite Precipitation Analysis (TMPA), available at 0·25° grid spatial and three-hourly resolution (Huffman *et al.*, 2007). The rainfall data, analyzed and integrated over the four closest 0·25° tiles, show that several days of intense rainfall preceded the landslide. Cumulative rainfall of the 40 days prior to the event amounted to 170 mm, likely implying soil saturation. Recent studies evaluating the performance of TRMM in the same Andean region have shown that correlation between TRMM-TMPA and ground rainfall gauge stations increases over longer periods of time, with a correlation of about 0·8 for seven-day periods (Scheel *et al.*, 2010). For the Salcantay landslide, but for many other cases as well, cumulative rainfall is therefore a more reliable measure than rainfall intensity, especially when using TRMM data (Figure 10). On a more general level, the Salcantay event demonstrates the importance of both precipitation and temperature for triggering large landslides in high mountains since high temperatures and large precipitation amounts together imply runoff generation and soil saturation up to high elevations. Although the unusual amount of rainfall and high air temperatures at Salcantay reflect climate variability, the relation of this case to climate change is primarily through the effects of long-term glacier retreat that exposed large amounts of unconsolidated sediment to erosion.

Process Coupling Among Landslides, Glaciers, and Rivers

Dynamic interactions among landslides and glacial and river processes in high mountains include: (i) outburst floods from glacier- and moraine-dammed lakes that have formed in response to recent glacier thinning and retreat; (ii) slope failures in rock, ice, or moraines that impact glacial lakes, causing outburst floods; (iii) damming of rivers by landslides, resulting in potentially unstable dams that may fail within hours, days, months, or years; (iv) deposition of landslide debris in channels of rivers or torrents, altering debris flow magnitude and frequency.

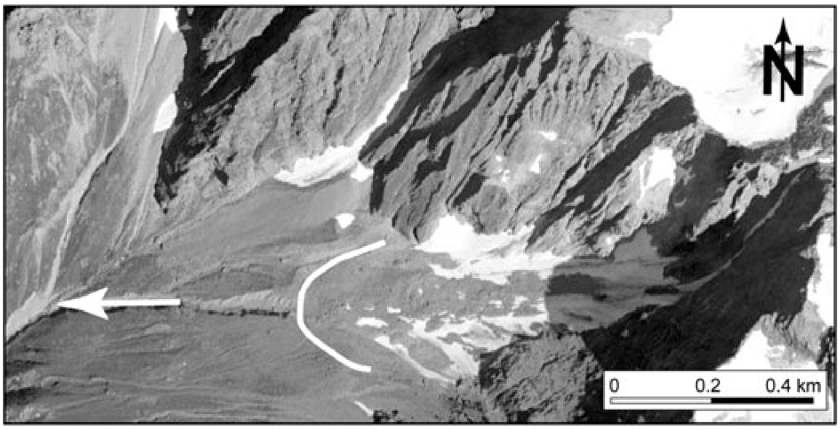

Figure 7. Area of initiation of the 2005 Rotlaui-Guttannen debris flow. The white line delineates the lower margin of a frozen body of glacial sediment. Remnants of Homaf Glacier are visible at the foot of the rock face at the right side of the photograph. The glacier formerly extended to the white line and left large amounts of sediment during its retreat. The sediment probably only became permanently frozen after retreat of the glacier. The arrow indicates the direction of debris flow propagation.

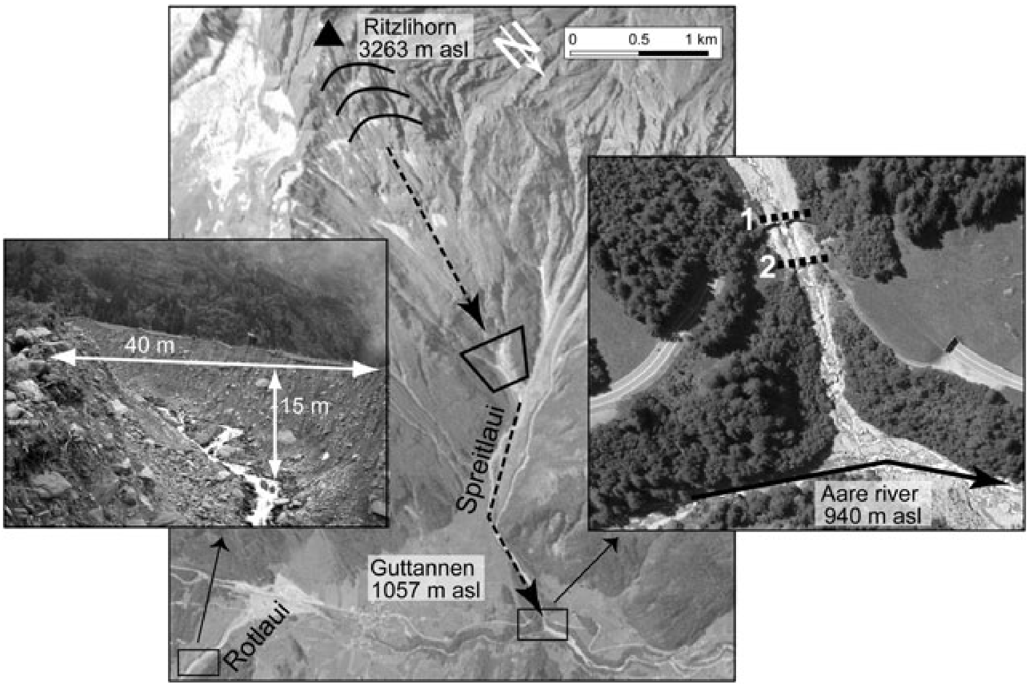

Figure 8. Cascades of landslide processes at Ritzlihorn-Guttannen, central Swiss Alps: black curved lines on the northeast flank of Ritzlihorn indicate rockfall source areas from frozen bedrock. The trapezoid marks the area where rockfall debris and avalanche snow accumulates, and debris flows initiate and propagate along the Spreitlaui channel to run out to the Aare River. The inset image on the right highlights the critical risk situation with the debris flow channel running over (1) the highway gallery, and (2) the transnational gas pipeline. Also shown on the main image is the runout zone of the 2005 Rotlaui-Guttannen debris flow, with the inset indicating a heavily eroded channel cross-section along the debris fan. (Photograph by C. Huggel and image by GoogleEarth.)

Many lakes dammed by ice and unstable moraines have formed in recent decades in high mountains around the world (Ames, 1998; Gardelle *et al.*, 2011), as glaciers retreated and downwasted at rates of up to 5–10 m yr^{-1} (Schiefer *et al.*, 2007; Paul and Haeberli, 2008). Many of these lakes have drained suddenly, producing huge downstream floods (Clague and Evans, 2000; Richardson and Reynolds, 2000). Others are bordered by slopes that show signs of instability and might fail. Landslides that enter these lakes or artificial reservoirs could generate displacement waves that overtop the dams. In April 2010, an ice–rock avalanche from the flanks of Hualcán (~5400 m a.s.l.) in the Cordillera Blanca in Peru generated a massive impact wave in a glacier lake; the wave overtopped the bedrock dam and flooded downstream communities (Carey *et al.*, In Press). Similar cases are also known from other mountain regions (Clague and Evans, 2000; Huggel *et al.*, 2004a; Kershaw *et al.*, 2005).

