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Permafrost in steep bedrock slopes and its temperature-related destabilization following climate change — Source link

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Published on: 01 Jun 2007 - Journal of Geophysical Research (John Wiley & Sons, Ltd)

Topics: Permafrost and Bedrock

Year: 2007

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Abstract

Permafrost in steep bedrock is abundant in many cold-mountain areas, and its degradation can cause slope instability that is unexpected and unprecedented in location, magnitude, frequency, and timing. These phenomena bear consequences for the understanding of landscape evolution, natural hazards, and the safe and sustainable operation of high-mountain infrastructure. Permafrost in steep bedrock is an emerging field of research. Knowledge of rock temperatures, ice content, mechanisms of degradation, and the processes that link warming and destabilization is often fragmental. In this article we provide a review and discussion of existing literature and pinpoint important questions. Ice-filled joints are common in bedrock permafrost and possibly actively widened by ice segregation. Broad evidence of destabilization by warming permafrost exists despite problems of attributing individual events to this phenomenon with certainty. Convex topography such as ridges, spurs, and peaks is often subject to faster and deeper thaw than other areas. Permafrost degradation in steep bedrock can be strongly affected by percolating water in fractures. This degradation by advection is difficult to predict and can lead to quick and deepdevelopment of thaw corridors along fractures in permafrost and potentially destabilize much greater volumes of rock than conduction would. Although most research on steep bedrock permafrost originates from the Alps, it will likely gain importance in other geographic regions with mountain permafrost.

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Permafrost in steep bedrock slopes and its temperature-related destabilization following climate change

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Received 30 April 2006; revised 12 February 2007; accepted 12 April 2007; published 8 June 2007.

[1] Permafrost in steep bedrock is abundant in many cold-mountain areas, and its degradation can cause slope instability that is unexpected and unprecedented in location, magnitude, frequency, and timing. These phenomena bear consequences for the understanding of landscape evolution, natural hazards, and the safe and sustainable operation of high-mountain infrastructure. Permafrost in steep bedrock is an emerging field of research. Knowledge of rock temperatures, ice content, mechanisms of degradation, and the processes that link warming and destabilization is often fragmental. In this article we provide a review and discussion of existing literature and pinpoint important questions. Ice-filled joints are common in bedrock permafrost and possibly actively widened by ice segregation. Broad evidence of destabilization by warming permafrost exists despite problems of attributing individual events to this phenomenon with certainty. Convex topography such as ridges, spurs, and peaks is often subject to faster and deeper thaw than other areas. Permafrost degradation in steep bedrock can be strongly affected by percolating water in fractures. This degradation by advection is difficult to predict and can lead to quick and deep development of thaw corridors along fractures in permafrost and potentially destabilize much greater volumes of rock than conduction would. Although most research on steep bedrock permafrost originates from the Alps, it will likely gain importance in other geographic regions with mountain permafrost.

Citation: Gruber, S., and W. Haeberli (2007), Permafrost in steep bedrock slopes and its temperature-related destabilization following climate change, *J. Geophys. Res.*, *112*, F02S18, doi:10.1029/2006JF000547.

1. Introduction

[2] Permafrost thaw is thought to be an important mechanism through which climate controls slope stability, landscape evolution and natural hazard potential in mountain areas. During the extremely hot and dry summer 2003 in the European Alps, many rockfall events that originated from permafrost in steep bedrock [*Schiermeier*, 2003; *Gruber et al.*, 2004b] were triggered without heavy precipitation or earthquake, and left massive ice visible in the exposed detachment zones. This supported earlier observations [*Dutto and Mortara*, 1991; *Dramis et al.*, 1995; *Deline*, 2001; *Haeberli et al.*, 2003; *Noetzli et al.*, 2003] and conceptual approaches [*Haeberli et al.*, 1997] indicating that permafrost and its climate-induced degradation influence the stability of steep bedrock permafrost. Although the geologic and geometric conditions are most important for stability, permafrost can be an important transient element that is subject to fast change. The fast warming of Alpine permafrost by 0.5 to 0.8C, in the upper decameters, during the last century has been confirmed by borehole measurements [*Harris et al.*, 2003].

[3] The aim of this article is to review and analyze existing knowledge on permafrost and ice in steep bedrock as well as the mechanisms that link warming and destabilization. On the basis of this, we point out some important future research.

