Ice loss from glaciers and permafrost and related slope instability in high-mountain regions

Philip Deline[a], Stephan Gruber[b], Florian Amann[c], Xavier Bodin[a], Reynald Delaloye[d], Jer^ome Failletaz[e], Luzia Fischer[m], Marten Geertsema[f], Marco Giardino[g], Andreas Hasler[h], Martin Kirkbride[i], Michael Krautblatter[j], Florence Magnin[a], Samuel McColl[k], Ludovic Ravanel[a], Philippe Schoeneich[l], and Samuel Weber[j]

EDYTEM, Universite Savoie Mont Blanc, CNRS, Chambery, France[a] Department of Geography and Environmental Studies, Carleton University, Ottawa, ON, Canada[b] Department of Geography and Environmental Studies, Carleton University, Ottawa, ON, Canada[c] Department of Geosciences, Geography, University of Fribourg, Fribourg, Switzerland[d] Department of Geography, University of Zu€rich, Zu€rich, Switzerland[e] Ministry of Forests, Lands, and Natural Resource Operations, Prince George, BC, Canada[f] GeoSitLab, Dipartimento di Scienze della Terra, Universita` di Torino, Torino, Italy[g] SensAlpin GmbH, Davos, Switzerland[h] Geography, University of Dundee, Dundee, United Kingdom[i] Technische Universita€t Mu€nchen, Mu€nchen, Germany[j] School of Agriculture and Environment, Massey University, Palmerston North, New Zealand[k] Institut d'Urbanisme et de Geographie Alpine, Universite Grenoble Alpes, CNRS, Grenoble, France[l] Lucerne University of Applied Sciences and Arts, Lucerne, Switzerland[m]

Abstract

The present time is a significant stage in the adjustment of mountain slopes to climate change and specifically atmospheric warming. This review examines the state of understanding of the responses of mid-latitude alpine landscapes to recent cryospheric change and summarizes the variety and complexity of documented landscape responses involving glaciers, moraines, rock and debris slopes, and rock glaciers. These indicate how a common general forcing translates into varied site-specific slope responses according to material structures and properties, thermal and hydrological environments, process rates, and prior slope histories. Warming of permafrost in rock and debris slopes has demonstrably increased instability, manifest as rock glacier acceleration, rockfalls, debris flows, and related phenomena. Changes in glacier geometry influence stress fields in rock and debris slopes, and some failures appear to be accelerating toward catastrophic failure. Several sites now require expensive monitoring and modeling to design effective risk-reduction strategies, especially where new lakes form and multiply hazard po- tential, and new activities and infrastructure are developed.

15.1 Introduction

Processes resulting from the instability of steep slopes are often compounded by the effects of the changing cryosphere: High-mountain slopes are frequently characterized by ice and firn cover. Less visible, but not less common, is the presence of permafrost: subsurface material with a temperature below 0°C during at least two consecutive years, which likely contains ice. In contrast to the well- studied relationship between rock-slope instability and glacier retreat (cf. Ballantyne, 2002), the effects of permafrost and its dynamics on bedrock slope stability are a more recent field of research (Gruber and Haeberli, 2007).

Cryosphere changes have been documented for historic and Holocene times. Since the termination of the Little Ice Age (LIA), the extent of mountain glaciers has decreased worldwide, with an accel- eration of their shrinkage during recent decades in relation to the increase in mean air temperature. Glaciers in the European Alps have lost 35% of their total area from 1850 until 1975, and almost 50% by 2000 (Zemp et al., 2006), while historically unprecedented global glacier decline in the early 21st century (Zemp et al., 2015) is currently exposing more rock and debris slopes at the terrain surface.

The Summary for Policymakers of the IPCC Special Report on the Ocean and Cryosphere in a Chang- ing Climate (2019) underscores the importance and policy relevance of slope instability related to ice loss from glaciers and permafrost. Based on its chapter on “High Mountain Areas” (Hock et al., 2019), the assessment finds that permafrost thaw and glacier retreat have decreased the stability of high-mountain slopes and are projected to further decrease the stability of the terrain. Corresponding events are projected to occur in new locations and to increase the risk for infrastructure, cultural, tourism, and recreational assets. Because climate change impacts in the cryosphere and their societal consequences operate on time horizons, which are longer than those of governance arrangements, the ability of societies to adequately prepare for and respond to long-term cryosphere changes is challenged. Here, scientific research and coordinated monitoring can contribute to empowering forward-looking decision making.

Monitoring of ground temperature in debris slopes started in the 1970s (Marchenko et al., 2007) and was extended to bedrock slopes in the 1990s. These measurements allow investigating multi-year trends and seasonal variations in the permafrost (PERMOS, 2013). Similarly, strong seasonal and multi-annual fluctuations and longer term acceleration of rock-glacier creep have been observed during the last two decades (Delaloye et al., 2008; Hartl et al., 2016; PERMOS, 2019). An apparent increase in slope fail- ures in steep rockwalls observed over recent decades in the European Alps (Ravanel and Deline, 2008, 2011; Allen and Huggel, 2013) also suggests the current degradation of the permafrost.

Due to the high potential energy inherent in steep environments and the possibility of compound events, the consequences of slope instability can be far-reaching. Such events can have devastating consequences, as illustrated by the dramatic Kolka-Karmadon ice-rock avalanche triggered in Septem- ber 2002 in the Caucasus (Haeberli et al., 2004; Evans et al., 2009). Lake formation due to glacier shrinkage, and expanding human activities are further increasing the risks (cf. Schaub et al., 2013; Haeberli et al., 2016, 2017).

Slope instability and mass movements, in general and when related to ice loss, occur at a wide range of magnitudes and frequencies. This chapter focuses on less-frequent events that often have larger mag- nitudes and are (or can be) tied to the unprecedented ice loss during recent decades. Often, these events are also less understood than the more frequent ones.

This chapter describes the consequences of the loss of perennial ice, at the surface and in the sub- surface, rather than processes related to seasonal change. It has been compiled and edited by P. Deline and S. Gruber and represents a joint effort of 17 authors contributing their specific expertise. It is struc- tured in two parts: The first revisits the mechanisms of cryosphere control on slope stability, dealing with instabilities in hanging glaciers (J. Failletaz, L. Fischer), debuttressing and ice unloading (S. McColl), and thermal and hydrologic changes in bedrock (A. Hasler, M. Krautblatter and S. Weber) and sediments (R. Delaloye); the second part illustrates the relationship between ice loss and slope instability with case studies, organized partly by mechanism and partly by morphology. Concerning

bedrock slopes, case studies are presented on the Monte Rosa East face (L. Fischer), and rockfalls from rockwalls in the Mont Blanc massif where permafrost is modeled (F. Magnin and L. Ravanel). Rela- tions between debuttressing and rock instability or deformation are illustrated by cases in the Mont Blanc massif (L. Ravanel), British Columbia and Alaska ranges (M. Geertsema), Mueller Glacier, New Zealand (S. McColl), at Alp B€aregg, Switzerland (P. Deline), and Mont de la Saxe, Italy (M. Giar- dino). Concerning debris slopes, the processes (sliding, gullying, and breaching) that impact the mo- raine stability because of glacier shrinkage are described (P. Deline and M. Kirkbride). Rock-glacier displacement changes, ground-ice melting, and debris supply from permafrost areas are exposed through several Alpine case studies (X. Bodin, R. Delaloye and P. Schoeneich). Finally, case studies exemplify interactions between differing processes, which translate cryospheric change into far- reaching and difficult-to-forecast events (F. Amann, P. Deline and M. Geertsema). A synopsis and an outlook on likely future challenges conclude this chapter.

15.2 Mechanisms of cryosphere control on slope stability

15.2.1 Unstable ice

Changes in the thermal regime and geometry of steep glaciers can impact the stability of glaciers and lead to ice avalanches (Fig. 15.1). The periodic or occasional breakoff of moderate volumes of ice is a common and predictable (Faillettaz et al., 2016) part of the mass balance of many hanging glaciers (Pralong and Funk, 2006). This section focuses on unusual instabilities and their relationship with changing temperature and topography.

Three general conditions of glacier instabilities can be distinguished based on the thermal properties at the ice/bedrock contact (Faillettaz et al., 2015): (1) cold-based glaciers frozen to the bedrock; (2) polythermal glaciers stabilized by the cold ice frozen to the bedrock in the frontal and lateral parts; and (3) temperate glaciers sliding on bedrock. The thermal state of the ice is related to the glacier lo- cation, aspect, and elevation (R€othlisberger, 1981; Alean, 1985; Faillettaz et al., 2012). The factors affecting the stability of steep ice and glaciers include en- and subglacial hydrology, support from flat- ter downslope glacier parts and lateral bedrock abutments, adhesion of cold and polythermal ice on bedrock, and cohesion with more stable upslope ice (R€othlisberger, 1981; Alean, 1985; Wagner, 1996; Haeberli et al., 1999; Pralong, 2005; Faillettaz et al., 2015).

Climate trends may introduce feedback mechanisms involving changes in englacial and subglacial temperature, water infiltration, accumulation rate, and surface geometry. Such changes may reduce the stability of parts of, or the entire, steep glacier (Wegmann et al., 1998; Pralong and Funk, 2006; Fischer et al., 2013). Atmospheric warming affects the hydraulic conditions of the ice and increases the pro- portion of temperate ice in steep high-mountain faces. Rising temperature increases flow velocities and sliding processes within the ice mass and at the glacier bed, causing a changed stress field in the glacier. Furthermore, the tensile strength of the warmed and water-containing ice is reduced (Haeberli et al., 1997, 1999). Forced by atmospheric warming, melting can occur at higher elevation, where increased infiltration of meltwater through firn and ice may lead to basal warming and cause a decrease in the effective basal pressure and friction at the glacier bed (Pralong, 2005; Faillettaz et al., 2011).

Changes in glacier geometry and the surrounding topography can influence ice stability. A retreat of the glacier terminus from a moderate into a steep slope (e.g., due to an ice avalanche from the frontal part or a decoupling from a valley glacier) can change the stress field within the glacier and may lead to in- stability. Rockfalls or debris flows can lead to a destabilization of hanging glaciers when decreasing the

FIG. 15.1

The deposit of the ice avalanche from the front of the Allalingletscher (Swiss Alps) after the 1 Mm^3 event on July 31, 2000.

Photograph: archive VAW.

lateral or downslope abutments (Fischer et al., 2011, 2013). Finally, the thickness and mass of a steep glacier are limited for a given slope gradient and thermal regime, as they affect the shear stresses deter- mining stability (Pralong, 2005). As a consequence, additional load from mass-movement deposits can trigger an ice failure when the basal shear resistance or tensile strength is exceeded (Fischer et al., 2013).

Driven by climate change, these factors and processes might lead to large instabilities in steep ice not only at locations with known events (e.g., Alean, 1985; Pralong, 2005) but also at locations without precedence for such events. Although some presently hazardous glaciers will become harmless in the near future because of their retreat, others may evolve toward a critical situation and become danger- ous. A timely identification of such newly developing critical situations represents a challenge in haz- ard assessment.

15.2.2 Glacial debuttressing and unloading

Aside from creating a void into which slope failures can freely move, the thinning and retreat of glaciers can destabilize slopes through: (1) glacial debuttressing; (2) thermo-mechanical damage; and (3) crustal rebound (McColl, 2012).

Glacial debuttressing refers to the loss of mechanical slope support provided by glacial ice, an ef- fect that diminishes as glaciers thin, warm, or experience a rise in groundwater (McColl and Davies, 2012). For rock slopes, where glacial erosion changes the slope geometry (notably glacial oversteepen- ing, sensu Augustinus, 1995), debuttressing can weaken the slope, possibly initiating failure synchro- nously with ice retreat (McColl and Davies, 2012) and involving failure of the remaining buttressing ice (e.g., the 1975 Devastation Glacier rock avalanche; Evans and Clague, 1988). More often, rock slopes may fail long after ice retreat (McColl, 2012). Rock-slope response may be large and cata- strophic failure, adjustment through smaller rockfalls, or more gradual as gravitational deformation. For moraines or drift-covered slopes that are constructed against the glacier body, glacier thinning in- duces an immediate slope adjustment through deformation or sliding (Blair, 1994; Hugenholtz et al., 2008). The slopes may continue to adjust through other mass-wasting processes once ice has retreated (e.g., Ballantyne and Benn, 1994; Draebing and Eichel, 2018), with deformation prolonged where ice- cored moraines thaw (Ravanel et al., 2018). Sometimes, debuttressing may induce the simultaneous failure of rock and sediment, as has been observed alongside the thinning Tasman Glacier in New Zeal- and (Blair, 1994; Fig. 15.2). Whatever the material, debuttressed slopes will continue to adjust until they reach a new strength-equilibrium independent of ice support.

Thermomechanical damage to rock slopes during the cycles of varying glacier cover is another pre- paratory factor for their failure. The removal of confining rock or ice overburden from glacial erosion or glacier retreat, or cycles of ice loading and unloading, can induce damage to rock (Nichols Jr., 1980; Gr€amiger et al., 2017) and propagate fractures. Further, paraglacial thermal shock, as glacier retreat exposes rock to new thermal regimes, may further enhance rock damage (Gr€amiger et al., 2018). This damage may weaken rock slopes, and coalescing fractures may act as potential failure surfaces for mass

FIG. 15.2

Downwasting of the Tasman Glacier, Southern Alps, New Zealand, since the Little Ice Age has exposed and unloaded the lateral moraine wall and the rockslope behind. The moraine has responded by a combination of shallow failures and sliding, damaging and threatening huts constructed on it. Scarps showing vertical displacements are visible in both moraine and rockslope.

Photograph: S. Winkler.

movements. A spatial coincidence between rock slope failures and locations experiencing damage or high stress changes (Cossart et al., 2008; Jarman and Harrison, 2019) provides support for this idea, and the time-dependent nature of damage accumulation may explain the lag-time of some post-glacial rock slope failures (Ballantyne et al., 2014a).

The unloading associated with melting glacial ice can cause larger scale crustal rebound via iso- static and elastic flexural adjustment. Its timescale depends on the rheological properties of the upper mantle and strength of the crust, and it may continue for tens of thousands of years (Stewart et al., 2000). In some locations, isostatic uplift in the order of hundreds of meters has occurred during the Holocene, and differential uplift and changes to stress-fields around faults is sufficient to cause coseis- mic faulting (Arvidsson, 1996; Hampel et al., 2010). Spatial and temporal links between glacial re- bound and slope instability have been documented (Sanchez et al., 2010; Cossart et al., 2013), providing a further explanation for lag-times between ice-retreat and mass movement activity (Ballantyne et al., 2014b).