Landslides can interact with river processes by forming short- or long-lived dams in valleys. Water impounded behind these dams poses significant flood hazards for downstream populations and infrastructure. An extensive literature exists on processes, hazards, remedial measures, and landslide dam case studies (Costa and Schuster, 1988; Korup and Tweed, 2007). Landslides can also alter the sediment budget of rivers (Hewitt *et al.*, 2008) and steep mountain torrents. Because sediment supply is an important control on debris flow magnitude and frequency (Lugon and Stoffel, 2010), both

Figure 9. Left: Scarp of the 1998 landslide on the northern slope of Salcantay (6264 m a.s.l.) in the southern Peruvian Andes. The landslide initiated in thick glacial sediments. For scale, the black bars indicate the height of a person. Right: Aster satellite image from July 4, 2002 showing the landslide initiation area (white rectangle) relative to the glaciers of Salcantay (summit at lower right of the image). (Photograph courtesy of Reynolds Geosciences Ltd.)

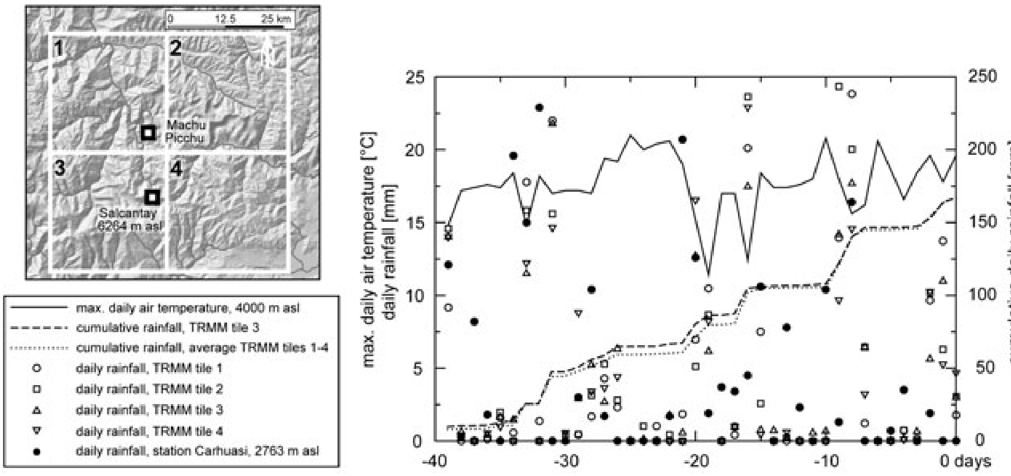

Figure 10. Rainfall and temperature record for the 40-day period prior to the Salcantay landslide. Locations of tiles for the TRMM satellite rainfall data are shown on the map at the left. The Carhuasi meteorological station is 30 km southwest of the landslide at 2763 m a.s.l. Maximum air temperatures were extrapolated to the top of the landslide failure area at 4000 m a.s.l. The cumulative rainfall record for TRMM tile 3 and an average of all four TRMM tiles are highly correlated, whereas daily rainfall amounts can differ considerably among TRMM tiles and the ground station.

slowly and rapidly moving mass movements can change debris flow hazards over many years.

The following case studies refer to the aforementioned interactions and serve as examples of what is likely to occur in the future with continued warming.

Ritzlihorn-Guttannen, Swiss Alps

Repeated rock falls began to occur in 2009 from the northeast face of Ritzlihorn (3263 m a.s.l.), ~2 km northwest of the Rotlaui-Guttannen event described earlier (Figures 7 and 8). The rock face consists of strongly shattered and weathered gneiss, and its upper part is frozen (permafrost). Many couloirs carry water, possibly derived from thawing permafrost. Further field observations provide evidence of frozen bodies of shattered rock that have become increasingly unstable and are the source of the rock falls (D. Tobler, personal communication, 2011). Blocky debris and avalanched snow accumulate at the apex of the large Holocene debris fan on which the community of Guttannen is located (Figure 8). Some of the accumulated snow persists through the summer. The debris fan is drained by the Spreitlaui torrent, which enters Aare River at 940 m a.s.l.. No damaging debris flows are known to have occurred prior to 2009 (Hählen, 2010), although the morphology and sedimentology of the fan indicate considerable debris flow activity earlier in the Holocene.

In the summer of 2009, debris flows began to occur in the Spreitlaui torrent. The initiation zone is the apex of the fan, where rock debris and snow accumulate. Saturation of the sediment by water derived from melting of avalanched snow is likely an important factor in triggering the debris flows. The largest debris flows, in July and August 2010, had volumes of ~100 000 m^3 and peak discharges of ~500 $m^3 s^{-1}$ (Hählen, 2010). The debris flows entrained large amounts of fan sediment along the Spreitaui channel and obstructed the flow of Aare River. Just before reaching the river, they passed over a gallery of the main highway and a transnational gas pipeline. Both structures were heavily damaged by debris flows in 2009 and 2010; traffic and transport of gas had to be temporarily suspended (Figure 8).

Field observations indicate that thawing permafrost has played an important role in the recent rock fall activity. The accumulated rock debris, in turn, plays a major role in the generation of debris flows, as illustrated by the historically unprecedented large events on the Holocene fan in 2009 and 2010. These events illustrate the dynamic interactions between rockfall and debris flow activity at Ritzlihorn.

More research is currently being conducted to better understand the implications of permafrost degradation on rock fall activity at this site. The Ritzlihorn events could be a model case for a major change in mass-movement frequency and magnitude related to atmospheric warming. An initial stimulus, in this case an increase in air temperature and its effect on ground surface temperatures, may have critically disturbed the system, leading to a process cascade that evolves independently of climate (see also the Monte Rosa case earlier). An additional perspective is the possibility that the Ritzlihorn system displays a threshold or tipping-point behavior, with changes in permafrost likely being at the head of a chain of subsequent landslide processes.

Lower Grindelwald Glacier, Swiss Alps

Changes in Lower Grindelwald Glacier have been well documented since at least the mid-nineteenth century, when tourism started to become popular in the Alps. Since it achieved its maximum Little Ice Age extent around 1860, the glacier has retreated ~2 km (Zumbühl *et al.*, 2008). The terminal area of the glacier downwasted 60 m to >80 m between 1985 and 2000 (Paul and Haeberli, 2008). The glacier presently terminates at the upper end of a gorge in a glacially overdeepened trough flanked by steep rock slopes.

In recent years, the steep rock slopes near the snout of the glacier have shown signs of instability that probably are related to glacier downwasting and permafrost degradation. In 2006, 2×10^6 m^3 of rock at the toe of the east flank of the Eiger, at the entrance of the glacial gorge (Figure 11), became unstable (Oppikofer *et al.*, 2008). The rock slope spectacularly collapsed later that year and still remains unstable. About the same time, a glacial lake started to form in the trough at the terminus of the glacier. Since 2005, the lake has grown, achieving volumes of ~$0{\cdot}24 \times 10^6$ m^3 in 2007, ~$1{\cdot}3 \times 10^6$ m^3 in 2008, and ~$2{\cdot}6 \times 10^6$ m^3 in 2009 (Werder *et al.*, 2010). The lake is dammed by rock debris emplaced by the 2006 and more recent landslides and by stagnant glacier ice. In late May 2008, ~$0{\cdot}8 \times 10^6$ m^3 of water suddenly drained from the lake, with a peak discharge of 110 m^3 s^{-1}. Like the face of the Eiger, the moraine at the northeast margin of the glacial lake has been destabilized by glacier downwasting in recent years. This instability culminated in a ~$0{\cdot}7 \times 10^6$ m^3 failure of the proximal flank of the moraine in 2005. Additional moraine failures in 2009 generated impact waves in the lake, but did not cause an outburst flood.