[4] Steep bedrock is abundant in many cold-mountain areas (Figures 1 and 2) and contains a significant proportion of the permafrost. However, permafrost in steep slopes has previously received little attention: In the early twentieth century, the construction of mountain railways such as the Gornergrat (3100 m asl.) and the Jungfraujoch (3500 m asl.) took place in perennially frozen bedrock. Much of the subsequently published evidence on permafrost in steep bedrock is connected to and focused on construction [*Keusen and Haeberli*, 1983; *Keusen and Amiguet*, 1987; *King*, 1996; *Wegmann and Keusen*, 1998]. Potential stability problems of perennially frozen bedrock were outlined by *Haeberli et al.* [1997, 1999], followed by publications about measurement and modeling of permafrost temperatures in the Jungfraujoch/Aletsch area [*Wegmann et al.*, 1998; *Wegmann*, 1998]. These were the first comprehensive

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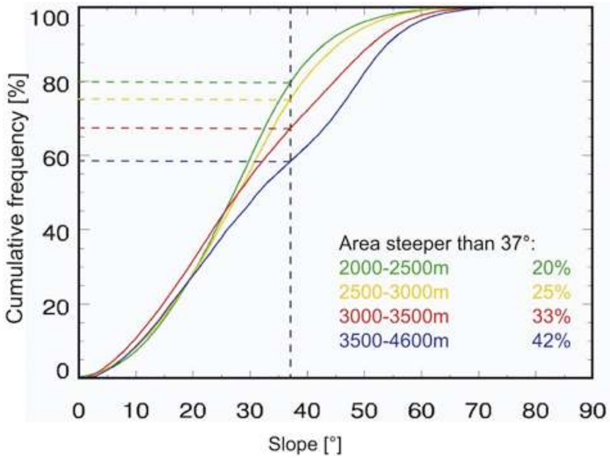

Figure 1. Cumulative histogram of slope steepness for different altitudinal zones in Switzerland. A slope steepness of 37 is assumed as a threshold for the separation of debris from bedrock slopes. Data source is Level 2 DHM25 of SwissTopo.

investigations of this topic. More recently, the combination of systematic measurements [*Gruber et al.*, 2003] and energy-balance modeling resulted in a more quantitative understanding of topographic and regional distribution as well as the temporal evolution of near-surface temperatures in steep Alpine bedrock [*Gruber et al.*, 2004a; *Gruber*, 2005].

[5] Observed [*Böhm et al.*, 2001; *Intergovernmental Panel on Climate Change* (*IPCC*), 2001; *Zemp et al.*, 2006] and projected [*IPCC*, 2001; *Knutti et al.*, 2002; *Stott and Kettleborough*, 2002; *Zwiers*, 2002] climate change is about to exceed the range of Holocene variability. Changes to the thermal regime of steep slopes may, therefore, lead to instabilities that are unexpected in location, magnitude, frequency, and timing. The high potential energy available for mass movements and the ability to trigger downslope cascades of hazardous events [*Huggel et al.*, 2004] increase the importance of steep bedrock permafrost. A considerable amount of infrastructure, especially in the European Alps, exists directly beneath or within steep bedrock permafrost and often makes a vital contribution to local economies. A better understanding of steep bedrock permafrost and the