15.2.3 Bedrock permafrost: Thermal and hydrologic change

Permafrost affects the hydrological and mechanical behavior of the naturally fractured bedrock. Water can freeze in cavities and fissures at 0°C, whereas the freezing point in the pore system is at 0.1°C to 1.5°C in water-saturated, tough (low-porosity) alpine and arctic bedrock (Krautblatter et al., 2010; Draebing and Krautblatter, 2012). Rock temperature change and water percolation in the cold-fractured rock may be accompanied by latent heat exchange, leading to altered hydraulic permeability, mechan- ical strength, and stress fields.

Pore diameter, ion concentration in pore water, and pressure control the proportion of unfrozen wa- ter in bedrock at temperatures below 0°C (Krautblatter, 2009; cf. Watanabe and Mizoguchi, 2002). As a consequence, the hydraulic permeability of compact rock (interjoint rock mass) changes gradually be- low the freezing point (Kleinberg and Griffin, 2005). In this temperature range, crack initiation and ice- lens formation are expected to occur where the cryogenic stresses exceed rock strength and overburden. For porous chalk, ice segregation has been illustrated in the laboratory (Murton et al., 2006, 2016) whereas acoustic emissions measured in metamorphic alpine bedrock indicate frost-cracking activity significantly below 0°C (Girard et al., 2013).

Hydrology is mainly controlled by the permeability of the fracture system. Pogrebiskiy and Chernyshev (1977) found that the permeability of frozen fissured granite is one to three orders of magnitude lower than the permeability of identical thawed rock. Hence, saturated fractured bedrock permafrost acts as an aquiclude and can support perched water tables. Even though the observations of ice-saturated fractures in bedrock exist (cf. Gruber and Haeberli, 2007), unfrozen fracture systems (micro-taliks) can lead water deeply into frozen systems (Tang and Wang, 2006; Haeberli and Gruber, 2008). Water flowing into permafrost can warm fractures at depth and is unlikely to freeze in subsur- face channels if not impounded (Hasler et al., 2011a). This effect may be reduced where basal ice layers between snow and rock impede the percolation of snowmelt into fractures (Phillips et al., 2016). The heating of water at the surface, for example, on warm rock, prior to percolation is important for the thermal erosion of ice within rock fractures, a process that may be tracked with ambient seismic vi- bration measurements (Weber et al., 2018).

The mechanical strength of bedrock permafrost and the subsurface stress field change with tem- perature and hydraulic conditions. The stress changes due to cryostatic and hydrostatic pressures and thermomechanical forcing. Increased cryostatic pressures up to a few MPa can derive from ice,

whereas volume expansion of (fast-) freezing trapped water may cause larger stresses that decline rapidly due to ice extrusion (Matsuoka and Murton, 2008). Hydrostatic pressure can derive from perched groundwater blocked by impermeable rock (Fischer et al., 2010) or from meltwater input into fractured bedrock with limited permeability. It can reduce the friction at a potential failure plane via effective stress or increase gravitational, down-slope forces. Thermal expansion and contraction cause stress change and movement in fractured bedrock (Gischig et al., 2011; Luethi et al., 2015). Changing rock- and ice-mechanical strength affects the macroscopic strength of intact bedrock and rock-ice interfaces, and the deformation of ice in fractures (Krautblatter et al., 2013). In intact rock, fracture initiation thresholds and friction within the rough fractures decrease when thawing (Mellor, 1973; Dwivedi et al., 2000). Along the rock-ice interfaces, fracturing is facilitated by increasing tem- peratures. Similarly, the deformability and fracturing of ice are also temperature-dependent (Sanderson, 1988; Budd and Jacka, 1989). A comprehensive criterion for ice-filled rock joints by Mamot et al. (2018) demonstrated that both warming and unloading lead to a significant drop in shear resistance, possibly resulting in self-reinforced propagation of slope failure (Fig. 15.3). All these pro- cesses work effectively for different stress levels and deformation velocities. Ice-filled fractures with less than 20-m rock overburden are likely to be more affected by ice mechanics, whereas for higher rock overburden, rock mechanical changes upon thaw become more relevant (Krautblatter et al., 2013). Thus, rockfall of lower magnitude is more prone to be affected by ice-mechanical changes, whereas slow deformation and rockfall with high magnitude is more likely to be affected by rock- mechanical changes. Although shallow stress changes tend to represent more short-term reactions (except ice segregation) to particular meteorological events and annual extremes, deep-seated strength changes all predict a stability reduction with increased rock temperature. It is also significant to consider that over timescales of years to thousands of years, rock strength degrades due to sub- critical fracture propagation (Kemeny, 2003; Voigtl€ander et al., 2018) as well as repeated freeze- thaw cycles (Jia et al., 2015).

Electrical resistivity tomography (Krautblatter and Hauck, 2007; Magnin et al., 2015a) and refraction seismic tomography (Draebing and Krautblatter, 2012; Krautblatter and Draebing, 2014) have demonstrated their capability to monitor frozen rock distribution in bedrock, both in the labora- tory and in the field, and assess unfrozen water content (Hauck, 2013), and may even differentiate ther- mal conditions in permafrost (Krautblatter et al., 2010). Capacitive resitivity imaging and polarization tomography are tested for future applications (Murton et al., 2016; Coperey et al., 2019).

Different explanations for the role of liquid water for rock instability exist. One hypothesis is that hydrostatic pressures due to the sealing of rock surfaces by ice may play a vital part in the destabili- zation of rock slopes (Terzaghi, 1962), as illustrated by coupled hydro-mechanical modeling of the Tschierva rock avalanche by Fischer and Huggel (2008). Another is that hydrothermal heat transport causes warming at depth and corresponding strength reductions (Hasler et al., 2012), which is sup- ported by field data (Weber et al., 2019). A study in Norway concluded that the primary control for the deformation process was meltwater percolation into fractures during summer, with refreezing, ice formation, and temperature increase in the lower part of the fractures from 1°C to 0°C (Blikra and Christiansen, 2014).

Permafrost degradation can occur on different scales in time and space. For example, extreme sum- mer or disappearance of thin ice cover leads to thicker active layers with fast (seasonal) reactions and small events. On the other hand, the effect of decadal-scale atmospheric warming is delayed at deep- seated potential failure planes even in steep topography where signal propagation is faster due to warm- ing penetrating the rock mass from opposing faces (Noetzli and Gruber, 2009).

FIG. 15.3

Progressive failure in a warming permafrost rock slope displaying thermal and normal stress conditions before and after detachment of a first slab (Mamot et al., 2018). Both can initiate failure: (B2) progressive thermal warming (i.e., permafrost degradation) occurs within years, but (B1) sudden unloading develops even faster within days.

15.2.4 Debris permafrost: Hydrological and thermal change

Thermal and hydrological conditions in frozen accumulations of coarse debris are primarily responding to the shifts in temperature occurring at the ground surface. An increased temperature thereby induces a thickening of the active layer, which is also controlled by the ice content in the upper permafrost layer (Scherler et al., 2013). In the Swiss Alps, for permafrost conditions close to 0°C, the active-layer thick- ness in debris terrain remains almost constant at some sites but has increased by up to 10 cm year^{1} over the last 15–20 years at others (Zenklusen Mutter and Phillips, 2012a; PERMOS, 2019). A thicker active layer induces changes in the supra-permafrost water circulation: Drainage through the active layer in- creases in autumn along with thaw depth (Buchli et al., 2013). In steep terrain, mass movements can start from the base of, or from within, the active layer whose thickness determines the volume of avail- able loose material. Frozen ground at depth limits the retrogressive erosion and therefore the magnitude of an event, but it can also favor the triggering of the mass movement by acting as an aquiclude (Fig. 15.4).

The warming of permafrost favors water percolation and talik formation (Zenklusen Mutter and Phillips, 2012b). During warming, the unfrozen water content in frozen ground increases and, if the water is not drained, the mechanical strength of frozen, and especially the ice-rich, ground decreases (Arenson et al., 2007). As a consequence, the permafrost creep rate increases nonlinearly, accelerating significantly when the temperature is approaching 0°C (K€a€ab et al., 2007; Staub et al., 2016); under cooling conditions, the response is reversed. Interannual variations in the displacement rate of rock glaciers are almost synchronous and similar in range in a given region (Delaloye et al., 2008; PERMOS, 2019). Destabilization of rock glaciers with strong acceleration has been reported

FIG. 15.4

Left: ground ice exposure (dark grey) in the starting zone of a 4000-m^3 rockfall at the front of a deep-seated slide in permafrost conditions, Grabengufer, Swiss Alps, September 2010; Right: ongoing regressive erosion scar produced by a series of debris flows triggered at the front of an active rock glacier, likely due to the limitation of snowmelt percolation by the frozen ground, Gugla-Bielzug, Swiss Alps, June 2013. The exposed ground ice started to melt for weeks to years after both events.

Photographs: R. Delaloye.

(see Section 15.3.5 Debris slopes), and in most cases, the onset appears to be in direct response to warmer conditions (e.g., Delaloye et al., 2013; Eriksen et al., 2018; Marcer et al., 2019).

When a coarse debris layer is permeable, air circulation is favored (Delaloye and Lambiel, 2005) and its efficiency increases with reduced sealing of voids by ice. The so-called chimney effect (air ad- vection) occurs in talus slopes and coarse-grained rock glaciers (Morard et al., 2008). It contributes to preserving cold ground conditions, particularly in the lower and deeper parts of the affected landforms, especially by the aspiration of cold air deep into the terrain in wintertime, while heat is expelled toward the upper part of the landform. Ground ice could thus be formed or preserved in inactive rock glaciers, and talus slopes under climatic conditions are unfavorable for permafrost formation in differing configurations.

15.3 Case studies

15.3.1 Bedrock slopes: Hanging glaciers and permafrost

15.3.1.1 Dynamics of hanging glaciers on the Monte Rosa east face

The Monte Rosa east face is a prominent example of dramatic changes in hanging glaciers, including large ice avalanches. Extending from 2200 m to >4600 m a.s.l., large sections are covered by hanging glaciers, firn fields, and steep glaciers connected downslope to the more gently sloping valley glaciers. Between the end of the LIA and the 1980s, the hanging glaciers and firn fields have changed only slightly (Fischer et al., 2011). Around the beginning of the 21st century, however, the ice cover has experienced an accelerated loss in extent and thickness (Haeberli et al., 2002; K€a€ab et al., 2004; Fischer et al., 2006, 2011), leaving large parts of the underlying rock unprotected against mechanical and thermal erosion (cf. Noetzli et al., 2003). Since around 1990, the events of frequent small- and several large-volume ice and rock avalanches and debris flows have led to a significant reduction in the ice-covered area. The total material loss in the face was around 25 Mm^3 from 1988 to 2007, both from steep glaciers and from subjacent bedrock.

The measured englacial temperatures at the Colle Gnifetti site, located directly above the east face, show cold ice throughout, with a mean annual surface temperature near 14°C and basal ice at a depth of 124 m slightly below 12°C (Haeberli and Funk, 1991; Hoelzle et al., 2011). The terminus of the largest steep glacier has long been at 3300m a.s.l., an elevation close to the lower boundary of the modeled permafrost conditions (Fischer et al., 2006). This implies changing thermal conditions from cold-based glaciers in the uppermost part of the face to polythermal ones in the middle elevations, and warm-based glaciers at the foot of the slope. The three-dimensional permafrost changes within the face are closely linked with the evolving ice and snow cover on its surface. Thus, the obvious decrease in glaciation is most likely accompanied by a less visible change in the permafost regime of the Monte Rosa east face (Fischer et al., 2006, 2013).

Increased ice avalanching caused the shrinking or disappearance of several hanging glaciers. A con- nection between failures in steep glaciers and the underlying bedrock was suggested based on inter- pretation of time series of high-resolution digital terrain models and terrestrial photographs (Fischer et al., 2011, 2013). Failure zones were found to be spatially correlated and commonly proceeded from lower elevation upward. Around 1990, an entire cold-based hanging glacier and large parts of the un- derlying bedrock disappeared. Thereby, a steep rock wall became newly exposed in the uppermost part

of the face. The instability in the hanging glacier was most likely influenced by previous rockfall and debris flow activity right below it. The changed bedrock geometry at its terminus and additional mass movement activity below it were subsequently inducing a stepwise failure of the glacier front and re- treat of the ice mass. A second phase of increased ice avalanching occurred in the central part of the face, where the lowest part of the hanging glacier, reaching down to 3300 m a.s.l., disappeared around 1999–2000 (Fig. 15.5, glacier part below marking H). The ice failures started in the lowest part of the glacier and proceeded progressively upward, with one major ice avalanche and several small events. As the lowest part of this steep glacier was likely polythermal or temperate, rising air temperature might have affected stability (Fischer et al., 2013).

In the following years, heavy ice avalanches and rockfalls originated from the newly ice-free zone right below the retreated glacier front (Fig. 15.5), while debris flows were presumably triggered by meltwater. The continuation of rockfall during the winter pointed to a substantial change in glacier and rock conditions rather than to seasonal melt effects alone. These mass-wasting processes culmi- nated in August 2005 in a major ice avalanche, where a large part of the remaining hanging glacier failed with a maximum detachment thickness of 40m and a total volume of c. 1.2Mm^3 (Fig. 15.5, at location H). Changing ice temperature might have locally increased water percolation, indicated by strong water outflow at the front of the remaining glacier at an elevation of 3700 m a.s.l., and reduced ice stability. Additional destabilizing factors were supposedly the lack of the supporting glacier parts below, which detached in 1999, and the ongoing accumulation of rock-fall debris on the hanging gla- cier. Considerable ice volume gain up to 20 m thick took place in the area of the detachment zone of the 2005 ice avalanche in the Imseng Couloir in 2005–2007, showing that rapid ice accumulation and build-up of steep glaciers in the eroded areas following a rock or ice avalanche are possible (Fischer et al., 2011).

FIG. 15.5

Upper part of the Monta Rosa east face in the mid-1980s (left) and on August 6, 2003 (right). Some steep glaciers have totally disappeared; some have significantly lost mass. Zones with highest slope failure activity and mass waste are located in the center of the face (H).

From Ka€a€b, A., Huggel, C., Barbero, S., Chiarle, M., Cordola, M., Epifani, F., Haeberli, W., Mortara, G., Semino, P., Tamburini, A., Viazzo, G., 2004. Glacier hazards at Belvedere Glacier and the Monte Rosa east face, Italian Alps: processes and mitigation. In: Proceedings of the Interpraevent 2004, Riva/Trient, pp. 67─78.

15.3.1.2 Permafrost control on rock-fall distribution in the Mont Blanc massif

An increase in rockfalls with volumes from a few hundred to millions of m^3 is recognized as a possible impact of the current warming in the permafrost-affected rockwalls in the mountain regions (Gruber and Haeberli, 2007). Ravanel and Deline (2008, 2011) showed a recent increase in the frequency of rockfall that detached in two iconic areas of the Mont Blanc massif since 1860, with a correlation of the occurrence of the 83 rockfalls with the warmest periods (Fig. 15.6). The prevalence of extremely hot days leading up to rockfalls has also been observed in the Swiss Alps over recent decades (Allen and Huggel, 2013).