Rockfalls started to occur on the Mättenberg, northeast of the glacier terminus, in 2000. The source area is located between 2500 and 3000 m a.s.l., likely in warm degrading permafrost (Städelin, 2008). The accumulating rock debris is periodically evacuated by large erosive debris flows; ~$0{\cdot}7 \times 10^6$ m^3 of sediment was eroded from the debris cone between 2000 and 2005 (Städelin, 2008). Many of these debris flows reached the lake. The lake is also vulnerable to impacts by ice avalanches from steep snouts of Challifirn and Fiescher Glacier. To date, ice avalanches from Challifirm have been restricted mainly to the glacial gorge and the dam impounding the lake, and those from Fiescher Glacier have come to rest before reaching the lake. Less favorable scenarios, however, are possible.

Lower Grindelwald Glacier is likely to continue to retreat, with a potential for significant enlargement of the lake. In late 2009, authorities completed the construction of a drainage tunnel more than 2 km long to lower the level of the lake and reduce the danger of outburst floods to people and property in the corridor between Grindelwald and Interlaken, 2·5 to 20 km downstream.

The coupling of hazardous processes around Lower Grindelwald Glacier is remarkable. Yet it remains to be seen whether the Grindelwald situation is representative or an exceptional case of landscape instability in de-glacierizing high-mountain areas. The experience at Grindelwald has nevertheless shown that hazard assessments are considerably complicated where multiple processes operate and interact.

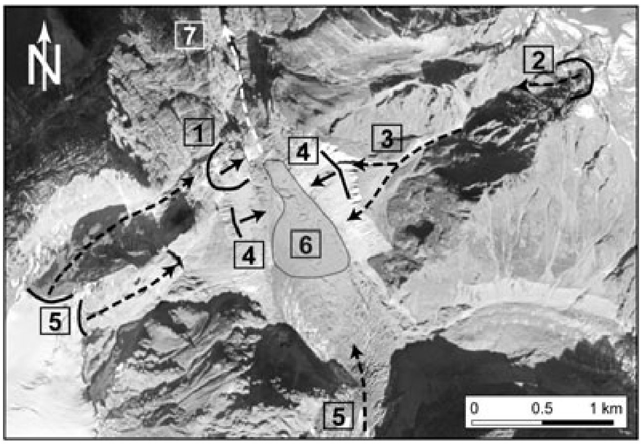

Figure 11. Interacting slope instability processes at Lower Grindelwald Glacier. (1) Rock slope failure at Schlossplatte, at the head of gorge. (2) Frequent rock falls from warm permafrost areas of Mättenberg. (3) Debris flows along ravines cut into the debris fan/moraine complex. (4) Lateral moraine slope failures. (5) Ice avalanches from Challifirn and Fiescher Glacier/Heissi Platte. (6) Glacier lake at terminus of Lower Grindelwald Glacier. (7) Outburst flood from the glacier lake. This area shows the complex interaction of processes that are closely linked to glacier retreat and downwasting and to permafrost degradation. Background image is from 1997 (Google Earth).

Mount Meager, Canada

One of the largest historic landslides in western Canada occurred at Mount Meager in southwest British Columbia on August 6, 2010. Approximately 48×10^6 m^3 of highly fractured and weathered, Pleistocene volcanic rocks detached from the southwest flank of the mountain between about 1800 and 2400 m a.s.l. (Figure 12). The detached rock mass, which contained large amounts of water, rapidly fragmented as it impacted the base of the mountain slope in the headwaters of Capricorn Creek, a steep tributary of Meager Creek. The impact, however, was sufficient to trigger seismic waves that were recorded by seismographs up to hundreds of kilometers from the source. As it struck the base of the steep mountain slope, the landslide transformed into a debris flow that traveled 7 km down Capricorn Creek at an average speed of 60 m s^{-1} to Meager Creek. The debris flow dramatically super-elevated at bends along this part of its path. It then entered the valley of Meager Creek and climbed 270 m up the northwest-facing wall of that valley. There it bifurcated into two lobes: one lobe ran nearly 4 km to the southwest up Meager Creek valley, and the other traveled northeast down Meager Creek to Lillooet River, about 12 km from the source. Landslide debris blocked Meager Creek at the mouth of Capricorn Creek for about 19 hours, impounding a lake that reached up to 1·5 km long. Debris in Lillooet River valley stemmed the flow of Lillooet River for about two hours. Concern over a possible outburst flood from the lake in the valley of Meager Creek led to the evacuation of 1500 people in the town of Pemberton.

The landslide is the third large landslide in the Capricorn Creek watershed since 1998 and the fifth such event in the Meager Creek watershed since 1930 (Jordan, 1987). Landslides at the head of Capricorn Creek in 1998 and 2009 were sourced in a colluvial apron near the toe of the 2010 failure and may

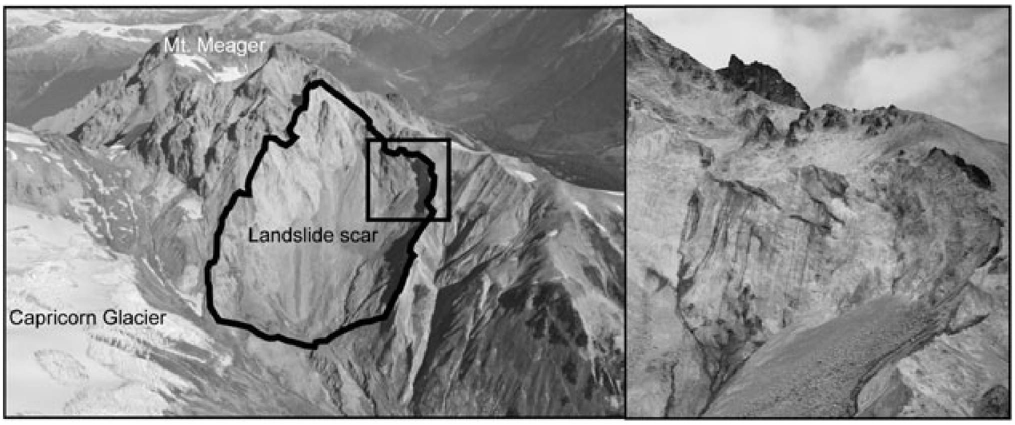

Figure 12. Failure zone of the Mount Meager landslide a few days after the event. The limit of the landslide scar is marked with a thick black line, which extends down to the surface of Capricorn Glacier (left image). A close-up of the failure zone (corresponding to the square in the left image) is shown on the right. Note that water is surfacing at several bedrock fractures.

have destabilized the rock slope above. The 1998 and 2009 landslides were large ($> 10^6\,m^3$) debris flows; the former, like the much larger 2010 landslide, blocked Meager Creek and impounded a short-lived lake (Bovis and Jakob, 2000).