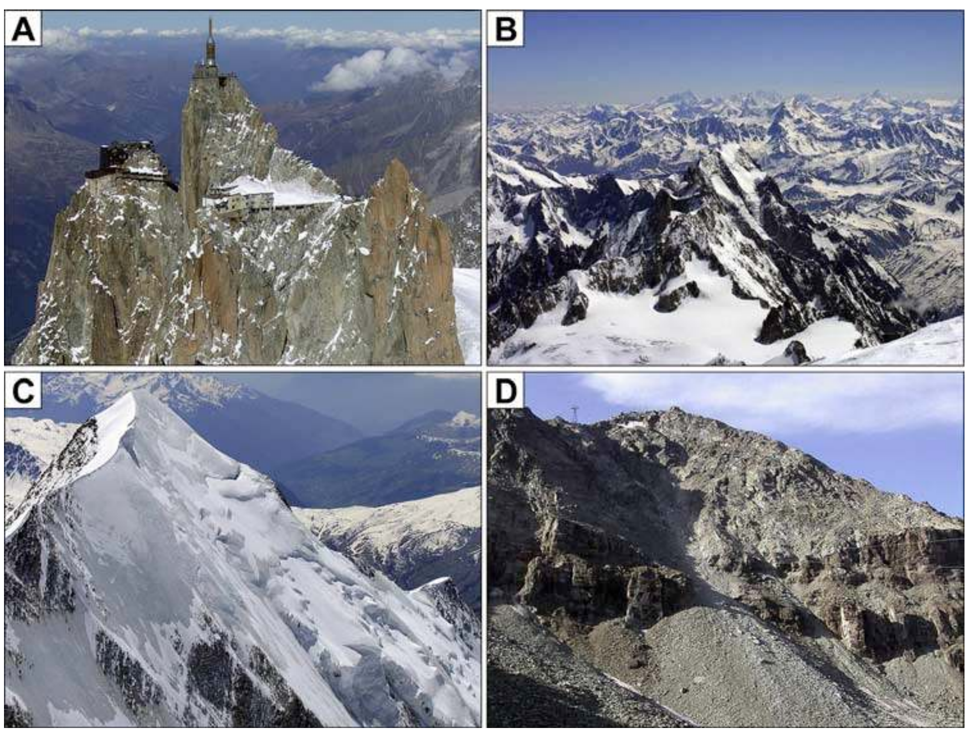

Figure 2. Examples of steep bedrock underlain by permafrost. (a) The Aiguille du Midi (granite) cable car top station at 3900 m asl near Chamonix, France, is one of many high-altitude buildings with considerable importance for local economies. (b) The Dent du Geant and the Grandes Jorasses (granite) are visible in the foreground, and the Alps of the Valais are visible in the background. The dominance of high and mostly snow-free steep slopes visually illustrates the importance of steep bedrock slopes within the zone of Alpine permafrost (photo: M. Hoelzle). (c) Ice faces and hanging glaciers on the surface of a steep bedrock slope (Aiguille de Bionnassay, granite) are a sure sign of permafrost. They constitute an additional surface element that can drastically influence temperature and water availability dependent on climatic conditions (photo: M. Hoelzle). (d) The degree of fracturing as well as the high spatial variability of surface conditions are illustrated by this slope (granodiorite) above the rock glacier Murtèl: Dark bands of steep rock with little fracturing as well as strongly fractured and debris-covered zones are visible in close proximity.

consequences of its thaw is very important for the safe, profitable, and sustainable operation of such infrastructure in the future. However, current understanding is limited, because knowledge about permafrost distribution, ground temperatures, surface characteristics and ice content in steep bedrock, and the mechanisms that link warming and destabilization is sparse.

2. Characteristics of Steep Bedrock Slopes

[6] Steep bedrock slopes do not accumulate thick covers of debris. On the basis of a typical angle of repose, an inclination greater than 37 is used as an approximate definition of "steep" slopes. In some instances (e.g., steep couloirs) thick debris can be present on steep slopes and, similarly, slopes less than 37 are sometimes free of debris.

[7] The lateral variability of microclimatic conditions and near-surface characteristics in steep slopes is high. Much of the variability in microclimatic conditions can be explained or modeled successfully with existing methods but near-surface characteristics and their influence on the mass and energy balance of steep bedrock are virtually unexplored [cf. *Gruber*, 2005]. The three most important dimensions of variation of surface conditions are (1) the degree of fracturing, (2) snow and ice cover, and (3) the availability of water. Vegetation is negligible in steep bedrock slopes that are subject to permafrost conditions.

[8] The degree of fracturing affects infiltration capacity, water content of the shallow subsurface, the ability to retain snow or clasts (surface roughness), and possible thermal offset mechanisms. It can be represented as a continuum bounded by the endmembers of (1) compact, unfractured rock and (2) a thick layer of blocks. Intermediate conditions include slightly fractured rock, deeply and heavily fractured rock, and fractured rock with individual detached clasts present at the surface (compare Figure 2d). The degree of fracturing is often inversely proportional to slope angle as it is related to stability.

[9] Snow and ice cover affects rock temperature and water availability. Snow cover on steep slopes is usually thin, laterally variable (Figure 2), and intermittent. The ability to retain snow is inversely proportional to the slope angle but rime or ice veneers can be accreted even on very steep slopes. High surface roughness and surface concavity favor the accumulation of snow. Rock temperature is important for the retention of snow because negative temperatures close to zero facilitate bonding by small amounts of liquid water and rapid metamorphosis. In very cold conditions, a high proportion of falling snow "flows" downslope. This effect frequently causes Sun-exposed rock faces to be snow-covered and shaded faces to be snow-free during winter. In summer the situation is reversed. Ice faces or hanging glaciers [cf. *Alean*, 1985; *Haeberli et al.*, 1997, 1999; *Lüthi and Funk*, 1997] (Figure 2c) are a sure sign of permafrost. Atmospheric warming will increase the proportion of temperate ice in hanging glaciers and affect the thermal and hydraulic conditions of the rock substrate.

[10] The availability of water affects advective transport of heat in fractured rock, rock weathering, and the turbulent exchange of latent energy at the surface. It can be approximated by the specific catchment area, i.e., the planimetric catchment area divided by the draining contour length. This favors concave over convex parts and areas toward the base of a wall over those near the crest. Melting snow and ice can contribute water during times without precipitation. Overhanging faces do not have a contributing area. Because of the rapid drainage of water in steep terrain, infiltration capacity is a crucial factor for the local effectiveness of water.

[11] The strong heterogeneity of near-surface characteristics is a major challenge in the interpretation of observations or measurements, as well as in the design, parametrization and validation of models.