FIG. 15.6

Mean annual air temperature in Chamonix (1040 m a.s.l.) since 1934 (A) and number of rockfalls per decade in the West face of the Drus and on the North side of the Aiguilles de Chamonix, Mont Blanc massif, France (B). For the decade 2010, only the period 2010–2015 is included. Dashed lines: linear regressions.

Rockfall from permafrost slopes in the European Alps during the hot summer of 2003 has set the stage for much of the current research into the topic. Although the back-analyses of rock temperature were consistent with the expected active-layer thickening during 2003, the timing was suspect: The observed rockfall occurred almost simultaneously with seasonal air-temperature maxima, whereas active-layer thickness would be expected to peak much later (Gruber et al., 2004). Fast and linear thaw by water percolation into fractures has been suggested to accommodate this fast reaction (Hasler et al., 2011a) and additionally, convex topography warms relatively fast due to geometric effects (Noetzli and Gruber, 2009).

In the Mont Blanc massif, rockfalls have been recorded annually since 2007, and a retrospective inventory of the rockfalls that occurred in 2003 was compiled based on satellite images (Ravanel et al., 2010, 2011, 2017). This distribution of rockfalls can be analyzed with a map of simulated per- mafrost distribution in rock walls of the massif, for testing the hypothesis of a link between permafrost degradation and rock-fall triggering. As steep alpine faces have reduced snow and debris cover, tem- perature control by atmospheric variables is simpler than in gentle topography. A linear regression model explaining the mean annual rock surface temperature (MARST) with the potential incoming solar radiation and mean annual air temperature calibrated for the entire European Alps (Boeckli et al., 2012) was used for mapping MARST on a high-resolution DEM of the Mont Blanc massif (Magnin et al., 2015a). The 0°C isotherm of MARST extends as far down as 2600 m a.s.l in the most shaded faces and rises up to 3800 m in the most sun-exposed rock surfaces (Fig. 15.7), with local de- viations due to the variable effects of snow and fractures. Snow cover can warm or cool the bedrock, depending on its thickness, timing, and sun exposure (Pogliotti, 2011; Haberkorn et al., 2017; Magnin et al., 2015b, 2017), and non-conductive heat transfer in bedrock clefts may locally influence the bed- rock thermal regime (Hasler et al., 2011a, b). Nevertheless, five electrical resistivity tomographies con- firmed an accurate representation of permafrost conditions (Magnin et al., 2015c).

The comparison of the 521 rockfalls inventoried between 2003 and 2014 with the MARST map shows a pattern indicative of permafrost thaw as a driver of rockfall: 83.5% of recorded events orig- inated from rock walls with simulated MARST <0°C (9.7 events per km^2), whereas only 16.5% oc- curred with MARST >0°CC (0°C to 4°C; 1.9 event per km^2). Of the rockfalls, 53% originated from rock walls with simulated MARST between 0°C and 3°C (10.6 events per km^2). Additionally, the link with permafrost degradation is evidenced by: (i) the presence of ice in many rockfall scars; (ii) higher scar elevations during hotter summers; (iii) a sharp contrast in scar elevation between north and south faces, consistent with modeled permafrost distribution; and (iv) increased rockfall from convex topog- raphy prone to rapid warming such as pillars, spurs, and ridges.

15.3.2 Bedrock slopes and debuttressing: Rockfall

Rockfalls and rock avalanches can threaten human infrastructure and activities in the mountain areas. One factor that could increase their frequency is the impact of the current glacier shrinkage. An ac- curate observation of active rockfalls is possible in places well-frequented by people, whereas only large events like rock avalanches are usually observed in the remote areas. The possible control that glacial debuttressing exerts on rockfall and rock avalanche magnitude and frequency in the areas likely containing warm permafrost can be explored by jointly interpreting the corresponding case studies.

FIG. 15.7

Distribution of predicted mean annual rock surface temperature (MARST; based on mean annual air temperature from 1961 to 1990) and rockfalls for the period 2003 and 2007–2014 in the Mont Blanc massif, and rockfall density (i.e., number of rockfall per km^2 for each MARST class).

15.3.2.1 Rockfalls at the Are^te des Cosmiques, Mont Blanc massif

The evolution of the ice masses since the end of the LIA, and especially since the 1990s, influences the triggering of rockfall close to the glacier tongues but also at higher elevation. The Mer de Glace lost more than 100 m of ice thickness below the Montenvers since 1990, and tens of meters at many loca- tions above 3000m a.s.l. This could lead to triggering of rockfalls due to debuttressing and, indeed, 20% of the 139 collapses recorded in the massif in 2007–2009 occurred at the base of the recently ex- posed rock slopes.

Ravanel et al. (2012) have examined rockfalls at the lower Ar^ete des Cosmiques over fifteen years. Its slopes differ strongly in terms of height (around 350 m for the NW face and 50 m for the SE face) and rock structure (no particular structure visible in the NW face, fractured SE face). This 400-m-long ridge is especially relevant because of a mountain hut with 140 beds situated on its crest (3613 m a.s.l.). After the 600-m^3 rockfall that destabilized a part of the refuge in 1998, the SE face of the ridge has been regularly observed, and annually surveyed by terrestrial laser scanning since 2009. 16 rockfalls (20–256 m^3) occurred along the ridge between 2003 and 2011, and 8 (0.7–18 m^3) were detected below the hut between 2011 and 2018 (Fig. 15.8).

Topographic maps from the 1950s and 1970s show ice cover east and south-east of the hut, which is not the case today. Ice thickness reduced by 40 m between 1979 and 2003, a change that could thus have interfered with the stability of the ridge. For example, the 1998 rockfall affected a slab that had an

FIG. 15.8

Rockfalls that occurred from 2011 to 2018 below the Cosmiques Hut, on the SE face of the lower Ar^ete des Cosmiques, Mont Blanc massif, France. (A) Yellow dashed line: scar of the August 1998 rockfall (600m^3)—a portion of the refuge became unstable; smaller event (6): 0.70.2m^3; larger event (7): 18.23.2m^3. (B) Two examples of comparison of terrestrial laser scanning point clouds (Duvillard, 2019).

ice-covered base until that year, whereas a 2010 rockfall in the couloir east of the hut was possibly related to the recent lowering of the glacier. The observed presence of permafrost within the rock mass and the concentration of rockfalls during or at the end of warm periods suggest that permafrost deg- radation could have contributed to their occurrence.

The Cosmiques case highlights how the monitoring of rock slopes, permafrost, and glaciers is nec- essary for assessing the sustainability of infrastructure and the management of risks in the high- mountain areas (Bommer et al., 2010.

15.3.2.2 Rockfalls related to glacier shrinkage in British Columbia and Alaska

Due to the low population density in these mountainous areas, data are scarcer than in the European Alps, particularly on smaller events. Many post-LIA rockfalls and avalanches in British Columbia (BC) occurred on rock slopes above glaciers (Evans and Clague, 1994; Holm et al., 2004). Two-thirds of the 18 rock avalanches that were recorded in northern BC between 1973 and 2003 affected cirque walls where glaciers thinned up to several hundred meters (Geertsema and Chiarle, 2013), like, for example, the 1999 Howson rock avalanche. Here, some 0.9 Mm^3 of fractured and jointed granodiorite toppled at 1900 m a.s.l., fell and slid 150 m over a 48-degree slope onto a glacier (Schwab et al., 2003); the rubble transformed into a rock avalanche (2.7km runout), incorporated some 0.6Mm^3 of colluvium and till along the slide path, severed a natural gas pipeline, and dammed a stream. Although several large rock avalanches in BC and Alaska recently originated from permafrost-affected rockwalls (Bessette-Kirton et al., 2018; Coe et al., 2018), more modest examples with travel distances of 2–3km show a clear association with either glacier cirque walls or zonation within LIA trimlines (Fig. 15.9).

FIG. 15.9

Alaskan rockfalls that transformed into rock avalanches. Left: on West Nunatak Glacier (1.8-km runout) near Nunatak Fjord (October 1998); right: on Takhin Glacier near Tsirku River, Takhinsha Mountains (3-km runout) in 2006. The rock slope failures initiated in the recently deglaciated zones.

Photographs: M. Geertsema.

15.3.3 Bedrock slopes and debuttressing: Rock sliding and deep-seated gravitational slope deformation

Active over decades to millennia, many rock slides and deep-seated gravitational slope deformations (DSGSD) in formerly glaciated valleys—characterized by a slow deformation evolving along planes or shear zones at depth—are considered as paraglacial (Ballantyne, 2002). However, lack of dating makes the testing of this hypothesis difficult as the rock slope failures often undergo multiple and different phases in their development. Some slides have been studied because they are clearly related to post- LIA shrinkage of glaciers, while others have been monitored because of their strong hazard potential. As bedrock and debris-cover stability are strongly affected by these slope deformations, they form a source for other mass-movements like earth flows or rock avalanches.

15.3.3.1 Mueller Rockslide, New Zealand

The Mueller Rockslide (c. 1.1 Mm^2; 100–200 Mm^3) is a slow-moving rock slide in the Sealy Range of the

New Zealand Southern Alps (Fig. 15.10). Its crown is near the crest of the slope (c. 1800 m a.s.l.), and its toe is some 600–800m farther down the slope, partly hidden beneath the surface of the

FIG. 15.10

Mueller Rockslide, Southern Alps, New Zealand. The circle on the map shows the approximate rockslide location. Solid lines on the oblique view show major headscarps; the dashed line shows the approximate outline of the lower slide and the partly debuttressed toe; red circle: alpine hut. The development and continuing movement of the rockslide is thought to be related to downwasting of the glacier. Oblique image: Google Earth, April 5, 2013; map data: LINZ.

Mueller Glacier. The dip-slope failure is inferred to be sliding in bedding within the limb of an an- ticline (Lillie and Gunn, 1964; Hancox, 1994), with erosion by the Mueller Glacier and debuttressing likely to have primed the slope for failure. The slope is currently moving at 0.5 4 m year^{1}, probably responding to post-LIA debuttressing from glacier downwasting of some 150m, but the initial de- velopment of instability and movement may be older. This rockslide provides an example of a para- glacial slope failure with a long history of deformation, with its slow evolution potentially moderated by the ductile behavior of the weakly buttressing ice (McColl and Davies, 2012). As of 2019, part of the rock slide toe is free of ice and appears to be slowing, but further reduction in the ice buttress and signs of upslope retrogression of movement activity make the future of this large rock slope failure uncertain.

15.3.3.2 Schlossplatte/Alp Ba€regg, Switzerland

The instability and then collapse of a 2-Mm^3, 250-m-high, compact limestone mass in 2006–2009 at Schlossplatte, Bernese Oberland, was caused by the Lower Grindelwald Glacier shrinkage, the LIA surface of which was 300m above its 2006 elevation (Fig. 15.11). The collapsed area was subject to a high glacial compression due to its position on a topographic ridge.

After the opening of two 250-m-long valley-parallel cracks, and one month of small rockfalls, 0.17Mm^3 toppled in July 2006 (Oppikofer et al., 2008). During one year, the total subsidence of the rear block was 52 m, whereas the front block slid forward by ca. 20 m (Fig. 15.11). The rear block was fully broken up in June 2008, whereas the front block collapsed in August 2008; two residual rock columns partly toppled two months later and fully collapsed during the summer of 2009.

FIG. 15.11

Cross-section of Schlossplatte/Alp B€aregg trough, Bernese Oberland, Switzerland, with Lower Grindelwald Glacier surface elevation at the end of the Little Ice Age (Oppikofer et al., 2008, modified). Black profile: prefailure state; red profile: Summer 2007 profile, with deposit at the glacier surface; R: rear block; F: front block.

15.3.3.3 Mont de la Saxe, Italy

As a result of Pleistocene glacier fluctuations (Porter and Orombelli, 1982) and the onset of deglaci- ation predating 16.4 ka BP (Wirsig et al., 2016), the NW side of the ridge of Mont de la Saxe, Aosta Valley, developed a DSGSD, forming aligned trenches and closed depressions, and scarps and counter- slope scarps in the intensely deformed metasedimentary rocks over an area larger than 4km^2 (Fig. 15.12). The DSGSD induced block-flexural toppling on rock masses acting on the subvertical tectonic and metamorphic discontinuities. A shallow and extremely active complex rock slope failure with a volume of \$8 Mm^3 has been affecting the SW end of the ridge since the 1980s, with velocities of 1.5mm per day during the winter and 15mm per day during the spring in 2010–2012 (Crosta et al., 2013). Since it poses a high risk to the village of Courmayeur, the highway, the national road, and the access to the Mont Blanc tunnel, an intensive monitoring system has been initiated since 2002 (Frattini et al., 2015).

15.3.3.4 Taan Fiord, Alaska

Taan Fiord (Icy Bay) has been reshaped by the 17-km-retreat of Tyndall Glacier since 1961, with more than 400m of ice thinning and the glacier receding to its approximate current grounded position by 1991 (Koppes and Hallet, 2006; Higman et al., 2018). Grabens and a large rotational rock slide were

FIG. 15.12

Landforms and slope instabilities on the NW side of Mont de la Saxe, Aosta Valley, Italy. (1) Landslide; (2) fracture; (3) trench; (4) active scarp; (5) relict scarp; (6) depression; (7) debris cone. Imagery SCT Geoportal Aosta Valley Region; design: W. Alberto.

identified in 1995 in the deglaciated head of the fjord, with progressive deformation until 2015 (Meigs et al., 2006; Higman et al., 2018). Finally, resulting from the rapid glacial debuttressing, a c. 76 Mm^3 rock-slope failure occurred above the terminus of Tyndall Glacier in October 2015. It travelled c. 3 km at $45 ms^{1}$, resulting in a M4.9 seismic signal, and generated a tsunami in the fjord (Section 15.3.6.3).

15.3.4 Moraine stability and glacier shrinkage

Glacial debuttressing of proximal moraine slopes is commonly associated with deep-seated slope fail- ure or smaller slumps, and once adjusted to the glacier withdrawal, gullying may continue for one to two centuries and solifluction and breaching can occur.

15.3.4.1 Slides

Many studies have documented translational sliding in moraines (Table 15.1), although debris is gen- erally released through a combination of sliding, slumping, debris flow, and blockfall (Ballantyne, 2002). Mattson and Gardner (1991) documented 25 slides from ice-cored moraines at Boundary Gla- cier (Alberta Rockies), which initiated at the ice-debris interface. Most were triggered by rainfall, with some resulting from ice melt reducing the strength of the overlying sediment.