Many of the landslides in the watershed, including those in Capricorn Creek, have occurred on slopes that were buttressed by glacier ice during the Little Ice Age (Holm *et al.*, 2004). Glaciers on Mount Meager have downwasted and retreated over the past century, removing lateral support from slopes that since have failed. Records of mean annual temperatures of climate stations in the region (Whistler, 650 m a.s.l.) show a warming of 2.3°C over the past three decades (Figure 5). Based on relations between rock and air temperature in the Alps and New Zealand (Gruber *et al.*, 2004b; Allen *et al.*, 2009), we suggest that warm permafrost (0 to −2°C) existed in the upper detachment zone of Mount Meager. Permafrost degradation due to strong decadal warming is likely and may have reduced the shear strength of the rock and facilitating accumulation of water in cracks, with concomitant local elevation of pore pressures.

It is furthermore interesting to note that all large historic landslides in the Meager Creek watershed occurred during or following spells of warm weather in summer. During the weeks prior to the August 6, 2010 landslide, air temperatures reached as high as about 20°C at the top of the south-facing failure zone (Figure 6). Snow and ice melt caused hydrostatic pressure and stress changes, and rapid thaw may have occurred along rock joints; both may have played a role in triggering the landslide. A more integrative perspective, however, needs to consider decadal to millennium-scale glacier retreat, poor volcanic rock strength, recent permafrost degradation, and brief warm events as factors reducing the stability of slopes at Mount Meager.

Discussion and Conclusions

Research into the detection of changes in landslide activity over the past several decades of observed atmospheric warming and identification of factors that control climate-driven changes in landslide magnitude and frequency is still limited. There is still no unambiguous evidence that the frequency or the magnitude of landslides has changed over this period. Even in Switzerland, which has a relatively complete database of mass movements, no statistically significant trend is evident in the occurrence of all types of landslides or their damage (Hilker *et al.*, 2009). However, for large rock slope failures ($> 10^5\,m^3$) in the Swiss Alps and neighboring areas, the frequency is higher during the past two decades than earlier during the twentieth century (Fischer *et al.*, 2010). The same trend has been found for small to medium rock falls in the Mont Blanc area (Ravanel and Deline, 2011).

Rather than seeking evidence of changes in landslide magnitude and frequency, some researchers have examined changes in impacts of landslides. For example, Petley (2010) concludes that in south, east, and southeast Asia both summer monsoon and tropical cyclones are important climatic triggers of landslides, although population growth is driving increases in damaging and fatal landslides. He argues that future socio-economic development will have a more important effect on landslide-related losses than climate change. These findings support a larger body of research in other fields of natural hazards that attributes the increasing losses due to floods, hurricanes, cyclones, and windstorms primarily to societal change and economic development (Pielke *et al.*, 2008). Most of this research uses a normalization approach to adjust for economic and societal change in order to identify a possible climatic signal (Bouwer *et al.*, 2007; Barredo, 2009, 2010).

In the case of landslides, more instructive evidence comes from case studies and field observations, which is the approach we took in this paper, with a summary, put into context of additional recent events, provided in Table I. It is striking that in recent years several events have occurred in the central Alps without historical precedence, for example the Ritzlihorn and Rotlaui debris flows in Guttannen and the slope failures at Monte Rosa. Recent changes in the source slopes related to glacier thinning and retreat, and also likely to permafrost degradation, may have caused these events. The frequent large landslides at Mount Meager during the twentieth and twenty-first centuries are also likely related to effects of glacier retreat and permafrost degradation.

Another important aspect of emerging instability events is the complex interplay among different landslide processes, and among landslide and river processes. Massive loss of glacier ice on the Monte Rosa east face over the past several decades and attendant changes in permafrost destabilized large parts of the mountain face and initiated a cascade of slope failures in the 1990s and 2000s. The 2010 landslide at Mount Meager strongly affected Meager Creek and Lillooet River through damming the two streams and introducing large amounts of sediment that are being transported downstream into populated areas, possibly increasing flood hazards. The Kolka-Karmadon ice–rock avalanche obliterated fluvial processes through long-term river damming and changes in sediment supply. Recent changes in the northeast flank of Ritzlihorn induced rock fall activity that triggered previously unrecorded debris flow activity on the fan at the base of the slope. Historically unprecedented landslide activity, however, may not necessarily be an indicator of climatically induced changes in slope stability; it could relate to natural variability. This possibility must be considered, given that the observation period is short compared to the return period of large landslides. For instance, the slope morphology at Mount Steele hints at a recurrence of slope failures in rock and ice, although the remoteness of the area precludes a reliable historical record. Static fatigue in the underlying bedrock, with stress changes caused by periodic loading by accumulating firn and ice, may exert an important control on slope instability, but we cannot exclude the possibility that strong decadal warming and related effects contributed to the 2007 avalanche. In this context, it should be noted that topography and geology, including rock mass strength (Moore *et al.*, 2009) or structural characteristics (Reid *et al.*, 2000; Wyllie and Mah, 2004) remain key controls on slope failures.

The case studies presented here raise the question of whether there are tipping points in geomorphic systems related to climatic change. Some researchers in climate science have argued that warming above certain thresholds can trigger a sudden shift to a contrasting dynamical regime. Examples include disruption of thermohaline circulation, complete melting of the Greenland ice sheet, or dieback of the Amazon rainforest (Schneider *et al.*, 2007; Kriegler *et al.*, 2009). Recent research on tipping points and reasons for concern that have been published since release of the IPCC Forth Assessment Report in 2007 and that are partly based on new evidence from paleo-climatic records have lowered the threshold of critical temperature increase (Füssel, 2009; Kriegler *et al.*, 2009; Smith *et al.*, 2009).

Non-linear threshold behavior has been observed in a number of geomorphic processes, including sediment entrainment and deposition (Phillips, 2003). Exceedance of such erosion thresholds may be recorded in the case of the extraordinary debris flow at Rotlaui-Guttannen in 2005, which was erosive enough to entrain large amounts of sediment on the Holocene debris fan (Huggel *et al.*, 2011).

However, sediments exposed by recent glacier retreat may remain in place for several decades or more before being mobilized by landslides or debris flows (Figure 2). This important lag effect has been observed in many events in high mountains in recent times (Rickenmann and Zimmermann, 1993; Chiarle *et al.*, 2007; Keiler *et al.*, 2010).

In summary, we identify the following mechanisms by which climate change can affect landslide activity in high mountains: (1) positive feedbacks acting on mass movement processes that can be reinforced after a climatic stimulus independently of climate change (Monte Rosa, Kolka, Ritzlihorn-Guttannen, Lower Grindelwald Glacier); (2) threshold behavior and tipping points in geomorphic systems (Ritzlihorn-Guttannen, Rotlaui-Guttannen); (3) storage of sediment and ice involving important lag-time effects (Salcantay, Rotlaui-Guttannen, Mount Meager, Mount Steele).

Acknowledgments—This work benefited from discussions and collaborations with Stephan Gruber, Luzia Fischer, Wilfried Haeberli, Andreas Kääb, Jeannette Noetzli, Andreas Hasler, Demian Schneider, Patrick Städelin, Daniel Tobler, Brian McArdell, Thomas Scheuner, John Reynolds, and several other colleagues. We much appreciate the support of Marlene Scheel and Mario Rohrer for TRMM data and Panya Lipovsky, Andy Heald, Luzia Fischer and Jürg Alean for photographs. The authors are also grateful for support provided by Fichtner, GmbH & Co. KG, EGEMSA S.A, and the Swiss Agency for Development and Cooperation. Climate data from Environment Canada, US National Oceanic and Atmospheric Administration (NOAA), the Peruvian National Meteorological and Hydrological Service (SENAMHI), and the Swiss Federal Office of Meterology and Climatology (MeteoSwiss) are also acknowledged. Finally the authors would like to thank two anonymous reviewers and the editors for their very constructive and useful comments that helped to improve the manuscript.