3. Rock Temperature

[12] For near-vertical, unfractured rock, a basis of systematic temperature measurements and models exists [*Gruber et al.*, 2003, 2004a; *Gruber*, 2005]. It allows derivation of permafrost distribution patterns in steep rock, and simulation of temperature time series for various locations and depths. However, the effects of surface and subsurface characteristics, snow, thermal offset, and water circulation in gentler slopes remain largely unknown.

3.1. Snow and Thermal Offset

[13] Snow cover affects ground temperatures by increasing albedo, consuming melt energy, and insulating the ground surface from cold atmospheric conditions. Several researchers have investigated the effect of snow on permafrost in more gentle topography [*Keller and Gubler*, 1993; *Zhang et al.*, 1996, 2001; *Bernhard et al.*, 1998; *Ishikawa*, 2003; *Lütschg et al.*, 2003]. These investigations have to be interpreted with great care in the case of steep terrain. Snow cover in steep bedrock is usually shallow and intermittent, and its insulation effects are minimal. Steep bedrock often extends to elevations higher than the local equilibrium-line altitude, above which gentle slopes would be covered by glaciers. With increasing elevation the proportion of solid precipitation rises and snowfall can contribute to the cooling of steep slopes, even during summer. The cooling effect of snow diminishes with extreme steepness because of reduced snow cover. Low slope angles promote thick, insulating snow cover that also reduces the cooling effect.

[14] Mean annual temperatures at the top of permafrost are sometimes reduced significantly with respect to the surface. This "thermal offset" [*Burn and Smith*, 1988; *Romanovsky and Osterkamp*, 1995] has been described for Arctic soils and coarse blocky layers. It likely exists in steep bedrock as well, and can be caused by three different mechanisms: (1) Variable thermal conductivity of rock resulting from changes in the phase or saturation of pore water. Even in low-porosity rock this can reduce thermal conductivity significantly [cf. *Clauser and Huenges*, 1995]. (2) Fractures in the active layer are ice-filled during winter but drained and air-filled during summer, causing a contrast of thermal conductivity. (3) A loose cover of blocky clasts can cause a thermal offset effect similar to that described in more gentle terrain or on rock glaciers [*Harris*, 1996; *Harris and Pedersen*, 1998; *Goering*, 2003; *Hanson and Hoelzle*, 2004]. All three effects transmit cooling more readily than warming and lower the temperature at depth with respect to the surface.

[15] The concept of cooling by snow or thermal offset is supported by rock temperatures of −12C measured during

the construction of the Chli Matterhorn cable-car top station [*Keusen and Haeberli*, 1983]. Even with the assumption of a 20th century atmospheric warming of 1.0C [*Haeberli and Beniston*, 1998; *Böhm et al.*, 2001], this is still about 3C colder than model results [*Gruber et al.*, 2004a, Plate 1] indicate for steep north-facing slopes at the same location. By contrast, temperatures measured at depth in the East Ridge of Jungfrau by *Wegmann* [1998, Figures 7.16 and 7.17] coincide with the modeling results presented by *Gruber et al.* [2004a, Plate 1]. This linkage of temperatures at the surface and below the active layer in steep rock is currently weak because of a lack of measurements.

3.2. Temperatures at Depth

[16] In perennially frozen rock, heat transfer at depth (deeper than the seasonal freeze/thaw) occurs primarily by heat conduction. *Kohl* [1999] described temperatures and transient effects below complex topography, but neglected differences due to insolation (i.e., northern and southern slopes have the same temperature). This work did, however, clearly demonstrate the effect of diverging heat flow toward the top of mountains. At that time, only a few investigations [*Haeberli et al.*, 1997; *Wegmann et al.*, 1998] had considered the effects of a two-dimensional shape on mountain permafrost. The three-dimensional subsurface temperature field is the result of four different phenomena [*Kohl and Gruber*, 2003; *Gruber et al.*, 2004c; *Gruber*, 2005]: (1) terrain geometry, (2) spatial temperature variability at the surface, (3) transient variability of surface temperatures, and (4) heterogeneity or anisotropy of rock thermophysical properties. *Noetzli et al.* [2007] have demonstrated that the isotherms below mountain peaks are steeply inclined and sometimes vertical between warm and cold faces of a ridge or peak. They also demonstrated that even under steady state conditions, permafrost bodies below slopes with positive mean annual ground surface temperatures can be induced by a colder surface nearby. The slow responses of the thermal field at depth to changing boundary conditions can cause permafrost to persist for decades or centuries below surfaces that have warmed to positive mean annual temperatures. Temperature gradients determine the direction and magnitude of heat fluxes and, to some degree, the movement of premelted pore water. They are induced primarily by temporal fluctuations or spatial differences of surface temperatures. In permafrost at depth, gradients are mostly small, of constant direction over long time periods [*Noetzli et al.*, 2007], and are subject to slow variations over decades to millennia.