Large deformation of lateral moraines has become more frequent since the end of the 2000s, from the Bionnassay, Mer de Glace, Tacul, and Miage Glaciers (Mont Blanc massif) to Findelen and Bel- vedere Glaciers (Monte Rosa massif), and Forni Glacier (Ortles-Cevedale Group). Deformation started earlier at other glaciers. The sharp-crested right lateral moraine of Athabasca Glacier (Alberta Rocky Mountains) rises 150m above the glacier foreland. Hugenholtz et al. (2008) found that deformation along a 540-m-long section began in the early 1950s, and the (possibly ice-cored) moraine underwent

Table 15.1 Records of deformation of alpine moraines.

Glacier	Number, magnitude, and mode of failures	Preparatory factors	Triggering factors	Reference
Boundary (Alaska, USA) Athabasca (Canada) Tortin (Switzerland) Lower Grindelwald (Switzerland) Belvedere (Italy)	25 slides Post-1960 gravity deformation with multiple failure surfaces 10 slides of 24—1138m^3 2009 slide of 0.3Mm^3 Two slides of 300m (post-	Ice-cored moraine degradation Glacier thinning >65 m Glacier thinning 17.7m (2007— 2014); warming of permafrost Glacier thinning > 80 m (1985—	Rainfall Traffic vibration? Rock glacier push from upslope? Contact between base of summer active layer and buried glacier ice Not	Mattson and Gardner (1991) Hugenholtz et al. (2008) Ravanel et al. (2018) Stoffel and Huggel (2012) Chiarle and Mortara (2001); Mortara et al. (2009);

	1889), 250m length (post-1954), and 500m length (post-2015)	2000) Glacier thinning	specified	
Locce (Italy)	Rapid rotation slide in 2005, collapsing in 2007.	Glacier thinning; melting of buried ice	Not specified	Mortara (pers. com., 2019) Tamburini (2009)
Tasman (New Zealand)	Progressive down-glacier failure >5km since 1960s. Multiple slow slides of moraine and valley side bedrock	Glacier thinning > 150 m (1890–2000)	Infiltration from moraine-dammed lake Rainfall; snowmelt	Blair (1994)

progressive bulging and gravitational deformation from the late 1960s, to create a gap in the moraine crest and a displacement of 55m toward the glacier with a vertical lowering of 41m. A network of fractures up to 40 m long has transformed the sharp crest into a series of discontinuous crests. The re- striction of deformation to this section (a quarter of its length) may be due to both the interaction with adjacent rock glaciers impinging on the distal side of the moraine, and vibrations from summer tourist vehicles since 1961.

At the Belvedere Glacier, the crest of a curved section of the right lateral moraine has been sliding since at least 1889 on a 40-degree plane, generating a 300-m-long, double-crested moraine (Chiarle and Mortara, 2001; Mortara et al., 2009) with a vertical lowering of 25–33 m in 2016; this system has sur- vived three glacier advances in the 20th century (Fig. 15.13). Sometime before 1951, a translational slide detached several meters below the crest of the proximal side of another section of the same mo- raine several hundred meters upstream; this 250-m-long section was probably eroded by the 1960s ad- vance (Mortara et al., 2009). Most recently, a 500-m-long crack formed in the upper section of the

moraine in May 2015. The vertical lowering reached 17m in July 2017 (A. Tamburini, pers. com.; Fig. 15.13), resulting from the lowering of 40m of the surface of the Belvedere from 2008 to 2017 after its surge-type flow acceleration of 1999–2004 (Haeberli et al., 2002).

Melting of internal ice has also caused surface deformation in some moraines, for example, at the terminal moraine of Locce Glacier (Monte Rosa massif; Fig. 15.13), that caused the hut Rifugio Para- disio to be abandoned by 1975 because of toppling (Mortara et al., 2009). Lateral moraines at several large valley glaciers in New Zealand show a range of large-scale failure types. At Tasman Glacier, the right-lateral moraine displays a complex suite of mass movement styles (Fig. 15.2). Failure com- menced 10 km upstream of the contemporary terminus, where post-1890 glacier thinning was greatest above the debris-covered tongue (Kirkbride and Warren, 1999). Gullying of the enlarging proximal moraine face preceded the initiation of deep-seated failure in the 1960s. By 1986, this had propagated 4.5km toward the terminus, at a rate apparently controlled by lowering of the glacier surface (Blair, 1994). Subsequently, failure surfaces activated by 1986 have continued to develop. Slumping com- prises translational sliding of rigid slabs down the proximal moraine face, rotational slumps incorpo- rating blocks of the entire superposed moraine ridge, slumps incorporating the alluvial fill of the adjacent lateral morainic trough and valley-side debris cones, and deep-seated rock slope failure extending c. 250 m above the trough. The entire valley side now appears to be at residual strength, per- haps posing the risk of catastrophic failure in coming decades after the supporting glacier commenced rapid calving retreat in 2008 (Dykes et al., 2011).

FIG. 15.13

Right-lateral moraine of the debris-covered Belvedere Glacier, Monte Rosa massif, Italy. Left: double-crested moraine formed since 1889; large terminal moraine of Locce Glacier in the middle ground. Right: 500-m-long double-crested moraine in the upper section, formed since spring 2015. Dashed white line: glacier level in 2002.

Photographs: A. Tamburini and G. Mortara.

15.3.4.2 Gullying

Gullying is the fretting of moraine faces by weathering and erosion, to form geometrically regular suites of gullies, most commonly on over-steepened proximal lateral moraine slopes, where gullied faces can extend for several kilometers up valley. Gullying results from frost weathering, surface wash, wind ablation, debris fall, and avalanche (Ballantyne, 2002; Curry et al., 2006).

The upper proximal faces of large moraines (>120-m-height, >30 degree) are characteristically covered by an array of parallel gullies separated by sharp ridges (Fig. 15.14), with a gully density up to 60 per km in the central Swiss Alps (Curry et al., 2006) or 110 per km in the western Norway (Curry, 1999). Gullies are wider and deeper when incised by streams flowing along the mountain side; debris flows along the poorly consolidated and sometimes ice-cored moraines are initiated by rain- storms or melting of snowbeds at gully heads (Ballantyne, 2002). The lower moraine faces comprise debris cones or debris aprons derived from the gullies above.

Glacier thinning of the Mer de Glace in 1890–2008 was 200 m at its terminus and 140 m at the Mon- tenvers. The right-lateral moraine is incised by many streams coming from the upper glacier basins (Fig. 15.14). The moraine's distal face above the two recent proglacial lakes has 50- to 200-m-high gullies, on average 10–20 m deep (up to 64 m), with longitudinal profile slope angles of 21–41 degrees. The rate of gully erosion here averages 50–100mmyear$_{1}$ with a maximum of 300mmyear$_{1}$, com- parable to the highest values measured on LIA moraines in Norway (Curry, 1999) and Switzerland (Curry et al., 2006).

However, the period of strong sediment release is of short duration: Gullies in the Swiss Alps reach their maximum dimensions within ca. 50 years and stabilization occurs within 80–140 years of degla- ciation (Curry et al., 2006; Draebing and Eichel, 2018). Once stabilized, the paraglacial gully system in Norway comprises a bedrock-floored upper source area, a midslope area of gullies with sidewall slope

FIG. 15.14

Active erosion on the right lateral moraine of Mer de Glace, Mont Blanc massif, France, is evident in comparing these two pictures separated by one century. Left: photograph by Spelterini from his balloon, August 1909 (Kramer and Stadler, 2007). Right: 2008 orthophoto draped on 4-m DEM from RGD 73-74.

Courtesy M. Le Roy.

inclination of c. 25 degree, and debris talus at the slope foot (Curry, 1999). In the Turtmann Valley (Switzerland), solifluction reworks the debris-flow lobes at the base of the larger gullies and forms turf-banked lobes and terraces onto the lower part of the distal face; this solifluction stage gives way to a stabilization phase when shrubs and trees start to colonize the slopes, at first where blocky material is dominant (Draebing and Eichel, 2018; Eichel et al., 2018).

15.3.4.3 Breaching

Moraine breaching may occur with the glacier still present, and also, a long time after paraglacial sed- iment supply from upslope has been exhausted. It represents the terminal grade of linear incision in this environment and can result from either internal processes (ice-core melting, saturation collapse) or the impact of external erosive events (e.g., ice avalanching, lake-outburst flooding). Where the breached moraine dams a lake, downstream flooding forms a serious hazard due to its magnitude, rapid onset, and unpredictability (e.g., Nostetuko Lake in 1983: Clague and Evans, 2000).

Water saturation due to precipitation was responsible for the September 1993 failure of the LIA moraine of the Mulinet Glacier, Levanne massif, Italy, at 2525m a.s.l. A heavy and prolonged rain- storm eroded a 50-m-deep and 450-m-long breach, facilitated by buried glacier ice, and resulted in a 0.8-Mm^{3} debris flow travelling over 5.6km and flooding the village of Forno Alpi Graie (Mortara et al., 1995).

Breaching usually affects terminal moraines (e.g., Dolent Glacier in 1990: Lugon et al., 2000), because they are located in more vulnerable positions, but lateral moraines may also be affected. The left lateral moraine of the Belvedere Glacier was progressively buried by an aggrading debris cone from a lateral proglacial stream. As the cone reached the moraine crest by the 1940s, it triggered incision and removal of a section of this moraine by the stream between 1950 and 1968 (Mortara et al., 2009).

15.3.5 Debris slopes in permafrost areas

Hazards related to permafrost degradation in debris slopes can be subdivided into (1) direct effects of accelerated creep of cohesive masses of frozen debris such as rock glaciers; (2) indirect effects of debris supply from moving masses; and (3) thermokarst.

15.3.5.1 Rock glacier movement

Several types of behavior of rock glaciers and other landforms affected by permafrost creep can be distinguished based on surface kinematics (Fig. 15.15; Schoeneich et al., 2014):

- Type 1: Moderate acceleration, modulated by multi-annual velocity fluctuations in the range of a few cmyear^{1} to >2myear^{1} (Delaloye et al., 2008; Staub et al., 2016; PERMOS, 2019).
- Type2a:Abnormalaccelerationoftheentireormostofthemovingmasswithopeningofcrevasses, scarps, or cracks on the surface (Roer et al., 2008; Darrow et al., 2016; Scotti et al., 2016; Marcer et al., 2019) with velocities ranging from about 1 to >10myear^{1}.
- Type 2b: Very strong acceleration with known velocities >80myear^{1}, which can last a few years (Delaloye et al., 2013; Eriksen et al., 2018).
- Type 2c: Acceleration and dislocation of the lower part of the moving mass, with the formation of scarps (Roer et al., 2008; Delaloye and Morard, 2011; Delaloye et al., 2013).

FIG. 15.15

Evolution trajectories of rock-glacier behavior with increasing temperature (Schoeneich et al., 2014).

Type3:Collapseofthelowerpartofthefrozenmovingmass,whichbreaksdownasadebrisflow;a new front develops from the scarp (Bodin et al., 2017).

Type 4: Deceleration of Types 2 and 3 during months to decades after destabilization.

Type 1 Laurichard rock glacier (Ecrins massif, France) provides one of the longest available series of surface displacement (Bodin et al., 2009, 2018). Its velocity has increased during 1988–2018, likely in response to permafrost warming, and its multi-annual behavior has been similar to observations in the Swiss Alps (PERMOS, 2019). After a first peak in 2004, velocity abruptly decreased until 2007, before increasing again until 2016. As shown by Staub et al. (2016) for the Becs-de-Bosson rock glacier, this behavior can be related to the lagged effect of the ground surface temperature deviation over the pre- vious two years.

Types 2–3 are often referred to as “destabilized” and many of these phenomena started in the post- 1980s warm decades, although possible cases of earlier initiation are also known (e.g., Marcer et al., 2016). For instance, the destabilization of Type 2c Petit-Velan and Type 2b Tsate rock glaciers (Valais) in 1988–1995 followed the strong increase in permafrost temperature that occurred around 1990 (Delaloye and Morard, 2011; Lambiel, 2011), and the collapse of the Berard rock glacier (Southern French Alps) was possibly triggered by the summer heat-waves of 2003 and 2006 (Bodin et al., 2017).

The internal composition of the moving mass is an important control of destabilization: The Type 4 Berard moving mass (Southern French Alps), composed of fine schist debris prone to sliding when water saturated, corresponds to a “pebbly” or “fine-grained” rock glacier according to Ikeda and Matsuoka (2006). The initiation of a destabilization phase results from the combined influence of ther- mal (permafrost temperature), geometrical/topographical (gradual changes in geometry of the moving mass over a given topography), and mechanical (e.g., increased loading induced by significant rockfall deposits) factors over different time scales. As such, it may not be only the response to permafrost warming (Delaloye et al., 2013).

The basal topography over which a rock glacier is moving is a significant factor influencing its destabilization (Avian et al., 2009; Marcer et al., 2019). A steep slope causes higher shear stress, and a convex long-profile topography induces extending flow and hence favors stretching and even splitting of the moving rock glacier, as shown by all reported destabilizations/ruptures of the lower part of rock glaciers. Deposits of rock avalanche or rockfall may add mass to a rock glacier and con- tribute to destabilization (Delaloye et al., 2013; Scotti et al., 2016). If this affects the rooting zone, a longer time is necessary for the effects to reach the terminal part of the rock glacier, e.g., 25 years to achieve the “mechanical surge” of the 400-m-long Grabengufer rock glacier (Valais, Switzerland) in 2008–2012 (Delaloye et al., 2013).

15.3.5.2 Debris supply from permafrost areas

The connections between permafrost degradation and debris flows have received increased attention in the aftermath of the catastrophic rain and flooding in the Swiss Alps during the summer of 1987, which triggered numerous debris flows on steep till-covered slopes deglaciated during the past 150 years (Zimmermann and Haeberli, 1992). Rock glacier fronts provide debris downstream into mountain tor- rents (Zischg et al., 2011), although the link is not necessarily direct. At active rock-glacier fronts, de- bris supply to the torrent system is modulated by velocity variations of the rock glacier (Kummert et al., 2018). The rate of debris supply is usually in the range of tens to a few hundred m^{3} year^{1}. Episodic acceleration or destabilization phases can increase this to several ten thousand m^{3} year^{1} (Kummert and Delaloye, 2018). Melting of ground ice in debris masses may leave a large amount of loose material available to erosion, which may be limited by the coarse size of debris and a facilitated infiltration of water into the ground.

The consequences of this variation in debris supply depend on the characteristics of the torrent sys- tem. In debris-limited systems, additional debris supply due to permafrost degradation may lead to an increase in the magnitude and frequency of debris flows (Kummert et al., 2018). For example, an en- hanced torrent activity was triggered by increased debris supply from the overhanging Derochoir rock glacier front in the Arandellys catchment in the Mont Blanc massif (France) in the late 1890s (Mougin, 1914; Marcer et al., 2016).