References

Abele G. 1974. Bergstürze in den Alpen. Ihre Verbreitung, Morphologie und Folgeerscheinungen. *Wissenschaftliche Alpenvereinshefte* **25**: 1–230.

Allen SK, Gruber S, Owens IF. 2009. Exploring steep permafrost bedrock and its relationship with recent slope failures in the Southern Alps of New Zealand. *Permafrost and Periglacial Processes* **20**: 345–356.

Ames A. 1998. A documentation of glacier tongue variations and lake development in the Cordillera Blanca, Peru. *Zeitschrift für Gletscherkunde und Glazialgeologie* **34**: 1–26.

Arenson LU, Springman SM. 2005. Triaxial constant stress and constant strain rate tests on ice-rich permafrost samples. *Canadian Geotechnical Journal* **42**: 412–430.

Augustinus PC. 1995. Glacial valley cross-profile development: the influence of *in situ* rock stress and rock mass strength, with examples from the Southern Alps, New Zealand. *Geomorphology* **14**: 87–97.

Ballantyne CK. 2002. Paraglacial geomorphology. *Quaternary Science Reviews* **21**: 1935–2017.

Barla G, Dutto F, Mortara G. 2000. Brenva glacier rock avalanche of 18 January 1997 on the Mont Blanc range, northwest Italy. *Landslide News* **13**: 2–5.

Barredo JI. 2009. Normalised flood losses in Europe: 1970–2006. *Natural Hazards and Earth System Sciences* **9**: 97–104.

Barredo JI. 2010. No upward trend in normalised windstorm losses in Europe: 1970–2008. *Natural Hazards and Earth System Sciences* **10**: 97–104.

Bathurst JC, Moretti G, El-Hames A, Moaven-Hashemi A, Burton A. 2005. Scenario modelling of basin-scale, shallow landslide sediment yield, Valsassina, Italian Southern Alps. *Natural Hazards and Earth System Sciences* **5**: 1–14.

Beniston M. 2006. August 2005 intense rainfall event in Switzerland: not necessarily an analog for strong convective events in a greenhouse climate. *Geophysical Research Letters* **33**: L05701.

Borgatti L, Soldati M. 2010. Landslides as a geomorphological proxy for climate change: a record from the Dolomites (northern Italy). *Geomorphology* **120**: 56–64.

Bouwer LM, Crompton RP, Faust E, Hoppe P, Pielke Jr RA. 2007. Confronting disaster losses. *Science* **318**: 753.

Bovis MJ, Jakob M. 2000. The July 29, 1998, debris flow and landslide dam at Capricorn Creek, Mount Meager Volcanic Complex, southern Coast Mountains, British Columbia. *Canadian Journal of Earth Sciences* **37**: 1321–1334.

Buma J. 2000. Finding the most suitable slope stability model for the assessment of the impact of climate change on a landslide in southeast France. *Earth Surface Processes and Landforms* **25**: 565–582.

Caine N. 1980. The rainfall intensity: duration control of shallow landslides and debris flows. *Geografiska Annaler, Series A, Physical Geography* **62**: 23–27.

Carey M, Huggel C, Bury J, Portocarrero C, Haeberli W. In Press. An integrated socio-environmental framework for climate change adaptation and glacier hazard management: lessons from Lake 513, Cordillera Blanca, Peru. *Climatic Change*.

Chiarle M, Iannotti S, Mortara G, Deline P. 2007. Recent debris flow occurrences associated with glaciers in the Alps. *Global and Planetary Change* **56**: 123–136.

Clague JJ, Evans SG. 2000. A review of catastrophic drainage of moraine-dammed lakes in British Columbia. *Quaternary Science Reviews* **19**: 1763–1783.

Collison A. Wade S, Griffiths J, Dehn M. 2000. Modelling the impact of predicted climate change on landslide frequency and magnitude in SE England. *Engineering Geology* **55**: 205–218.

Corominas J. 2000. Landslides and climate. In *Proceedings, 8th International Symposium on Landslides*, Bromhead E, Dixon N, Ibsen ML (eds). A.A. Balkema: Rotterdam; vol. **4**, 1–33.

Cossart E, Braucher R, Fort M, Bourlès DL, Carcaillet J. 2008. Slope instability in relation to glacial debuttressing in alpine areas (Upper Durance catchment, southeastern France): evidence from field data and ^{10}B cosmic ray exposure ages. *Geomorphology* **95**: 3–26.

Costa J, Schuster R. 1988. The formation and failure of natural dams. *Geological Society of America Bulletin* **100**: 1054–1068.

Crosta GB, Clague JJ. 2009. Dating, triggering, modelling, and hazard assessment of large landslides. *Geomorphology* **103**: 1–4.

Crosta GB, Frattini P. 2003. Distributed modelling of shallow landslides triggered by intense rainfall. *Natural Hazard Earth System Sciences* **3**: 81–93.

Crozier M. 2010. Deciphering the effect of climate change on landslide activity: a review. *Geomorphology* **124**: 260–267.

Cruden DM, Varnes DJ. 1996. Landslide types and processes. In *Landslides, Investigation and Mitigation*, Turner AK, Schuster RL (eds). National Research Council, Transportation Research Board, Special Report 247. National Research Council: Washington, DC; 36–75.

Dapples F, Lotter A, van Leeuwen J, van der Knaap W, Dimitriadis S, Oswald D. 2002. Paleolimnological evidence for increased landslide activity due to forest clearing and land-use since 3600 cal BP in the western Swiss Alps. *Journal of Paleolimnology* **27**: 239–248.

Davies MCR, Hamza O, Harris C. 2001. The effect of rise in mean annual temperature on the stability of rock slopes containing ice-filled discontinuities. *Permafrost and Periglacial Processes* **12**: 137–144.

Dehn M, Burger G, Buma J, Gasparetto P. 2000. Impact of climate change on slopestability using expanded downscaling. *Engineering Geology* **55**: 193–204.

Eisbacher GH, Clague JJ. 1984. *Destructive Mass Movements in High Mountains: Hazard and Management*, Geological Survey of Canada Paper 84–16. Geological Survey of Canada: Ottawa; 230 pp.

Evans SG, Clague JJ. 1988. Catastrophic rock avalanches in glacial environments. In *Proceedings, 5th International Symposium on Landslides*, Lausanne; 1153–1158.

Evans SG, Clague JJ. 1994. Recent climatic change and catastrophic geomorphic processes in mountain environments. *Geomorphology* **10**: 107–128.

Evans SG, Tutubalina OV, Drobyshef VN, Chernomorets SS, Dougall S, Petrakov DA, Hungr O. 2009. Catastrophic detachment and high-velocity long-runout flow of Kolka Glacier, Caucasus Mountains, Russia in 2002. *Geomorphology* **105**: 314–321.

Fischer L, Amann F, Moore J, Huggel C. 2010. Assessment of periglacial slope stability for the 1988 Tschierva rock avalanche (Piz Morteratsch, Switzerland). *Engineering Geology* **116**: 32–43.

Fischer L, Eisenbeiss H, Kääb A, Huggel C, Haeberli W. 2011. Monitoring topographic changes in a periglacial high-mountain face using high-resolution DTMs, Monte Rosa East Face, Italian Alps. *Permafrost and Periglacial Processes* **22**(2): 140–152.