[17] Warming and degradation of permafrost in bedrock is rapid because it usually contains less ice than other types of permafrost. Additionally, warming progresses from several sides in steep terrain and is faster and larger at depth than in situations of one-dimensional warming below flat surfaces [*Noetzli et al.*, 2007]. This means that on the basis of geometry alone, permafrost degradation is faster and deeper near a ridge or peak than in a straight slope. A modeling approach that uses climate change scenarios downscaled from regional and global climate models [*Salzmann et al.*, 2007] to drive surface temperature [*Gruber et al.*, 2004a] and 3D subsurface heat conduction models [*Noetzli et al.*, 2007] has been evaluated on synthetic topography. This approach can readily be applied to real topography to project, for example, thermal evolution in rock below infrastructure. However, the effect of heat advected by water in fractured rock remains an important unknown.

[18] Water percolation in highly fractured rock can contribute rather uniformly to heat transfer and lead to a thicker and earlier development of the active layer as compared to pure heat conduction. In larger clefts, moving water can cause discrete thaw zones that extend significantly into surrounding permafrost. Percolation of water into the tunnels of the Jungfraujoch (3500 m asl.) and the Aiguille du Midi (3830 m asl.) mountain stations, noticed for the first time during summer 2003, was likely caused by this effect. Geophysical monitoring in solid rock walls [*Krautblatter and Hauck*, 2007] has recently identified thawed cleft systems influenced by moving water. There may be several important aspects to the degradation of steep bedrock and its associated instability: rock below 0C can self-heal cracks by freezing, whereas moving water can progressively widen and deepen its passages in thawing rock. Because these deep thaw corridors develop along fractures, this process can contribute to rapid destabilization of much larger volumes of rock than would be expected in a purely conductive system.

4. Ice in Bedrock Permafrost

4.1. Fissures and Fractures

[19] Several accounts of massive ice in Alpine bedrock exist. During construction of the summit station (3820 m asl.) of the Chli Matterhorn cable car near Zermatt, Switzerland, ice-filled cracks were found near the entrances of a tunnel traversing the summit pinnacle [*Keusen and Haeberli*, 1983]. The nearby construction of foundations for a cable car from Hohtälli (3286 m asl.) to Rote Nase (3250 m asl.) revealed ice-filled fractures up to 20 cm wide at depth [*King*, 1996]. Ice-filled joints have also been reported from the Sphinx station (3500 m asl.) of the Jungfraujoch railway, Switzerland [*Wegmann*, 1998] and from the summit of Chli Titlis, Switzerland [*Haeberli et al.*, 1979]. Massive ice has also been found at depths of 42 and 90 m in a borehole drilled in bedrock near Stelvio Pass (3000 m asl.), Italy [*Guglielmin et al.*, 2001] and in a highly fractured zone between 12 and 14 m depth in a borehole drilled in the Colle Nord di Cime Bianche (3100 m asl.), Aosta Valley, Italy (M. Guglielmin, personal communication, 11 April 2006). Personnel at the Aiguille du Midi cable car summit station (3830 m asl.) near Chamonix, France (Figure 2a) noticed water flow into the station tunnels for the first time during summer 2003. Massive ice in detachment zones of rockfall events (Figure 3) bears further witness to this phenomenon.

[20] Similar evidence is available from areas of Arctic and high-elevation permafrost: Massive sill and segregated ice up to 2 m thick has been reported in poorly consolidated bedrock on Ellesmere Island at depths between 10 and 20 m, and was attributed to downward aggrading permafrost following marine regression [*Robinson and Pollard*, 1998]. Relatively pure ice overlain by coal and sandstone was found at depths of 46–68 m in northeastern China and was attributed to a high-pressure injection of water [*Wang*, 1990]. An ice-rich zone in brecciated shale extends to a depth of 5 m below a 35 cm thick active layer on Melville

Figure 3. Detachment surface of a rockfall that happened on the Matterhorn just below the Carrel hut during summer 2003. On the left, massive ice is visible (photo: L. Trucco). A similar event nearby during summer 2006 also revealed massive ice.

Island, Canada [*French et al.*, 1986]. Fractures with massive ice up to 50 cm wide are also found at depths in the order of 100 m in coal mining corridors below mountainous topography with permafrost and polythermal glaciers in Spitsbergen [*Christiansen et al.*, 2005]. Geophysical investigations in Siberia found highly fractured zones that have 6–10% volumetric ice content bounding a permafrost body [*Afanasenko et al.*, 1983]. This zone was interpreted to reach a thickness of up to 100 m and to be related to its long-term permafrost history. This kind of zone is absent in areas of recent, Holocene incision. Ice-rich fractured zones and wide ice-filled cracks to depths of several tens of meters are not unusual in alpine bedrock permafrost.