15.3.5.3 Ground ice melting and thermokarst

In the European Alps, thermokarst phenomena are commonly related to buried glacier ice or avalanche deposits, and more rarely to excess ice formed in the ground. The long-term conservation of buried ice is favored by permafrost conditions. The most striking phenomena are thermokarst lakes and associated outburst floods (e.g., Gruben Glacier area: Haeberli et al., 2001). In the southern French Alps, the Lac Chauvet (2800 m a.s.l.) is an ephemeral lake on debris-covered dead ice in a proglacial area that grows for a couple of years before draining through a glacial tunnel, triggering debris flows that can dam the main river

(Assier, 1996). The phenomenon repeated at least six times since the 1930s, once in 2008. Geophysical investigations show that the ice and frozen debris are still >40m thick.

In recent years, several cases of ground ice revealed by thermokarst features were reported in the Alps in places where no presence of ground ice was suspected. This could point to an increase in the melt rate of buried ice bodies, representing an emerging hazard in periglacial mountain areas. Mapping of proglacial areas combined with permafrost distribution modeling could help identifying the potential areas of thermokarst development.

Ice-cemented debris layers can occur in aggrading talus slopes where debris and avalanche snow are deposited. The Dents Blanches (Valais) rockfall in 2006 provided a rare exposure of such a permafrost body (Gruber and Haeberli, 2009). The melt of the pore ice and the interwoven ice layers make sed- iment hitherto preserved from erosion available for transport. As a consequence, debris flows of unpre- dictable magnitude may originate from unexpected locations within talus slopes that contain ice.

15.3.6 Interactions and complex mass movements

Large slope failures in the mountain environments typically evolve into complex mass movements (Cruden and Varnes, 1996), that is, events during which one mode of movement transforms into an- other. The following cases involve transitions between movement types and illustrate how complex slope failures can lead to unexpected consequences, whereby events related to ice loss or interacting with ice can have consequences reaching much further than commonly anticipated (cf. Evans et al., 2021).

15.3.6.1 Rock avalanches onto glaciers

In 2002, ice-cored rubble in a cirque near Harold Price Creek (BC) failed and triggered a 4-km-long mass movement (Geertsema et al., 2006). The volcanic rock rubble travelled some 500m before it loaded the till and colluvium lower on the slope, transforming the moving mass into a debris avalanche that travelled 1.5km before it channelized into a debris flow. Large rock avalanches travel a much greater distance, even when little or no ice is involved: The 53 Mm^3 rock avalanche that affected Mount Meager (BC) in August 2010 travelled for 12.7 km with a small amount of ice from Capricorn Glacier incorporated in the detached rock mass (Roberti et al., 2017, 2018). On the contrary, the Alaskan rock avalanches at Mount Lituya in 2012, at Mount La Perouse in 2014 (Fig. 15.16), and on Lamplugh Gla- cier in 2016 originated in permafrost-affected rockwalls and propagated onto glaciers with horizontal travel distances of 9.2, 7.4 and 10.5 km, respectively, incorporating a huge volume of ice (Geertsema, 2012; Coe et al., 2018; Bessette-Kirton et al., 2018). The higher mobility of rock avalanches onto gla- ciers results from low-friction surface, basal pore pressure generated by frictional melting, debris flu- idization by ice and snow melting, and debris channelization by lateral moraines (Evans and Clague, 1988, 1994; Deline et al., 2015).

15.3.6.2 Rockfall-debris storage-debris flow (chain reaction)

Deposits of mass movements can be subject to time-lagged remobilization. A 1.5-Mm^3 rock mass de- tached from the north face of Piz Cengalo, Switzerland, in December 2011, and evolved into a rock avalanche, travelling 2.7 km down the Bondasca Valley. During summer 2012, heavy rainfall remobi- lized parts of the deposit, and four debris flows reached the village of Bondo, at a distance of 6 km (Baer et al., 2017). The travel angle of the rock avalanche was exceptionally steep (32 degree) for an event of this size, and debris flows extended the reach of the combined event.

On August 23, 2017, another 3 Mm^3 collapsed in the NE face of Piz Cengalo and impacted a small glacier located at its foot, removing 0.6Mm^3 of ice by an undetermined combination of melting and erosion (Mergili et al., 2019; Fig. 15.17). The rock-ice avalanche reached a runout distance of 3.4 km (travel angle of 26 degrees) and entrained a large amount of debris deposited by the 2011 rock ava- lanche; eight hikers located in the runout zone were killed. In contrast to 2011, a succession of large debris flows immediately followed the rock avalanche, notably in the absence of rainfall. The first

FIG. 15.16

The c. 7.4-km-long rock avalanche at Mount La Perouse (3269 m a.s.l.), Coast Mountains, Alaska, occurred on February 16, 2014 (Coe et al., 2018). Note the dust deposited beyond the limits of the rock avalanche deposit. Photograph: M. Geertsema.

debris flow occurred with a delay of c. 30 s, followed by 10 debris flows within 9.5 h, and 3 more 2 days later. The debris flows successively reached and filled the debris retention basin in the main valley, which had been constructed following the 2011 events: Later debris flows overtopped the basin and caused severe damage to houses and public infrastructure in the village of Bondo.

15.3.6.3 Rock slide-generated waves

When large rock slides (often originating from steep and recently deglaciated areas) enter water bodies, they create local tsunamis. These differ from tectonic tsunamis that have wave periods tens of minutes long, and maximum run-ups of 30 m (Higman et al., 2018), with periods typically an order of magnitude smaller but much larger waves. The largest historical tsunami at Lituya Bay, generated in 1958 by a c. 30-Mm^3 rock slide that crossed the 1350-m-wide Gilbert Inlet, had a period of 76 seconds and a peak runup of 524 m (Fritz et al., 2009; Fig. 15.18). A 2- to 6-m-high wave in Knight Inlet some 500 years ago destroyed a First Nation’s Village; it was generated by a 4 Mm^3 rock slide detached at a distance of 5km on an 840m high cliff oversteepened during the Last Glacial Maximum/Fraser Glaciation (Bornhold et al., 2007). A more recent rock-slide-generated wave was triggered in 2015 by the 73 Mm^3 Taan Fiord rock avalanche (see Section 15.3.3.4) that entered the 1.5-km-wide and 90-m-deep

FIG. 15.17

Piz Cengalo rock avalanche on August 23, 2017. The photographic sequence (from a video taken by Reto Salis) shows the formation of an ice jet as a consequence of the impact of the rock mass on the small glacier at the toe of the Piz Cengalo NE face.

recently deglaciated fjord to emerge and deposit debris on the opposite shore, producing a 6-km-long and 2km^2 subaqueous flow deposit (Dufresne et al., 2018; Haeussler et al., 2018). A tsunami with a maximum height of 193 m travelled the length of the 17-km-long fjord, removing forest and soil from the shores (Fig. 15.18); it extended the destructive reach of the initial rock avalanche almost six-fold (Higman et al., 2018)

FIG. 15.18

Left: A 30-Mm^3 rock avalanche at the head of Lituya Bay, Alaska, generated a wave running up more than 500 m and removing the forest from the slopes along the bay (red line: scar); the trimline of the tsunami (yellow line) is still visible today due to the difference in vegetation type. Right: A 73-Mm^3 rock avalanche in Taan Fiord, Alaska, generated a 193-m-high wave on October 17, 2015, that stripped the shores of vegetation (dashed line: scar, top at 650 m a.s.l). Tsunami inundation zone in the foreground; arrows: landslide hummocks which travelled across the fjord.

Photographs: M. Geertsema.

15.4 Conclusion and outlook

The mountain cryosphere has been changing rapidly since the late 20th century, with strong con- sequences for morphodynamics. Glaciers are shrinking at a rate never observed since the termina- tion of the Little Ice Age. This influences slope instability at different spatial and temporal scales through glacial debuttressing, stress-release fracturing, and crustal rebound. Therefore, large rock slides and deep-seated gravitational slope deformations (DSGSD) are triggered or reactivated. Denudation of moraine slopes due to glacier surface lowering allows increased sliding, gullying and breaching of moraines due to rainfall, infiltration or outburst floods. Changes in the geometry of temperate glaciers reduce abutment and can generate ice avalanches. Warming of cold-based glaciers changes rheology, basal cohesion, water content, and tensile strength of ice, which enables glacier sliding.

At the same time, but less visible at the terrain surface, permafrost also is strongly affected by at- mospheric warming. In steep rockwalls, rock temperature increase and water percolation can change hydraulic regime and mechanical strength that alter the stress field. As a consequence, rockfall fre- quency is increasing in many mountain ranges worldwide. In frozen accumulations of coarse debris, thermal changes thicken the active layer with consequences at depth on water percolation and creep velocity. This can lead to rock-glacier destabilization including collapse, whereas ice-cemented talus slopes and rock glaciers can become a source for more frequent and/or larger debris flows and rockfalls. The processes specific to a high-elevation geomorphic belt can affect lower areas and generate cascad- ing processes, with catastrophic consequences when long runout phenomena (e.g., a lake impacted by an ice-rock avalanche) threaten downvalley populations.

Many scientific advances concerning the relationship between ice loss and slope instability in high- mountain regions have been realized in recent decades, often stimulated by extreme events such as the 1987 summer flooding in the Swiss Alps, or the very hot summer of 2003 in the European Alps. As impacts of the changing climate on the cryosphere and the related morphodynamics will be reinforced during the 21st century, a race has begun between the acceleration of those phenomena and the under- standing of the underlying processes including their sensitivity and response times. To do this, coor- dinated research, observations, and monitoring are needed, as illustrated, for instance, by the WGMS for glaciers or PERMOS for Swiss permafrost, supporting long-term scientific research as well as phys- ical and numerical modeling. These future research steps will be based on an increasing international scientific collaboration, of which this chapter is in its own way an illustration.

Acknowledgments

We acknowledge the two editors, Wilfried Haeberli and Colin Whiteman, and the two external reviewers of the 2nd edition of this chapter, Norikazu Matsuoka and Martin Mergili, whose remarks and suggestions helped to improve it.

References

Alean, J., 1985. Ice avalanches: some empirical information about their formation and reach. J. Glaciol. 31 (109), 324–333.

Allen, S., Huggel, C., 2013. Extremely warm temperatures as a potential cause of recent high mountain rockfall. Glob. Planet. Chang. 107, 59–69.

Arenson, L.U., Springman, S.M., Sego, D.C., 2007. The rheology of frozen soils. Appl. Rheol. 17 (1), 1–14. Arvidsson, R., 1996. Fennoscandian earthquakes: whole crustal rupturing related to postglacial rebound. Science

274 (5288), 744–746.
Assier, A., 1996. Glaciers et glaciers rocheux de l'Ubaye. Association Sabenç a de la Valeia, Barcelonnette, p. 32.
Augustinus, P.C., 1995. Rock mass strength and the stability of some glacial valley slopes. Z. Geomorphol. 39, 55–68.
Avian, M., Kellerer-Pirklbauer, A., Bauer, A., 2009. LiDAR for monitoring mass movements in permafrost environments at the cirque Hinteres Langtal, Austria, between 2000 and 2008. Nat. Hazards Earth Syst. Sci. 9, 1087–1094.
Baer, P., Huggel, C.H., McArdell, B.W., Frank, F., 2017. Changing debris flow activity after sudden sediment input: a case study from the Swiss Alps. Geol. Today 33 (6), 216–223. https://doi.org/10.1111/gto.12211. Ballantyne, C.K., 2002. Paraglacial geomorphology. Quat. Sci. Rev. 21, 1935–2017. https://doi.org/10.1016/S0277-3791(02)00005-7.
Ballantyne, C.K., Benn, D.I., 1994. Paraglacial slope adjustment and resedimentation following recent glacier retreat, Få bergstølsdalen, Norway. Arct. Alp. Res. 26 (3), 255–269.
Ballantyne, C.K., Sandeman, G.F., Stone, J.O., Wilson, P., 2014a. Rock-slope failure following Late Pleistocene deglaciation on tectonically stable mountainous terrain. Quat. Sci. Rev. 86, 144–157.
Ballantyne, C.K., Wilson, P., Gheorghiu, D., Rodes, A` ., 2014b. Enhanced rock-slope failure following ice-sheet deglaciation: timing and causes. Earth Surf. Process. Landf. 39 (7), 900–913.
Bessette-Kirton, E.K., Coe, J.A., Zhou, W., 2018. Using stereo satellite imagery to account for ablation, entrainment, and compaction in volume calculations for rock avalanches on glaciers: application to the 2016 Lamplugh rock avalanche in Glacier Bay National Park, Alaska. J. Geophys. Res. 123. https://doi.org/10.1002/2017JF004512.
Blair, R.W., 1994. Moraine and valley wall collapse due to rapid deglaciation in Mount Cook National Park, New Zealand. Mt. Res. Dev. 14, 347–358.
Blikra, L.H., Christiansen, H.H., 2014. A field-based model of permafrost-controlled rockslide deformation in northern Norway. Geomorphology 208, 34–39. https://doi.org/10.1016/j.geomorph.2013.11.014.
Bodin, X., Thibert, E., Fabre, D., Ribolini, A., Schoeneich, P., Francou, B., Reynaud, L., Fort, M., 2009. Two de- cades of responses (1986–2006) to climate by the Laurichard rock glacier, French Alps. Permafr. Periglac. Process. 20, 331–344.
Bodin, X., Krysiecki, J.-M., Schoeneich, P., Le Roux, O., Lorier, L., Echelard, T., Peyron, M., Walpersdorf, A., 2017. The 2006 collapse of the Berard Rock Glacier (Southern French Alps). Permafr. Periglac. Process. 28, 209–223. https://doi.org/10.1002/ppp.1887.
Bodin, X., Thibert, E., Sanchez, O., Rabatel, A., Jaillet, S., 2018. Multi-annual kinematics of an active rock glacier quantified from very high-resolution DEMs: an application-case in the French Alps. Remote Sens. 10, 547. https://doi.org/10.3390/rs10040547.
Boeckli, L., Brenning, A., Gruber, S., Noetzli, J., 2012. A statistical approach to modelling permafrost distribution in the European Alps or similar mountain ranges. Cryosphere 6, 125–140. https://doi.org/10.5194/tc-6-125-2012. Bommer, C., Phillips, M., Arenson, L.U., 2010. Practical recommendations for planning, constructing and maintaining infrastructure in mountain permafrost. Permafr. Periglac. Process. 21, 97–104.
Bornhold, B.D., Harper, J.R., McLaren, D., Thomson, R.E., 2007. Destruction of the First Nations village of Kwa-