Fischer L, Huggel C. 2008. Methodogical design for stability assessments of permafrost-affected high-mountain rock walls. In *Proceedings, 9th International Conference on Permafrost*, Kane DL, Hinkel KM (eds). Fairbanks, Alaska; 439–444.

Fischer L, Huggel C, Kääb A, Haeberli W. In Press. Rapid topographic changes and increased erosion rates on a steep glacierized high-mountain face caused by massive slope failures. *Earth Surface Processes and Landforms*.

Fischer L, Kääb A, Huggel C, Noetzli J. 2006. Geology, glacier retreat and permafrost degradation as controlling factors of slope instabilities in a high mountain rock wall: the Monte Rosa east face. *Natural Hazards and Earth System Sciences* **6**: 761–772.

Fischer, L. 2010. *Slope Instabilities on Perennially Frozen and Glacierized Rock Walls: Multi-scale Observations, Analysis and Modelling. PhD Thesis*. University of Zürich: Zürich. pp. 218.

Frei C, Schöll R, Fukutome S, Schmidli J, Vidale PL. 2006. Future change of precipitation extremes in Europe: intercomparison of scenarios from regional climate models. *Journal of Geophysical Research* **111**: D06105.

Füssel H. 2009. An updated assessment of the risks from climate change based on research published since the IPCC Fourth Assessment Report. *Climatic Change* **97**: 469–482.

Gardelle J, Arnaud Y, Berthier E. 2011. Contrasted evolution of glacial lakes along the Hindu Kush Himalaya mountain range between 1990 and 2009. *Global and Planetary Change* **75**: 47–55.

Geertsema M, Clague JJ, Schwab JW, Evans SG. 2006. An overview of recent large catastrophic landslides in northern British Columbia, Canada. *Engineering Geology* **3**: 120–143.

Grove JM. 1987. Glacier fluctuations and hazards. *The Geographical Journal* **153**: 351–369.

Gruber S, Haeberli W. 2007. Permafrost in steep bedrock slopes and its temperature-related destabilization following climate change. *Journal of Geophysical Research* **112**: F02S18. DOI. 10.1029/2006JF000547

Gruber S, Hoelzle M, Haeberli W. 2004a. Permafrost thaw and destabilization of Alpine rock walls in the hot summer of 2003. *Geophysical Research Letters* **31**: L13504. DOI. 10.1029/2006JF000547

Gruber S, Hoelzle M, Haeberli W. 2004b. Rock wall temperatures in the Alps: modelling their topographic distribution and regional differences. *Permafrost and Periglacial Processes* **15**: 299–307.

Günzel FK. 2008. Shear strength of ice-filled rock joints. In *Proceedings, 9th International Conference on Permafrost*, Kane DL, Hinkel KM (eds), Fairbanks, Alaska; 581–586.

Guzzetti F, Peruccacci S, Rossi M, Stark CP. 2007. Rainfall thresholds for the initiation of landslides in central and southern Europe. *Meteorology and Atmospheric Physics* **3**: 239–267.

Guzzetti F, Peruccacci S, Rossi M, Stark CP. 2008. The rainfall intensity–duration control of shallow landslides and debris flows: an update. *Landslides* **5**: 3–17.

Haeberli W. 1983. Frequency and characteristics of glacier floods in the Swiss Alps. *Annals of Glaciology* **4**: 85–90.

Haeberli W. 2008. Changing views of changing glaciers. In *Darkening Peaks – Glacial Retreat, Science and Society*, Orlove B, Wiegandt E, Luckman BH (eds). University of California Press: Berkeley, CA; 23–32.

Haeberli W, Alean J. 1985. Temperature and accumulation of high altitude firn in the Alps. *Annals of Glaciology* **6**: 161–163.

Haeberli W, Alean J-C, Müller P, Funk M. 1989. Assessing risks from glacier hazards in high mountain regions: some experiences in the Swiss Alps. *Annals of Glaciology* **13**: 77–101.

Haeberli W, Huggel C, Kääb A, Oswald S, Polkvoj A, Zotikov I, Osokin N. 2004. The Kolka-Karmadon rock/ice slide of 20 September 2002 – an extraordinary event of historical dimensions in North Ossetia (Russian Caucasus). *Journal of Glaciology* **50**: 533–546.

Haeberli W, Huggel C, Kääb A, Polkvoj A, Zotikov I, Osokin N. 2003. Permafrost conditions in the starting zone of the Kolka-Karmadon rock/ice slide of 20 September 2002 in North Osetia (Russian Caucasus). *Proceedings, 8th International Conference on Permafrost*, Zürich, Switzerland; extended abstracts; 49–50.

Haeberli W, Kääb A, Paul F, Chiarle M, Mortara G, Mazza A, Deline P, Richardson S. 2002. A surge-type movement at Ghiacciaio del Belvedere and a developing slope instability in the east face of Monte Rosa, Macugnaga, Italian Alps. *Norwegian Journal of Geography* **56**: 104–111.

Haeberli W, Wegmann M, Vonder Mühll D. 1997. Slope stability problems related to glacier shrinkage and permafrost degradation in the Alps. *Eclogae Geologica Helvetica* **90**: 407–414.

Hählen N. 2010. Murgänge Spreitgraben Guttannen. In *Fachleute Naturgefahren Schweiz*, Herbstkurs 2010, FAN; 11.

Harris C, Arenson LU, Christiansen HH, Etzelmüller B, Frauenfelder R, Gruber S, Haeberli W, Hauck C, Hölzle M, Humlum O, Isaksen K, Kääb A, Lehning M, Lütschg MA, Matsuoka N, Murton JB, Nötzli J, Phillips M, Ross N, Seppälä M, Springman SM, Vonder Mühll D. 2009. Permafrost and climate in Europe: geomorphological impacts, hazard assessment and geotechnical response. *Earth Science Reviews* **92**: 117–171.

Hasler A. 2011. *Thermal Conditions and Kinematics of Steep Bedrock Permafrost*. PhD thesis, Department of Geography, University of Zurich: Switzerland. pp. 164.

Hasler A, Gruber S, Haeberli W. 2011. Thermal offset and advective heat fluxes in fractured bedrock permafrost. *The Cryosphere Discussion* **5**: 721–753.

Hewitt K, Clague JJ, Orwin JF. 2008. Legacies of catastrophic rock slope failures in mountain landscapes. *Earth Science Reviews* **87**: 1–38.

Hilker N, Badoux A, Hegg C. 2009. The Swiss flood and landslide damage database 1972–2007. *Natural Hazards and Earth System Sciences* **9**: 913–925.

Hoelzle M, Darms G, Suter S. 2010. Evidence of accelerated englacial warming in the Monte Rosa area, Switzerland/Italy. *The Cryosphere Discussion* **4**: 2277–2305.

Holm K, Bovis M, Jakob M. 2004. The landslide response of alpine basins to post-Little Ice Age glacial thinning and retreat in southwestern British Columbia. *Geomorphology* **57**: 201–216.

Huffman GJ, Adler RF, Bolvin DT, Gu G, Nelkin EJ, Bowman KP, Hong Y, Stocker EF, Wolff DB. 2007. The TRMM Multisatellite Precipitation Analysis (TMPA): quasi-global, multiyear, combined-sensor precipitation estimates at fine scales. *Journal of Hydrometeorology* **8**: 38–55.