4.2. Pores and Microfractures

[21] Rock (or concrete) frozen at temperatures typical of terrestrial permafrost usually contains varying amounts of ice, liquid water, and air, depending on the degree of water saturation and temperature. The transient thermal field can be influenced by latent heat effects, even below 0C. The freezing characteristic curve (i.e., the fraction of liquid water as a function of temperature) depends on the shape and diameter distribution of the pore space and the solute content of the pore water and shows a hysteresis between freezing and thawing. Because of its dependence on rock properties, it is likely to be spatially variable. The few published freezing characteristic curves for basalt [*Anderson and Tice*, 1973], tuff [*Akagawa and Fukuda*, 1991], sandstone and limestone [*Mellor*, 1970], and concrete [*Cai and Liu*, 1998] hint at a substantial fraction of pore water that remains liquid even at temperatures around −10C. Similar calorimetric measurements of concrete, which are not plotted as traditional freezing characteristic curves, [*Jacobsen et al.*, 1996; *Penttala*, 1998; *Kaufmann*, 2004] indicate a bimodal distribution, with much freezing taking place in the range above −10C and around −40C. The existence of unfrozen (premelted) water below the bulk freezing temperature is related to two effects [*Dash et al.*, 1995, 2006; *Rempel et al.*, 2004]: (1) curvature-induced premelting, where the equilibrium freezing temperature is depressed at an ice-liquid water interface with its center of curvature in the ice, and (2) interfacial premelting caused by long-range intermolecular forces between different materials (i.e., ice and rock) or different phases. Interfacial melt films have a thickness on the order of nanometers [*Engemann et al.*, 2004; *Rempel et al.*, 2004], whereas pore diameters are often larger [e.g., *Penttala*, 1998; *Inigo et al.*, 2000; *Nicholson*, 2001]. The threshold pore radius for ice intrusion [e.g., *Chatterji*, 1999; *Chen et al.*, 2004] and the thickness of the interfacial melt film [*Fagerlund*, 1973] can be approximated as a function of temperature and a theoretical freezing characteristic curve can be derived on the basis of known pore size distributions [e.g., *Zuber and Marchand*, 2000]. Unsaturated conditions can, as a first approximation, be regarded as a reduction of pore radii by air. Freezing point depression with decreasing degree of saturation, similar to soil systems, can be observed in rock [*Fukuda*, 1983]. Similarly, both air and ice content restrict the movement of liquid water. As a consequence, the permeability of frozen rock is a function of pore space characteristics, saturation, and temperature. This illustrates four important points: (1) Liquid water exists in frozen rock; (2) liquid water can move slowly in frozen rock; (3) permeability decreases with temperature; and (4) because of their dependence on pore structure, these effects will be highly variable between lithologies and within single slope segments.

5. Linking Warming and Destabilization

[22] Instability in rock slopes is usually related to existing fractures [*Hoek and Bray*, 1981; *Abramson et al.*, 2001] along which a rock mass is destabilized by a triggering event. Fractures in perennially frozen rock are likely to contain ice and to experience strong changes during thaw. Five physical processes may link warming permafrost and the destabilization of steep bedrock by altering the conditions of fractures: (1) loss of bonding, (2) ice segregation, (3) volume expansion, (4) hydrostatic pressure, and (5) reduction of shear strength.

[23] Bonding of ice-filled fissures and its reduction or loss during warming or thaw is related to a combination of ice/rock interlocking [*Davies et al.*, 2001] and ice-rock adhesion [*Ryzhkin and Petrenko*, 1997]. This concept is also inherent in the frequently used term "ice-cemented."

[24] Frost heave is the upward displacement of the ground surface by ice segregation during freezing. This phenomenon has been described by *Taber* [1929, 1930] who demonstrated that it is caused by the movement of unfrozen water toward the freezing front and unrelated to volume expansion during phase change. During the 1980s, experimental and theoretical work has established the role of ice segregation for rock breakdown [*Hallet*, 1983; *Walder and Hallet*, 1985, 1986] close to the surface and subject to strong thermal gradients. Nevertheless, small thermal gradients and long freezing duration were recognized early on to increase the importance of ice segregation over volume expansion [*Powers and Helmuth*, 1953; *Walder and Hallet*, 1985]. *Rempel et al.* [2004] have shown that the heaving or disjointing pressure is governed by the temperature

depression below the bulk-melting point, even in the absence of thermal gradients. Slow ice segregation is thus possible in compact rock at depth and its heave rate is limited by the supply of liquid water through the frozen rock. This has been demonstrated by laboratory experiments for high-porosity rocks and strong temperature gradients [*Fukuda*, 1983; *Akagawa and Fukuda*, 1991; *Murton et al.*, 2001]. Ice segregation could affect the stability of steep bedrock permafrost by slowly widening fractures and thus preparing the way for later failure during degradation, or by expanding fractures past a critical value, either slowly or in response to temperature (and permeability) cycles.