late by a rock avalanche-generated tsunami. Atmosphere-Ocean 45, 123–128.
Buchli, T., Merz, K., Zhou, X., Kinzelbach, W., Springman, S., 2013. Characterization and monitoring of the Furggwanghorn rock glacier, Turtmann Valley, Switzerland: results from 2010 to 2012. Vadose Zone J. 12 (1). https://doi.org/10.2136/vzj2012.0067.
Budd, W.F., Jacka, T.H., 1989. A review of ice rheology for ice-sheet modeling. Cold Reg. Sci. Technol. 16, 107–144.
Chiarle, M., Mortara, G., 2001. Esempi di rimodellamento di apparati morenici nell'arco alpino italiano. Suppl. Geogr. Fis. Dinam. Quat., 41–54.
Clague, J.J., Evans, S.G., 2000. A review of catastrophic drainage of moraine-dammed lakes in British Columbia. Quat. Sci. Rev. 19, 1763–1783.
Coe, J.A., Bessette-Kirton, E.K., Geertsema, M., 2018. Increasing rock-avalanche size and mobility in Glacier Bay National Park and Preserve, Alaska detected from 1984–2016 Landsat imagery. Landslides 15, 393–407. https://doi.org/10.1007/s10346-017-0879-7.
Coperey, A., Revil, A., Abdulsamad, F., Stutz, B., Duvillard, P.-A., Ravanel, L., 2019. Low-frequency induced polarization of porous media undergoing freezing: preliminary observations and modeling. J. Geophys. Res. 124. https://doi.org/10.1029/2018JB017015.
Cossart, E., Braucher, R., Fort, M., Bourles, D.L., Carcaillet, J., 2008. Slope instability in relation to glacial debuttressing in alpine areas (Upper Durance catchment, southeastern France): evidence from field data and $_{10}$Be cosmic ray exposure ages. Geomorphology 95, 3–26.
Cossart, E., Mercier, D., Decaulne, A., Feuillet, T., Jonsson, H.A., Saemundsson, P., 2013. Impacts of post-glacial rebound on landslide spatial distribution at a regional scale in Northern Iceland (Skagafj€orður). Earth Surf. Process. Landf. 39, 336–350. https://doi.org/10.1002/esp.3450.
Crosta, G.B., di Prisco, C., Frattini, P., Frigerio, G., Castellanza, R., Agliardi, F., 2013. Chasing a complete understanding of the triggering mechanisms of a large rapidly evolving rockslide. Landslides 11 (5), 747–764. https://doi.org/10.1007/s10346-013-0433-1.
Cruden, D.M., Varnes, D.J., 1996. Landslide types and processes. In: Turner, A.K., Schuster, R.L. (Eds.), Special Report 247: Landslides Investigation, Mitigation. National Research Council, Transportation Research Board, Washington, DC, pp. 36–75.

Curry, A.M., 1999. Paraglacial modification of slope form. Earth Surf. Process. Landf. 24, 1213–1228.
Curry, A.M., Cleasby, V., Zukowskyj, P., 2006. Paraglacial response of steep, sediment-mantled slopes to post-'Little Ice Age' glacier recession in the central Swiss Alps. J. Quat. Sci. 21, 211–225.
Darrow, M.M., Gyswyt, N.L., Simpson, J.M., Daanen, R.P., Hubbard, T.D., 2016. Frozen debris lobe morphology and movement: an overview of eight dynamic features, southern Brooks Range, Alaska. Cryosphere 10, 977–993. https://doi.org/10.5194/tc-10-977-2016.
Delaloye, R., Lambiel, C., 2005. Evidence of winter ascending air circulation throughout talus slopes and rock glaciers situated in the lower belt of alpine discontinuous permafrost (Swiss Alps). Nor. J. Geogr. 59 (2), 194–203.
Delaloye, R., Morard, S., 2011. Le glacier rocheux destabilize du Petit-Velan (Val d'Entremont, Valais): morpho-

logie de surface, vitesses de deplacement et structure interne. In: Lambiel, C., Reynard, E., Scapozza, C. (Eds.), La geomorphologie alpine: entre patrimoine et contrainte. Actes du colloque de la Societe Suisse de Geomorphologie, Olivone, Geovisions. vol. 36. IGUL, Lausanne, pp. 195–210.

Delaloye, R., Perruchoud, E., Avian, M., Kaufmann, V., Bodin, X., Hausmann, H., Ikeda, A., K€a€ab, A., Kellerer-Pirklbauer, A., Krainer, K., Lambiel, C., Mihajlovic, D., Staub, B., Roer, I., Thibert, E., 2008. Recent inter- annual variations of rock glacier creep in the European Alps. In: Kane, D.L., Hinkel, K.M. (Eds.), Proceeding of the 9th International Conference on Permafrost. University of Alaska, Fairbanks, pp. 343–348.

Delaloye, R., Morard, S., Barboux, C., Abbet, D., Gruber, V., Riedo, M., Gachet, S., 2013. Rapidly moving rock glaciers in Mattertal. In: Jahrestagung der Schweizerischen Geomorphologischen Gesellschaft, pp. 21–31.

Deline, P., Hewitt, K., Reznichenko, N., Shugar, D., 2015. Chapter 9. Rock avalanches onto glaciers (Volume Ed.). In: Davies, T.R.H. (Ed.), Landslide Hazards, Risks and Disasters. Elsevier, Amsterdam, ISBN: 978-0-12- 396452-6, pp. 263–319. in: Shroder, J. (Series Ed.), Hazards and Disasters Series.

Draebing, D., Eichel, J., 2018. Divergence, convergence, and path dependency of paraglacial adjustment of alpine lateral moraine slopes. Land Degrad. Dev. 29 (6), 1979–1990. https://doi.org/10.1002/ldr.2983.

Draebing, D., Krautblatter, M., 2012. P-wave velocity changes in freezing hard low-porosity rocks: a laboratory- based time-average model. Cryosphere 6, 1163–1174. https://doi.org/10.5194/tc-6-1163-2012.

Dufresne, A., Geertsema, M., Shugar, D., Koppes, M., Higman, B., Haessler, P., Stark, C., Venditti, J., Bonno, D., Larsen, C., Gulick, S., McCall, N., Walton, M., Loso, M., Willis, M., 2018. Sedimentology and geomorphol- ogy of a large tsunamigenic landslide, Taan Fiord, Alaska. Sediment. Geol. 364, 302–318.

Duvillard, P.-A., 2019. Destabilisation des terrains-supports d'infrastructures de haute montagne en contexte de rechauffement climatique dans les Alpes franc ̧aises. PhD Thesis, Chambery, France, Universite Grenoble Alpes.

Dwivedi, R.D., Soni, A.K., Goel, R.K., Dube, A.K., 2000. Fracture toughness of rocks under sub-zero temperature conditions. Int. J. Rock Mech. Min. Sci. 37, 1267–1275.

Dykes, R., Brook, M., Roberston, C., Fuller, I., 2011. Twenty-first century calving retreat of Tasman Glacier, New Zealand. Arct. Antarct. Alp. Res. 43, 1–10.

Eichel, J., Draebing, D., Meyer, N., 2018. From active to stable: paraglacial transition of Alpine lateral moraine slopes. Land Degrad. Dev. 29 (11), 4158–4172. https://doi.org/10.1002/ldr.3140.

Eriksen, H.Ø., Rouyet, L., Lauknes, T.R., Berthling, I., Isaksen, K., Hindberg, H., Larsen, Y., Corner, G.D., 2018. Recent acceleration of a rock glacier complex, Adjet, Norway, documented by 62 years of remote sensing observations. Geophys. Res. Lett. 45, 8314–8323. https://doi.org/10.1029/2018GL077605.

Evans, S.G., Clague, J.J., 1988. Catastrophic rock avalanches in glacial environments. In: Bonnard, C. (Ed.), Pro- ceedings of 5th International Symposium on Landslides. vol. 2. Balkema, Rotterdam, pp. 1153–1158.

Evans, S.G., Clague, J.J., 1994. Recent climatic change and catastrophic geomorphic processes in mountain en- vironments. Geomorphology 10 (1–4), 107–128.

Evans, S.G., Delaney, K.B., Rana, N.M., 2021. The occurrence and mechanism of catastrophic mass flows in the mountain cryosphere. In: Haeberli, W., Whiteman, C. (Eds.), Snow and Ice-related Hazards, Risks and Disas- ters. Elsevier. pp. 541–596.

Evans, S.G., Tutubalina, O.V., Drobyshev, V.N., Chernomorets, S.S., McDougall, S., Petrakov, D.A., Hungr, O., 2009. Catastrophic detachment and high-velocity long-runout flow of Kolka Glacier, Caucasus Mountains, Russia in 2002. Geomorphology 105, 314–321.

Faillettaz, J., Sornette, D., Funk, M., 2011. Numerical modeling of a gravity-driven instability of a cold hanging glacier: reanalysis of the 1895 break-off of Altelsgletscher, Switzerland. J. Glaciol. 57 (205), 817–831. https://doi.org/10.3189/002214311798043852.

Faillettaz, J., Funk, M., Sornette, D., 2012. Instabilities on Alpine temperate glaciers: new insights arising from the numerical modelling of Allalingletscher (Valais, Switzerland). Nat. Hazards Earth Syst. Sci. 12, 2977–2991.

Faillettaz, J., Funk, M., Vincent, C., 2015. Avalanching glacier instabilities: review on processes and early warning perspectives. Rev. Geophys. 53, 203–224. https://doi.org/10.1002/2014RG000466.

Faillettaz, J., Funk, M., Vagliasindi, M., 2016. Time forecast of a break-off event from a hanging glacier. Cryo- sphere 10, 1191–1200. https://doi.org/10.5194/tc-10-1191-2016.

Fischer, L., Huggel, C., 2008. Methodical design for stability assessments of permafrost-affected high-mountain rock walls. In: Kane, D.L., Hinkel, K.M. (Eds.), Proceeding of the 9th International Conference on Permafrost. University of Alaska, Fairbanks, pp. 439–444.

Fischer, L., K€a€ab, A., Huggel, C., Noetzli, J., 2006. Geology, glacier retreat and permafrost degradation as con- trolling factor of slope instabilities in a high-mountain rock wall: the Monte Rosa east face. Nat. Hazards Earth Syst. Sci. 6, 761–772. https://doi.org/10.5194/nhess-6-761-2006.

Fischer, L., Amann, F., Moore, J., Huggel, C., 2010. Assessment of periglacial slope stability for the 1988 Tschierva rock avalanche (Piz Morteratsch, Switzerland). Eng. Geol. 116, 32–43. https://doi.org/10.1016/j. enggeo.2010.07.005.

Fischer, L., Eisenbeiss, H., K€a€ab, A., Huggel, C., Haeberli, W., 2011. Monitoring topographic changes in a peri- glacial high-mountain face using high-resolution DTMs, Monte Rosa East Face, Italian Alps. Permafr. Peri- glac. Process. 22, 140–152. https://doi.org/10.1002/ppp.717.

Fischer, L., K€a€ab, A., Huggel, C., Haeberli, W., 2013. Slope failures and erosion rates on a glacierized high- mountain face under climatic changes. Earth Surf. Process. Landf. 38 (8), 836–846. https://doi.org/ 10.1002/esp.3355.

Frattini, P., Crosta, G., Valagussa, A., Tamburini, A., Alberto, W., 2015. Rockfall risk analysis along transportation facilities (Volume Eds.). In: Lollino, G., Giordan, D., Crosta, G., Corominas, J., Azzam, R., Wasowski, J., Sciarra, N. (Eds.), Landslide Processes. Springer, pp. 2061–2064. In Lollino, G. (Series Ed.) Engineering Ge- ology for Society and Territory, Proceedings of the IAEG 12th Congress, Torino, September 2014.

Fritz, H.M., Mohammed, F., Yoo, J., 2009. Lituya Bay landslide impact generated mega-tsunami 50th anniversary. Pure Appl. Geophys. 166, 153–175. https://doi.org/10.1007/s00024-008-0435-4.

Geertsema, M., 2012. Initial observations of the 11 June 2012 rock/ice avalanche, Lituya Mountain, Alaska. In: The First Meeting of ICL Cold Region Landslides Network, Harbin, pp. 49–54.

Geertsema, M., Chiarle, M., 2013. Mass-movement causes: glacier thinning. In: Shroder, J.F., Marston, R.A., Stof- fel, M. (Eds.), Treatise on Geomorphology, volulme 7, Mountain and Hillslope Geomorphology. Academic Press, San Diego, pp. 217–222.

Geertsema, M., Clague, J.J., Schwab, J.W., Evans, S.G., 2006. An overview of recent large landslides in northern British Columbia, Canada. Eng. Geol. 83, 120–143.

Girard, L., Gruber, S., Weber, S., Beutel, J., 2013. Environmental controls of frost cracking revealed through in situ acoustic emission measurements in steep bedrock. Geophys. Res. Lett. 40 (9), 1748–1753. https://doi.org/ 10.1002/grl.50384.

Gischig, V.S., Moore, J.R., Evans, K.F., Amann, F., Loew, S., 2011. Thermomechanical forcing of deep rock slope deformation: 1. Conceptual study of a simplified slope. J. Geophys. Res. 116 (F4), F04010.

Gr€amiger, L.M., Moore, J.R., Gischig, V.S., Ivy-Ochs, S., Loew, S., 2017. Beyond debuttressing: mechanics of paraglacial rock slope damage during repeat glacial cycles. J. Geophys. Res. 122, 1004–1036.

Gr€amiger, L.M., Moore, J.R., Gischig, V.S., Loew, S., 2018. Thermomechanical stresses drive damage of Alpine valley rock walls during repeat glacial cycles. J. Geophys. Res. 123, 2620–2646.

Gruber, S., Haeberli, W., 2007. Permafrost in steep bedrock slopes and its temperature-related destabilization fol- lowing climate change. J. Geophys. Res. 112, F02S18. https://doi.org/10.1029/2006JF000547.

Gruber, S., Haeberli, W., 2009. Mountain permafrost. In: Margesin, R. (Ed.), Permafrost Soils, Biology Series. vol. 16. Springer, pp. 33–44, https://doi.org/10.1007/978-3-540-69371-0_3.

Gruber, S., Hoelzle, M., Haeberli, W., 2004. Permafrost thaw and destabilization of Alpine rock walls in the hot summer of 2003. Geophys. Res. Lett. 31, L13504. https://doi.org/10.1029/2004GL020051.

Haberkorn, A., Wever, N., Hoelzle, M., Phillips, M., Kenner, R., Bavay, M., Lehning, M., 2017. Distributed snow and rock temperature modelling in steep rock walls using Alpine3D. Cryosphere 11, 585–607. https://doi.org/ 10.5194/tc-11-585-2017.

Haeberli, W., Funk, M., 1991. Borehole temperatures at the Colle Gnifetti core-drilling site (Monte Rosa, Swiss Alps). J. Glaciol. 37 (125), 37–46.