Huggel C. 2009. Recent extreme slope failures in glacial environments: effects of thermal perturbation. *Quaternary Science Reviews* **28**: 1119–1130.

Huggel C, Caplan-Auerbach J, Gruber S, Molnia B, Wessels R. 2008a. The 2005 Mt. Steller, Alaska, rock–ice avalanche: a large slope failure in cold permafrost. In *Proceedings, 9th International Conference on Permafrost*, Kane DL, Hinkel KM (eds). Fairbanks, Alaska; 747–752.

Huggel C, Caplan-Auerbach J, Wessels R. 2008b. Recent extreme avalanches: triggered by climate change? *EOS, Transactions, American Geophysical Union* **89**: 469–470.

Huggel C, Haeberli W, Kääb A, Bieri D, Richardson S. 2004a. Assessment procedures for glacial hazards in the Swiss Alps. *Canadian Geotechnical Journal* **41**: 1068–1083.

Huggel C, Kääb A, Reynolds JM, Heald A. 2004b. Impact-oriented models for potential lake outbursts and ASTER-based application in the Peruvian Andes. In *Proceedings Fachtagung Schweizerische Geomorphologische Gesellschaft*, March 28–29, 2003, Erstfeld; 129–143.

Huggel C, McArdell B, Scheuner T, Haeberli W, Stoffel M. 2011. Remobilization of Alpine Holocene debris fans: do climate change impacts increase debris flow hazards? *European Geoscience Union, Geophysical Research Abstracts* **13**: EGU2011–5316.

Huggel C, Salzmann N, Allen S, Caplan-Auerbach J, Fischer L, Haeberli W, Larsen C, Schneider D, Wessels R. 2010. Recent and future warm extreme events and high-mountain slope stability. *Philosophical Transactions of the Royal Society* **A368**: 2435–2459.

Huggel C, Zgraggen-Oswald S, Haeberli W, Kääb A, Polkvoj A, Galushkin I, Evans SG. 2005. The 2002 rock/ice avalanche at Kolka/Karmadon, Russian Caucasus: assessment of extraordinary avalanche formation and mobility, and application of QuickBird satellite imagery. *Natural Hazards and Earth System Sciences* **5**: 173–187.

Intergovernmental Panel on Climate Change (IPCC). 2007. Climate change 2007: the physical science basis. In *Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*, Solomon S, Qin D, Manning M, Chen Z, Marquis M, Averyt KB, Tignor M, Miller HL (eds). Cambridge University Press: Cambridge. http://www.ipcc.ch/ipccreports/assessments-reports.htm

Iverson RM. 2000. Landslide triggering by rain infiltration. *Water Resources Research* **36**: 1897–1910.

Ivy-Ochs S, von Poschinger A, Synal H-A, Maisch M. 2009. Surface exposure dating of the Flims landslide, Graubünden, Switzerland. *Geomorphology* **103**: 104–112.

Jakob M, Hungr O. 2005. *Debris-flow Hazards and Related Phenomena*. Springer: Berlin; 739.

Jakob M, Lambert S. 2009. Climate change effects on landslides along the southwest coast of British Columbia. *Geomorphology* **107**: 275–284.

Jakob M, Weatherly H. 2003. A hydroclimatic threshold for landslide initiation on the North Shore Mountains of Vancouver, British Columbia. *Geomorphology* **54**: 131–156.

Jordan RP. 1987. *Impacts of Mass Movements on Rivers in the Southern Coast Mountains, British Columbia: Summary Report*, Environment Canada, Inland Waters Directorate, Report IWD- HQ-WRB-SS-87-3. Environment Canada: New Brunswick.

Keiler M, Knight J, Harrison S. 2010. Climate change and geomorphological hazards in the eastern European Alps. *Philosophical Transactions of the Royal Society A: Mathematical, Physical and Engineering Sciences* **368**: 2461–2479.

Kershaw JA, Clague JJ, Evans SG. 2005. Geomorphic and sedimentological signature of a two-phase outburst flood from moraine-dammed Queen Bess Lake, British Columbia, Canada. *Earth Surface Process and Landforms* **30**: 1–25.

Korup O, Clague JJ. 2009. Natural hazards, extreme events, and mountain topography. *Quaternary Science Reviews* **28**: 977–990.

Korup O, Tweed F. 2007. Ice, moraine, and landslide dams in mountainous terrain. *Quaternary Science Reviews* **26**: 3406–3422.

Kotlyakov VM, Rototaeva OV, Nosenko GA. 2004. The September 2002 Kolka glacier catastrophe in North Ossetia, Russian Federation: evidence and analysis. *Mountain Research and Development* **24**: 78–83.

Kriegler E, Hall JW, Held H, Dawson R, Schellnhuber HJ. 2009. Imprecise probability assessment of tipping points in the climate system. *Proceedings of the National Academy of Sciences* **106**: 5041–5046.

Kyselỳ J, Beranová R. 2009. Climate-change effects on extreme precipitation in central Europe: uncertainties of scenarios based on regional climate models. *Theoretical and Applied Climatology* **95**: 361–374.

Larsen MC, Simon A. 1993. A rainfall intensity-duration threshold for landslides in a humid-tropical environment, Puerto Rico. *Geografiska Annaler, Series A, Physical Geography* **75A**: 13–23.

Lateltin O, Haemmig C, Raetzo H, Bonnard C. 2005. Landslide risk management in Switzerland. *Landslides* **2**: 313–320.

Lipovsky PS, Evans SG, Clague JJ, Hopkinson C, Couture R, Bobrowsky P, Ekström G, Demuth MN, Delaney KB, Roberts NJ. 2008. The July 2007 rock and ice avalanches at Mount Steele, St. Elias Mountains, Yukon, Canada. *Landslides* **5**: 445–455.

Lugon R, Stoffel M. 2010. Rock–glacier dynamics and magnitude–frequency relations of debris flows in a high-elevation watershed: Ritigraben, Swiss Alps. *Global and Planetary Change* **73**: 202–210.

Malet J-P, Durand Y, Remaître A, Maquaire O, Etchevers P, Guyomarc'h G, Déqué M, van Beek LPH. 2007. Assessing the influence of climate change on the activity of landslides in the Ubaye Valley. In *Proceedings, International Conference on Landslides and Climate Change – Challenges and Solutions*, McInnes R, Jakeways J, Fairbank H, Mathie E (eds). Taylor & Francis: London; 195–205.

Marchi L, Arattano M, Deganutti AM. 2002. Ten years of debris-flow monitoring in the Moscardo Torrent (Italian Alps). *Geomorphology* **46**: 1–17.

Meehl G, Stocker T, Collins W, Friedlingstein P, Gaye A, Gregory J, Kitoh A, Knutti R, Murphy J, Noda A, Raper S, Watterson I, Weaver A, Zhao Z. 2007. Global climate projections. In *Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*, Solomon S, Qin D, Manning M, Chen Z, Marquis M, Averyt KB, Tignor M, Miller HL (eds). Cambridge University Press: Cambridge; 747–845.

Montgomery DR, Dietrich WE. 1994. A physically based model for the topographic control of shallow landsliding. *Water Resources Research* **30**: 1153–1171.

Moore JR, Sanders JW, Dietrich WE, Glaser SD. 2009. Influence of rock mass strength on the erosion rate of alpine cliffs. *Earth Surface Processes and Landforms* **34**: 1339–1352.