[25] Freezing of water is accompanied by a 9% volume expansion. The pressure that can be exerted on confining surfaces is proportional to the temperature depression below the bulk-melting point. A mechanism to transport water into a confined location with cold temperatures is required to strongly affect the stability of steep bedrock permafrost.

[26] Hydrostatic pressure (e.g., in a crevice) is governed by the vertical height that the interconnected saturated zone extends above that location. Hydrostatic pressure can reduce the effective stress in fractures. Thaw can influence hydrostatic pressure and stability in many forms. For example, an ice-filled volume becomes water-filled after melt; melt of an ice-filled volume enables the flow of water to lower areas; or ice (seasonal or perennial) acts as an aquiclude and thus facilitates higher hydrostatic pressure.

[27] Ice strength (internal angle of friction, cohesion, shear and tensile strength) is usually higher at lower temperatures and decreases toward the bulk-melting temperature [*Fish and Zaretsky*, 1997]. Hydrostatic pressure, formation conditions, crystal size [*Voytkovskiy and Golubev*, 1973], impurities, or the inclusion of fine material [*Hooke et al.*, 1972; *Paterson*, 1994; *Arenson et al.*, 2004] can strongly affect ice strength. In addition, the strength of ice-filled joints depends on the ice/rock contact by interlocking or adhesion [*Ryzhkin and Petrenko*, 1997]. These effects result in a complex response in the shear strength of ice-filled discontinuities to warming. This has been demonstrated in centrifuge experiments [*Davies et al.*, 2001], where a stability minimum was shown at temperatures between -1.5 and 0C, and in which the discontinuity was less stable than in the thawed state.

6. Destabilization of Warming Permafrost

6.1. Observed Events

[28] A considerable number of rockfall events originating from Alpine permafrost areas have been described and investigated [*Deline*, 2001, 2002; *Noetzli et al.*, 2003; *Porter and Orombelli*, 1980; *Dutto and Mortara*, 1991; *Dramis et al.*, 1995; *Barla et al.*, 2000; *Fischer et al.*, 2006; *Keller*, 2003]. Outside Europe, many events of often catastrophic dimensions have been documented [*Evans and Clague*, 1988; *Hewitt*, 1988; *Haeberli et al.*, 2003, 2004; *Huggel et al.*, 2005; *Geertsema et al.*, 2006]. This list is incomplete but shows that many events of significant dimensions exist. For none of these events is it certain that permafrost was an important factor in destabilization, and, one can only infer that in many cases permafrost appears to be important. The robust inference of an actual increase of rockfall from these observations is difficult. This is due to the low frequency of events and the unknown bias due to better observation during recent times. It is nevertheless striking, that at least four events with volumes of about 1 million m^3 or more originated from Alpine steep bedrock permafrost areas during the last decade: Brenva (Italy) in 1997, Punta Thurwieser (Italy) in 2004, and Dents du Midi and Dents Blanches (both Switzerland) in 2006.

6.2. Timing and Magnitude of Events

[29] Rockfall from steep bedrock permafrost differs in magnitude and timing and can be classified roughly into active layer formation, active layer thickening, and warming at depth. Active layer formation is an immediate and shallow reaction, active layer thickening takes place on the scale of one season to several years and deep degradation may be delayed by decades, centuries, or millennia. Immediately after the disappearance of ice faces or cold hanging glaciers, an initial active layer develops in the rock exposed. Many rockfall events from recently deglaciated steep slopes [*Fischer et al.*, 2006] may, in part, be attributed to this. A slow temperature rise or an extremely hot summer can degrade permafrost directly beneath the active layer. The rockfall activity during summer 2003 in the Alps is interpreted as a phenomenon of active layer thickening. Large events [e.g., *Dramis et al.*, 1995; *Deline*, 2001] with correspondingly deeper detachment surfaces are caused by the slow reaction of the subsurface temperature field to changed surface conditions and temperatures. Events can progress in retrograde fashion, where thaw-induced rockfall again exposes a fresh surface subject to new temperature conditions. Water circulation and corresponding thaw along fractures can lead to deeper/faster thaw and larger events than suggested by a magnitude-delay relationship based on the assumption that heat diffusion dominates.