Haeberli, W., Gruber, S., 2008. Research challenges for steep and cold terrain. In: Kane, D.L., Hinkel, K.M. (Eds.), Proceeding of the 9th International Conference on Permafrost, Fairbanks, pp. 597–605.

Haeberli, W., Wegmann, M., Vonder M€uhll, D., 1997. Slope stability problems related to glacier shrinkage and permafrost degradation in the Alps. Eclogae Geol. Helv. 90, 407–414.

Haeberli, W., K€a€ab, A., Hoelzle, M., B€osch, H., Funk, M., Vonder M€uhll, D., Keller, F., 1999. Eisschwund und Naturkatastrophen im Hochgebirge. vdf Hochschulverlag an der ETH Z€urich, Z€urich, p. 190.

Haeberli, W., K€a€ab, A., Vonder M€uhll, D., Teysseire, P., 2001. Prevention of outburst floods from periglacial lakes at Grubengletscher, Valais, Swiss Alps. J. Glaciol. 47 (156), 111–122.

Haeberli, W., K€a€ab, 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. Nor. Geogr. Tidsskr.—Nor. J. Geogr. 56, 104–111.

Haeberli, W., Huggel, C., K€a€ab, A., Zgraggen-Oswald, S., Polkvoj, A., Galushkin, I., 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. J. Glaciol. 50, 533–546.

Haeberli, W., Buetler, M., Huggel, C., Lehmann Friedli, T., Schaub, Y., Schleiss, A.J., 2016. New lakes in degla- ciating high-mountain regions—opportunities and risks. Clim. Chang. 139 (2), 201–214. https://doi.org/ 10.1007/s10584-016-1771-5.

Haeberli, W., Schaub, Y., Huggel, C., 2017. Increasing risks related to landslides from degrading permafrost into new lakes in de-glaciating mountain ranges. Geomorphology 293, 405–417. https://doi.org/10.1016/j. geomorph.2016.02.009.

Haeussler, P.J., Gulick, S.P.S., McCall, N., Walton, M., Reece, R., Larsen, C., Shugar, D.H., Geertsema, M., Ven- ditti, J.G., Labay, K., 2018. Submarine deposition of a subaerial landslide in Taan Fiord, Alaska. J. Geophys. Res. 123 (10), 2443–2463.

Hampel, A., Hetzel, R., Maniatis, G., 2010. Response of faults to climate-driven changes in ice and water volumes on Earth's surface. Philos. Trans. R. Soc. A Math. Phys. Eng. Sci. 368 (1919), 2501–2517.

Hancox, G.T., 1994. Report on Mt Cook National Park Hut Site Inspections and Establishment of Monitoring Lines. Institute of Geological and Nuclear Sciences. Client Report 353902.20, prepared for Department of Conservation, Mt Cook National Park.

Hartl, L., Fischer, A., Stocker-Waldhuber, M., Abermann, J., 2016. Recent speed-up of an alpine rock glacier: an updated chronology of the kinematics of Outer Hochebenkar rock glacier based on geodetic measurements. Geogr. Ann. Ser. A 98 (2), 129–141. https://doi.org/10.1111/geoa.12127.

Hasler, A., Gruber, S., Font, M., Dubois, A., 2011a. Advective heat transport in frozen rock clefts—conceptual model, laboratory experiments and numerical simulation. Permafr. Periglac. Process. 22, 387–398. https://doi.org/10.1002/ppp.737.

Hasler, A., Gruber, S., Haeberli, W., 2011b. Temperature variability and offset in steep alpine rock and ice faces. Cryosphere 5, 977–988. https://doi.org/10.5194/tc-5-977-2011.

Hasler, A., Gruber, S., Beutel, J., 2012. Kinematics in steep bedrock permafrost. J. Geophys. Res. 117, F01016. https://doi.org/10.1029/2011JF001981.

Hauck, C., 2013. New concepts in geophysical surveying and data interpretation for permafrost terrain. Permafr. Periglac. Process. 24, 131–137. https://doi.org/10.1002/ppp.1774.

Higman, B., Shugar, D.H., Stark, C.P., Ekstr€om, G., Koppes, M.N., Lynett, P., Dufresne, A., Haeussler, P.J., Geertsema, M., Gulick, S., Mattox, A., 2018. The 2015 landslide and tsunami in Taan Fiord, Alaska. Sci. Rep. 8, 12993.

Hock, R., Rasul, G., Adler, C., Ca´ceres, B., Gruber, S., Hirabayashi, Y., Jackson, M., K€a€ab, A., Kang, S., Kutuzov, S., Milner, A., Molau, U., Morin, S., Orlove, B., Steltzer, H., 2019. Chapter 2: High mountain areas. In: P€ortner, H.-O., Roberts, D.C., Masson-Delmotte, V., Zhai, P., Tignor, M., Poloczanska, E., Weyer, N. (Eds.), IPCC Special Report on the Ocean and Cryosphere in a Changing Climate, pp. 2.1–2.94.

Hoelzle, M., Darms, G., L€uthi, M.P., Suter, S., 2011. Evidence of accelerated englacial warming in the Monte Rosa area, Switzerland/Italy. Cryosphere 5, 231–243. https://doi.org/10.5194/tc-5-231-2011.

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

Hugenholtz, C., Moorman, B., Barlow, J., Wainstein, P., 2008. Large-scale moraine deformation at the Athabasca Glacier, Jasper National Park, Alberta, Canada. Landslides 5 (3), 251–260.

Ikeda, A., Matsuoka, N., 2006. Pebbly versus bouldery rock glaciers: morphology, structure and processes. Geomorphology 73, 279–296.

IPCC, 2019. IPCC Special Report on the Ocean and Cryosphere in a Changing Climate [H.-O. Po¨rtner, D.C. Roberts, V. Masson-Delmotte, P. Zhai, M. Tignor, E. Poloczanska, K. Mintenbeck, A. Alegŕ ıa, M. Nicolai, A. Okem, J. Petzold, B. Rama, N.M. Weyer (eds.)]. In press.

Jarman, D., Harrison, S., 2019. Rock slope failure in the British Mountains. Geomorphology 340, 202–233. https://doi.org/10.1016/j.geomorph.2019.03.002.

Jia, H., Xiang, W., Krautblatter, M., 2015. Quantifying rock fatigue and decreasing compressive and tensile strength after repeated freeze-thaw cycles. Permafr. Periglac. Process. 26, 368–377. https://doi.org/10.1002/ ppp.1857.

K€a€ab, A., Huggel, C., Barbero, S., Chiarle, M., Cordola, M., Epifani, F., Haeberli, W., Mortara, G., Semino, P., Tamburini, A., Viazzo, G., 2004. Glacier hazards at Belvedere Glacier and the Monte Rosa east face, Italian Alps: processes and mitigation. In: Proceedings of the Interpra event 2004, Riva/Trient, pp. 67–78.

K€a€ab, A., Frauenfelder, R., Roer, I., 2007. On the response of rock glacier creep to surface temperature increase. Glob. Planet. Chang. 56 (1-2), 172–187.

Kemeny, J., 2003. The time-dependent reduction of sliding cohesion due to rock bridges along discontinuities: a fracture mechanics approach. Rock Mech. Rock. Eng. 36, 27–38. https://doi.org/10.1007/s00603-002-0032-2. Kirkbride, M.P., Warren, C.R., 1999. Tasman Glacier, New Zealand: twentieth-century thinning and predicted

calving retreat. Glob. Planet. Chang. 22 (1-4), 11–28.
Kleinberg, R.L., Griffin, D.D., 2005. NMR measurements of permafrost: unfrozen water assay, pore-scale distri-

bution of ice, and hydraulic permeability of sediments. Cold Reg. Sci. Technol. 42, 63–77.
Koppes, M., Hallet, B., 2006. Erosion rates during rapid deglaciation in Icy Bay, Alaska. J. Geophys. Res. Earth

Surf. 111, F02023. https://doi.org/10.1029/2005JF000349.
Kramer, T., Stadler, H., 2007. Eduard Spelterini. Photographs of a Pioneer Balloonist. Scheidegger & Spiess,

Z€urich, p. 147.
Krautblatter, M., 2009. Detection and Quantification of Permafrost Change in Alpine Rock Walls and Implications

for Rock Instability (Ph.D. Thesis). University of Bonn, Bonn, Germany.

Krautblatter, M., Draebing, D., 2014. Pseudo 3D—P-wave refraction seismic monitoring of permafrost in steep unstable bedrock. J. Geophys. Res. 119, 287–299. https://doi.org/10.1002/2012JF002638.

Krautblatter, M., Hauck, C., 2007. Electrical resistivity tomography monitoring of permafrost in solid rock walls. J. Geophys. Res. 112, F02S20. https://doi.org/10.1029/2006JF000546.

Krautblatter, M., Verleysdonk, S., Flores-Orozco, A., Kemna, A., 2010. Temperature-calibrated imaging of sea- sonal changes in permafrost rock walls by quantitative electrical resistivity tomography (Zugspitze, German/ Austrian Alps). J. Geophys. Res. 115, F02003.

Krautblatter, M., Funk, D., G€unzel, F.K., 2013. Why permafrost rocks become unstable: a rock-ice-mechanical model in time and space. Earth Surf. Process. Landf. 38 (8), 876–887. https://doi.org/10.1002/esp.3374. Kummert, M., Delaloye, R., 2018. Mapping and quantifying sediment transfer between the front of rapidly moving

rock glaciers and torrential gullies. Geomorphology 309, 60–76. https://doi.org/10.1016/j.

geomorph.2018.02.021.
Kummert, M., Delaloye, R., Braillard, L., 2018. Erosion and sediment transfer processes at the front of rapidly

moving rock glaciers: systematic observations with automatic cameras in the western Swiss Alps. Permafr.

Periglac. Process. 29, 21–33. https://doi.org/10.1002/ppp.1960.
Lambiel, C., 2011. Le glacier rocheux destabilise de Tsate-Moiry (VS): caracteristiques mor- phologiques et

vitesses de deplacement. In: Lambiel, C., Reynard, E., Scapozza, C. (Eds.), La geomorphologie alpine: entre patrimoine et contrainte. Actes du colloque de la Societe Suisse de Geomorphologie, Olivone, Geovisions, 36. IGUL, Lausanne, pp. 211–224.

Lillie, A.R., Gunn, B.M., 1964. Steeply plunging folds in the Sealy Range, southern Alps. N.Z.J. Geol. Geophys. 7, 403–423.

Luethi, R., Gruber, S., Ravanel, L., 2015. Modelling transient ground surface temperatures of past rockfall events: towards a better understanding of failure mechanisms in changing periglacial environments. Geogr. Ann. Ser. B 97 (4), 753–767. https://doi.org/10.1111/geoa.12114.

Lugon, R., Gardaz, J.-M., Vonder M€uhll, D., 2000. The partial collapse of the Dolent glacier moraine (Mont Blanc range, Swiss Alps). Z. Geomorphol. Supplement-Band 122, 191–208.

Magnin, F., Krautblatter, M., Deline, P., Ravanel, L., Malet, E., Bevington, A., 2015a. Determination of warm, sensitive permafrost areas in near-vertical rockwalls and evaluation of distributed models by electrical resis- tivity tomography. J. Geophys. Res. 120, 745–762. https://doi.org/10.1002/2014JF003351.

Magnin, F., Brenning, A., Bodin, X., Deline, P., Ravanel, L., 2015b. Statistical modelling of rock wall permafrost distribution: application to the Mont Blanc massif. Geomorphologie 21, 145–162. https://doi.org/10.4000/geomorphologie.10965.

Magnin, F., Deline, P., Ravanel, L., Noetzli, J., Pogliotti, P., 2015c. Thermal characteristics of permafrost in the steep alpine rock walls of the Aiguille du Midi (Mont Blanc Massif, 3842 m a.s.l). Cryosphere 9, 109–121. https://doi.org/10.5194/tc-9-109-2015.

Magnin, F., Westermann, S., Pogliotti, P., Ravanel, L., Deline, P., Malet, E., 2017. Snow control on active layer thickness in steep alpine rock walls (Aiguille du Midi, 3842ma.s.l., Mont Blanc massif). Catena 149, 648–662. https://doi.org/10.1016/j.catena.2016.06.006.

Mamot, P., Weber, S., Schr€oder, T., Krautblatter, M., 2018. A temperature- and stress-controlled failure criterion for ice-filled permafrost rock joints. Cryosphere 12, 3333–3353. https://doi.org/10.5194/tc-12-3333-2018.

Marcer, M., Bodin, X., Schoeneich, P., 2016. Reconstructing the last century evolution of the Derochoir rock gla- cier using a multi-approach monitoring. In: Proceedings of the 11th International Conference on Permafrost, Potsdam, p. 953.

Marcer, M., Serrano, C., Brenning, A., Bodin, X., Goetz, J., Schoeneich, P., 2019. Evaluating the destabilization susceptibility of active rock glaciers in the French Alps. Cryosphere 13, 141–155. https://doi.org/10.5194/tc- 13-141-2019.

Marchenko, S.S., Gorbunov, A.P., Romanovsky, V.E., 2007. Permafrost warming in the Tien Shan mountains, Central Asia. Glob. Planet. Chang. 56, 311–327. https://doi.org/10.1016/j.gloplacha.2006.07.02.

Matsuoka, N., Murton, J., 2008. Frost weathering: recent advances and future directions. Permafr. Periglac. Pro- cess. 19 (2), 195–210.

Mattson, L.E., Gardner, J.S., 1991. Mass wasting on valley-side ice-cored moraines, Boundary Glacier, Alberta, Canada. Geogr. Ann. 73A, 123–128.

McColl, S.T., 2012. Paraglacial rock-slope stability. Geomorphology 153–154, 1–16.

McColl, S.T., Davies, T.R.H., 2012. Large ice-contact slope movements; glacial buttressing, deformation and ero- sion. Earth Surf. Process. Landf. 38 (10), 1102–1115. https://doi.org/10.1002/esp.3346.

Meigs, A., Krugh, W.C., Davis, K., Bank, G., 2006. Ultra-rapid landscape response and sediment yield following glacier retreat, Icy Bay, southern Alaska. Geomorphology 78, 207–221.

Mellor, M., 1973. Mechanical properties of rocks at low temperatures. In: 2nd International Conference on Per- mafrost, Yakutsk, pp. 334–344.

Mergili, M., Jaboyedoff, M., Pullarello, J., Pudasaini, S.P., 2019. Back-calculation of the 2017 Piz Cengalo-Bondo landslide cascade with r. avaflow. Nat. Hazards Earth Syst. Sci. https://doi.org/10.5194/nhess-2019-204.

Morard, S., Delaloye, R., Dorthe, J., 2008. Seasonal thermal regime of a mid-latitude ventilated debris accumu- lation. In: Kane, D.L., Hinkel, K.M. (Eds.), Proceeding of the 9th International Conference on Permafrost, Fairbanks, pp. 1233–1238.