Noetzli J, Gruber S. 2009. Transient thermal effects in Alpine permafrost. *The Cryosphere* **3**: 85–99.

Noetzli J, Gruber S, Kohl T, Salzman N, Haeberli W. 2007. Three-dimensional distribution and evolution of permafrost temperatures in idealized high-mountain topography. *Journal of Geophysical Research* **112**: F02S13. DOI. 10.1029/2006JF000545

Oppikofer T, Jaboyedoff M, Keusen HR. 2008. Collapse of the eastern Eiger flank in the Swiss Alps. *Nature Geosciences* **1**: 531–535.

Paul F, Haeberli W. 2008. Spatial variability of glacier elevation changes in the Swiss Alps obtained from two digital elevation models. *Geophysical Research Letters* **35**: L21502. DOI: 10.1029/2008GL034718

Petley DN. 2010. On the impact of climate change and population growth on the occurrence of fatal landslides in south, east and SE Asia. *Quarterly Journal of Engineering Geology & Hydrogeology* **43**: 487–496.

Phillips JD. 2003. Sources of nonlinearity and complexity in geomorphic systems. *Progress in Physical Geography* **27**: 1–23.

Pielke J, Gratz J, Landsea CW, Collins D, Saunders MA, Musulin R. 2008. Normalized hurricane damage in the United States: 1900–2005. *Natural Hazards Review* **9**: 29–42.

Prager C, Ivy-Ochs S, Ostermann M, Synal H-A, Patzelt G. 2009. Geological considerations and age of the catastrophic Fernpass rockslide (Tyrol, Austria). *Geomorphology* **103**: 93–103.

Pralong A, Funk M. 2006. On the instability of avalanching glaciers. *Journal of Glaciology* **52**: 31–48.

Ravanel L, Deline P. 2011. Climate influence on rockfalls in high-Alpine steep rockwalls: the north side of the Aiguilles de Chamonix (Mont Blanc massif) since the end of the 'Little Ice Age'. *The Holocene* **21**: 357–365.

Rebetez M, Lugon R, Baeriswyl PA. 1997. Climatic change and debris flows in high mountain regions: the case study of the Ritigraben torrent (Swiss Alps). *Climatic Change* **36**: 371–389.

Reid ME, Christian SB, Brien DL. 2000. Gravitational stability of three-dimensional stratovolcano edifices. *Journal of Geophysical Research* **105**: 6043–6056.

Richardson SD, Reynolds JM. 2000. An overview of glacial hazards in the Himalayas. *Quaternary International* **65**: 31–47.

Rickenmann D, Zimmermann M. 1993. The 1987 debris flows in Switzerland: documentation and analysis. *Geomorphology* **8**: 175–189.

Sauber JM, Ruppert NA. 2008. Rapid ice mass loss: does it have an influence on earthquake occurrence in southern Alaska? *Geophysical Monograph* **179**: 369–384.

Scheel M, Rohrer M, Huggel C, Santos D, Silvestre E, Huffman GJ. 2010. Evaluation of TRMM Multi-satellite Precipitation Analysis (TMPA) performance in the central Andes region and its dependency on spatial and temporal resolution. *Hydrology and Earth System Sciences Discussion* **7**: 8445–8586.

Scheuner T, Keusen HR, McArdell B, Huggel C. 2009. Murgangmodellierung mit dynamisch-physikalischem und GIS-basiertem Fliessmodell. *Wasser Energie Luft* **101**: 15–21.

Schiefer E, Menounos B, Wheate R. 2007. Recent volume loss of 371 British Columbian glaciers, Canada. *Geophysical Research Letters* **34**(L16503): 372. DOI: 10.1029/2007GL030780

Schneider S, Semenov S, Patwardhan A, Burton I, Magadza C, Oppenheimer M, Pittock A, Rahman A, Smith J, Suarez A, Yamin F. 2007. Assessing key vulnerabilities and the risk from climate change. In *Climate Change 2007: Impacts, Adaptation and Vulnerability. Contribution of Working Group II to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*, Parry ML, Canziani OF, Palutikof JP, van der Linden JP, Hanson CE (eds). Cambridge University Press: Cambridge; 779–810.

Sidle RC, Ochiai H. 2006. Landslides, processes, prediction and land use. *American Geophysical Union, Water Resources Monograph* **18**: 312.

Smith JB, Schneider SH, Oppenheimer M, Yohe GW, Hare W, Mastrandrea MD, Patwardhan A, Burton I, Corfee-Morlot J, Magadza CH, Füssel H-M, Pittock AB, Rahman A, Suarez A, van Ypersele J-P. 2009. Assessing dangerous climate change through an update of the Intergovernmental Panel on Climate Change (IPCC) “reasons for concern”. *Proceedings of the National Academy of Sciences* **106** (11): 4133.

Sosio R, Crosta GB, Hungr O. 2008. Complete dynamic modeling calibration for the Thurwieser rock avalanche (Italian Central Alps). *Engineering Geology* **100**: 11–26.

Städelin P. 2008. Integrative Process Analysis at Lower Grindelwald Glacier: Current Problems and Future Scenarios, Master Thesis, University of Zurich (in German).

Stokes CR, Gurney SD, Shahgedanova M, Popovnin V. 2006. Late-20th-century changes in glacier extent in the Caucasus Mountains, Russia/Georgia. *Journal of Glaciology* **52**: 99–109.

Suter S, Hoelzle M. 2002. Cold firn in the Mont Blanc and Monte Rosa areas, European Alps: spatial occurrence and statistical models. *Annals of Glaciology* **35**: 9–18.

Trenberth K, Jones P, Ambenje P, Bojariu R, Easterling D, Klein Tank A, Parker D, Rahimzadeh F, Renwick J, Rusticucci M, Soden B, Zhai P. 2007. Observations: surface and atmospheric climate change. In *Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*, Solomon S, Qin D, Manning M, Chen Z, Marquis M, Averyt KB, Tignor M, Miller HL (eds). Cambridge University Press: Cambridge; 235–336.

Wegmann M, Gudmundsson GH, Haeberli W. 1998. Permafrost changes in rock walls and the retreat of alpine glaciers: a thermal modelling approach. *Permafrost and Periglacial Processes* **9**: 23–33.

Werder MA, Bauder A, Funk M, Keusen HR. 2010. Hazard assessment investigations in connection with the formation of a lake on the tongue of Unterer Grindelwaldgletscher, Bernese Alps, Switzerland. *Natural Hazards Earth System Sciences* **10**: 227–237.

Wieczorek GF, Glade T. 2005. Climatic factors influencing occurrence of debris flows. In *Debris Flow Hazards and Related Phenomena*, Jakob M, Hungr O (eds). Springer: Berlin; 325–362.

Wyllie DC, Mah CC. 2004. *Rock Slope Engineering Civil and Mining.* Spon Press: New York.

Zimmermann M, Haeberli W. 1992. Climatic change and debris flow activity in high-mountain areas – a case study in the Swiss Alps. *Catena Supplement* **22**: 59–72.

Zimmermann M, Mani P, Romang H. 1997. Magnitude–frequency aspects of alpine debris flows. *Eclogae Geologica Helvetica* **90**: 415–420.

Zumbühl HJ, Steiner D, Nussbaumer SU. 2008. 19th century glacier representations and fluctuations in the central and western European Alps: an interdisciplinary approach. *Global and Planetary Change* **60**: 42–57.