Mortara, G., Dutto, F., Godone, F., 1995. Effetti degli eventi alluvionali nell'ambiente proglaciale. La sovrainci- sione della morena del Ghiacciaio del Mulinet (Stura di Valgrande, Alpi Graie). Geogr. Fis. Din. Quat. 18,

295–304.
Mortara, G., Tamburini, A., Chiarle, M., Haeberli, W., Epifani, F., 2009. Le morene: sull'orlo del collasso. In: Mortara, G., Tamburini, A. (Eds.), Il ghiacciao del Belvedere e l'emergenza del lago Effimero. Regione Piemonte e Edizioni Societa` Meteorologica Subalpina, pp. 123–134.
Mougin, P., 1914. Les torrents de la Savoie. Imprimerie Generale, Grenoble, p. 1251.
Murton, J.B., Peterson, R., Ozouf, J.-C., 2006. Bedrock fracture by ice segregation in cold regions. Science 314, 1127–1129. https://doi.org/10.1126/science.1132127.
Murton, J., Kuras, O., Krautblatter, M., Cane, T., Tschofen, D., Uhlemann, S., Schober, S., Watson, P., 2016. Monitoring rock freezing and thawing by novel geoelectrical and acoustic techniques. J. Geophys. Res. 121, 2309–2332. https://doi.org/10.1002/2016JF003948.
Nichols Jr., T.C., 1980. Rebound: its nature and effect on engineering works. Q. J. Eng. Geol. Hydrogeol. 13 (3), 133–152.
Noetzli, J., Gruber, S., 2009. Transient thermal effects in Alpine permafrost. Cryosphere 3 (1), 85–99. https://doi.org/10.5194/tc-3-85-2009.
Noetzli, J., Hoelzle, M., Haeberli, W., 2003. Mountain permafrost and recent Alpine rock-fall events: a GIS-based approach to determine critical factors. In: Proceedings of the 8th International Conference on Permafrost, Z€urich, pp. 827–832.
Oppikofer, T., Jaboyedoff, M., Keusen, H.-R., 2008. Collapse at the eastern Eiger flank in the Swiss Alps. Nat. Geosci. 1, 531–535. https://doi.org/10.1038/ngeo258.
PERMOS, 2013. Permafrost in Switzerland 2008/2009 and 2009/2010. In: Noetzli, J. (Ed.), Glaciological Report (Permafrost) No. 10/11 of the Cryospheric Commission of the Swiss Academy of Sciences.
PERMOS, 2019. Permafrost in Switzerland 2014/2015 to 2017/2018. In: Noetzli, J., Pellet, C., Staub, B. (Eds.), Glaciological Report (Permafrost) No. 16/19 of the Cryospheric Commission of the Swiss Academy of Sci- ences, p. 104, https://doi.org/10.13093/permos-rep-2019-16-19.
Phillips, M., Haberkorn, A., Draebing, D., Krautblatter, M., Rhyner, H., Kenner, R., 2016. Seasonally intermittent water flow through deep fractures in an Alpine rock ridge: Gemsstock, Central Swiss Alps. Cold Reg. Sci. Technol. 125, 117–127. https://doi.org/10.1016/j.coldregions.2016.02.010.
Pogliotti, P., 2011. Influence of Snow Cover on MAGST Over Complex Morphologies in Mountain Permafrost Regions. PhD Thesis, Universita` degli Studi di Torino, Turin, Italy.
Pogrebiskiy, M.I., Chernyshev, S.N., 1977. Determination of the permeability of the frozen fissured rock massif in the vicinity of the Kolyma Hydroelectric Power Station. Cold Reg. Res. Eng. Lab.—Draft Translat. 634, 1–13.

Porter, S.C., Orombelli, G., 1982. Late-glacial ice advances in the western Italian Alps. Boreas 11, 125–140. Pralong, A., 2005. On the instability of hanging glaciers. In: Mitteilungen, Laboratory of Hydraulics Hydrology and Glaciology 189. Switzerland, ETH Z€urich, p. 156.
Pralong, A., Funk, M., 2006. On the instability of avalanching glaciers. J. Glaciol. 52 (176), 31–48.
Ravanel, L., Deline, P., 2008. La face ouest des Drus (massif du Mont-Blanc): evolution de linstabilite d'une paroi rocheuse dans la haute montagne alpine depuis la fin du Petit Age Glaciaire. Geomorphologie 4, 261–272. 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. Ravanel, L., Allignol, F., Deline, P., Gruber, S., Ravello, M., 2010. Rock falls in the Mont Blanc massif in 2007 and

2008. Landslides 7, 493–501.
Ravanel, L., Allignol, F., Deline, P., Bruno, G., 2011. Les ecroulements rocheux dans le massif du Mont-Blanc

pendant l'ete caniculaire de 2003. In: Lambiel, C., Reynard, E., Scapozza, C. (Eds.), La geomorphologie alpine : entre patrimoine et contrainte. Actes du colloque de la Societe Suisse de Geomorphologie, Olivone, Geovisions, 36. IGUL, Lausanne, pp. 245–261.

Ravanel, L., Deline, P., Lambiel, C., Vincent, C., 2012. Instability of a highly vulnerable high alpine rock ridge: the lower Ar^ete des Cosmiques (Mont Blanc massif, France). Geogr. Ann. Ser. A 95 (1), 51–66. https://doi.org/10.1111/geoa.12000.

Ravanel, L., Magnin, F., Deline, P., 2017. Impacts of the 2003 and 2015 summer heatwaves on permafrost-affected rock-walls in the Mont Blanc massif. Sci. Total Environ. 609, 132–143. https://doi.org/10.1016/j. scitotenv.2017.07.055.

Ravanel, L., Duvillard, P.-A., Jaboyedoff, M., Lambiel, C., 2018. Recent evolution of an ice-cored moraine at the Gentianes Pass, Valais Alps, Switzerland. Land Degrad. Dev. 29 (10), 3693–3708. https://doi.org/10.1002/ldr. Roberti, G., Friele, P., de Vries, B.V.W., Ward, B., Clague, J.J., Perotti, L., Giardino, M., 2017. Rheological evolution of the Mount Meager 2010 debris avalanche, southwestern British Columbia. Geosphere 13 (2),

369–390.
Roberti, G., Ward, B., de Vries, B.V.W., Friele, P., Perotti, L., Clague, J.J., Giardino, M., 2018. Precursory slope

distress prior to the 2010 Mount Meager landslide, British Columbia. Landslides 15 (4), 637–647.
Roer, I., Haeberli, W., Avian, M., Kaufmann, V., Delaloye, R., Lambiel, C., K€a€ab, A., 2008. Observations and considerations on destabilizing active rockglaciers in the European Alps. In: Kane, D.L., Hinkel, K.M.

(Eds.), Proceeding of the 9th International Conference on Permafrost, Fairbanks, pp. 1505–1510. R€othlisberger, H., 1981. Eislawinen und Ausbr€uche von Gletscherseen. In: Kasser, P. (Ed.), Gletscher und Klima—glaciers et climat, Jahrbuch der Schweizerischen Naturforschenden, Gesellschaft, wissenschaftlicher

Teil 1978, pp. 170–212.
Sanchez, G., Rolland, Y., Corsini, M., Braucher, R., Bourle`s, D., Arnold, M., Aumaˆıtre, G., 2010. Relationships

between tectonics, slope instability and climate change: cosmic ray exposure dating of active faults, landslides

and glacial surfaces in the SW Alps. Geomorphology 117 (1–2), 1–13.
Sanderson, T., 1988. Ice Mechanics and Risks to Offshore Structures. Springer, Amsterdam, p. 270.
Schaub, Y., Haeberli, W., Huggel, C., K€unzler, M., Br€undl, M., 2013. Landslides and new lakes in deglaciating

areas: a risk management framework. In: Margottini, C., Canuti, P., Sassa, K. (Eds.), Landslide Science and Practice (Proceedings of the 2nd World Landslide Forum, Rome), Social and Economic Impacts and Policies. vol. 7. Springer, pp. 31–38, https://doi.org/10.1007/978-3-642-31313-4_5.

Scherler, M., Hauck, C., Hoelzle, M., Salzmann, N., 2013. Modeled sensitivity of two alpine permafrost sites to RCM-based climate scenarios. J. Geophys. Res. 118, 780–794. https://doi.org/10.1002/jgrf.20069.

Schoeneich, P., Bodin, X., Echelard, T., Kellerer-Pirklbauer, A., Krysiecki, J.-M., Lieb, G.K., 2014. Velocity changes of rock glaciers and induced hazards. In: Proceedings IAEG 12th Congress, 15–19 September 2014, Torino.

Schwab, J.W., Geertsema, M., Evans, S.G., 2003. Catastrophic rock avalanches, west-central B.C., Canada. In: 3rd Canadian Conference on Geotechnique and Natural Hazards, Edmonton, pp. 252–259.

Scotti, R., Crosta, G.B., Villa, A., 2016. Destabilisation of creeping permafrost: the Plator Rock Glacier Case Study (Central Italian Alps). Permafr. Periglac. Process. 28, 224–236. https://doi.org/10.1002/ppp.1917.

Staub, B., Lambiel, C., Delaloye, R., 2016. Rock glacier creep as a thermally-driven phenomenon: a decade of inter-annual observation from the Swiss Alps. In: Proceedings of the 11th International Conference on Perma- frost, Potsdam, pp. 96–97.

Stewart, I.S., Sauber, J., Rose, J., 2000. Glacio-seismotectonics: ice sheets, crustal deformation and seismicity. Quat. Sci. Rev. 19 (14–15), 1367–1389.

Stoffel, M., Huggel, C., 2012. Effects of climate change on mass movements in mountain environments. Prog. Phys. Geogr. 36 (3), 421–439. https://doi.org/10.1177/0309133312441010.

Tamburini, A., 2009. L'evoluzione della morena del Ghiacciao delle Locce. In: Mortara, G., Tamburini, A. (Eds.), Il Ghiacciao del Belvedere e l'emergenza del lago Effimero. Regione Piemonte e Edizioni Societa` Meteoro- logica Subalpina, pp. 140–144.

Tang, G.Z., Wang, X.H., 2006. Modeling the thaw boundary in broken rock zones in permafrost in the presence of surface water flows. Tunn. Undergr. Space Technol. 21, 684–689.

Terzaghi, K., 1962. Stability of steep slopes in hard unweathered rock. Geotechnique 12, 251–270. Voigtl€ander, A., Leith, K., Krautblatter, M., 2018. Subcritical crack growth and progressive failure in Carrara mar- ble under wet and dry conditions. J. Geophys. Res. 123, 3780–3798. https://doi.org/10.1029/2017JB014956. Wagner, S., 1996. Dreidimensionale Modellierung zweier Gletscher und Deformationsanalyse von eisreichem Permafrost. Mitteilungen, Laboratory of Hydraulics Hydrology and Glaciology 146, ETH Z€urich, Switzerland. Watanabe, K., Mizoguchi, M., 2002. Amount of unfrozen water in frozen porous media saturated with solution.

Cold Reg. Sci. Technol. 34 (2), 103–110. https://doi.org/10.1016/S0165-232X(01)00063-5.
Weber, S., F€ah, D., Beutel, J., Faillettaz, J., Gruber, S., Vieli, A., 2018. Ambient seismic vibrations in steep bedrock permafrost used to infer variations of ice-fill in fractures. Earth Planet. Sci. Lett. 501, 119–127. https://doi.org/

10.1016/j.epsl.2018.08.042.
Weber, S., Beutel, J., Da Forno, R., Geiger, A., Gruber, S., Gsell, T., Hasler, A., Keller, M., Lim, R., Limpach, P.,

Meyer, M., Talzi, I., Thiele, L., Tschudin, C., Vieli, A., Vonder M€uhll, D., Y€ucel, M., 2019. A decade of de- tailed observations (2008–2018) in steep bedrock permafrost at Matterhorn H€ornligrat (Zermatt, CH). Earth Syst. Sci. Data 11, 1203–1237. https://doi.org/10.5194/essd-11-1203-2019.

Wegmann, M., Gudmundsson, G.H., Haeberli, W., 1998. Permafrost changes in rock walls and the retreat of Al- pine glaciers: a thermal modelling approach. Permafr. Periglac. Process. 9, 23–33.

Wirsig, C., Zasadni, J., Christl, M., Akc ̧ar, N., Ivy-Ochs, S., 2016. Dating the onset of LGM ice surface lowering in the High Alps. Quat. Sci. Rev. 143, 37–50. https://doi.org/10.1016/j.quascirev.2016.05.001.

Zemp, M., Haeberli, W., Hoelzle, M., Paul, F., 2006. Alpine glaciers to disappear within decades? Geophys. Res. Lett. 33, L13504. https://doi.org/10.1029/2006GL026319.

Zemp, M., Frey, H., G€artner-Roer, I., Nussbaumer, S.U., Hoelzle, M., Paul, F., Haeberli, W., Denzinger, F., Ahl- stroem, A.P., Anderson, B., Bajracharya, S., Baroni, C., Braun, L.N., Caceres, B.E., Casassa, G., Cobos, G., Davila, L.R., Delgado Granados, H., Demuth, M.N., Espizua, L., Fischer, A., Fujita, K., Gadek, B., Ghazanfar, A., Hagen, J.O., Holmlund, P., Karimi, N., Li, Z., Pelto, M., Pitte, P., Popovnin, V.V., Portocarrero, C.A., Prinz, R., Sangewar, C.V., Severskiy, I., Sigurdsson, O., Soruco, A., Usubaliev, R., Vincent, C., 2015. Histor- ically unprecedented global glacier decline in the early 21st century. J. Glaciol. 61, 745–762. https://doi.org/ 10.3189/2015JoG15J017.

Zenklusen Mutter, E., Phillips, M., 2012a. Active layer characteristics at ten borehole sites in Alpine permafrost terrain, Switzerland. Permafr. Periglac. Process. 23, 138–151. https://doi.org/10.1002/ppp.1738.

Zenklusen Mutter, E., Phillips, M., 2012b. Thermal evidence of recent talik formation in Riti-graben rock glacier: Swiss Alps. In: Hinkel, K.M. (Ed.), Proceedings of the 10th International Conference on Permafrost, Sale- khard. The Northern Publisher, Salekhard, pp. 479–483.

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

Zischg, A., Curtaz, M., Galuppo, A., Lang, K., Mayr, V., Riedl, C., Schoeneich, P., 2011. Chapter 2. Permafrost and debris-flows. In: Schoeneich, P., Dall'Amico, M., Deline, P., Zischg, A. (Eds.), Hazards Related to Per- mafrost and to Permafrost Degradation., ISBN: 978-2-903095-59-8, pp. 29–66. PermaNET project, state-of- the-art report 6.2. Online publication.