6.3. Interpreting Observations

[30] The interpretation of events in terms of their triggering mechanism is challenging in many ways. Exact timing, initial topography, and rock properties are usually difficult to reconstruct and temperature distributions can only be modeled. On the basis of this and the variety of candidate processes for linking thaw and destabilization, more than one course of events can often be hypothesized to have led to failure. However, some inferences can be made. The visibility of ice at the detachment of several events (Figure 3) during a summer heat wave supports the relevance of thawing ice-filled fractures. Many events have originated from ridges, spurs, and peaks, possibly due to more rapid thaw in such geometries [*Noetzli et al.*, 2007]. Most Alpine rockfall from permafrost during 2003 was observed between mid-June and August, and can be attributed with some certainty to the summer heat wave. It is striking, that this active layer thickening is earlier than would be expected on the basis of one-dimensional heat conduction (Figure 4) [cf. *Gruber et al.*, 2004b]. Earlier thawing in complex topography than in one-dimensional cases and thawing by water circulation are possible reasons for this.

[31] The concept of destabilization by warming or thawing ice-filled rock joints points to an important question: Why did the slope not fail when the joint froze? Implicitly, the assumption is made that either other relevant parameters

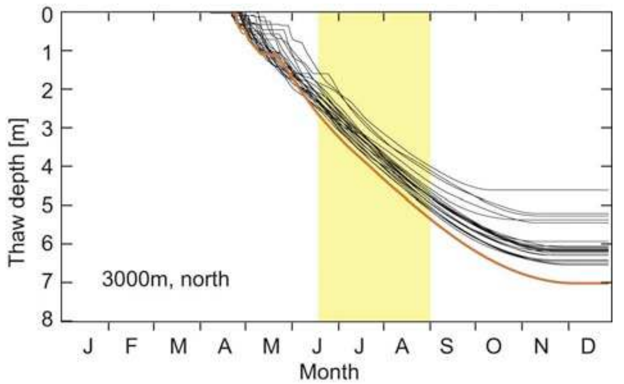

Figure 4. Active layer development in Alpine permafrost, modeled using one-dimensional heat conduction [*Gruber et al.*, 2004a]. The depth of the 0C isotherm is shown for individual years 1982–2002 (black) and 2003 (orange). The maximum depth is kept until the end of the year for clarity. The time of most rockfall activity in the Alps during 2003 (yellow area) is before the time when previous maxima of thaw depth are exceeded.

have changed (e.g., glacial erosion has steepened the rock wall; deglaciation has changed the stress field) or, that the joint has experienced alterations in the frozen state that promote later failure during thaw (Figure 5). Slow ice segregation in existing fractures at depth could, over very long time spans, cause a widening that promotes later failure. While this is theoretically possible and an exciting mechanism of climate control of topography, the importance of this is uncertain at present. As current atmospheric temperatures are about to exceed Holocene maxima, fractures that become thawed and destabilized now may have been frozen for many millennia, allowing ample time for both processes to take place.

7. Conclusion

[32] Permafrost in steep bedrock slopes is important because it is abundant in cold-mountain areas, it reacts rapidly to climate change, and it possesses large potential energy available for mass movements. Degradation of permafrost in steep bedrock can cause destabilization. Although for individual events the role of permafrost cannot be ascertained at present, the assumption is supported by broad evidence: (1) Physical processes exist that link warming and destabilization, (2) thick ice-filled fissures are common in bedrock permafrost, (3) much rockfall originates from permafrost areas, (4) ice has been observed in starting zones, (5) permafrost degradation has been measured and is consistent with atmospheric warming, and (6) the strong rockfall activity in the Alps during the 2003 heat wave points to permafrost thaw as the only plausible explanation. A review of existing research shows that ice-filled joints are common in bedrock permafrost and that they may be widened by ice segregation.

[33] Quantitative understanding and models of temperatures within steep rock faces in complex topography exist and have been validated with near-surface measurements. However, the cooling influence of snow and thermal offset mechanisms are currently not quantified. The investigation and modeling of both effects is difficult because of the strong spatial and temporal variability of surface conditions, energy flows, and mass balance (snow) in steep bedrock terrain. Measurements of active layer processes and permafrost temperatures in steep bedrock are expensive and will likely remain rare. Because of the strong heterogeneity and the cost of measurements, it is likely that considerable uncertainty will remain in the interpretation of measurements, the modeling of rock temperatures, and the analysis of rockfall thermal conditions.

[34] Permafrost degradation in steep bedrock takes place by heat conduction, and by advection of heat by percolating water in fractures. Degradation by conduction can be modeled with some confidence and zones of rapid deep thaw have been demonstrated in ridges and peaks due to warming from several sides. Degradation by advection can lead rapidly to development of deep thaw corridors along fractures in permafrost. This process has the potential to destabilize much greater volumes of rock than conduction can in the same time. Quantitative data and understanding of advective thaw in steep bedrock permafrost is rare. Because a realistic modeling strategy requires much more input data than is the case with conductive models, it will likely remain difficult to make realistic simulations. An increasing proportion of temperate ice in hanging glaciers can cause thawing of the permafrost underneath.

[35] Several conclusions can be drawn for hazard assessment. The heterogeneity of surface and subsurface conditions, possible fast thaw by advection, and the uncertainty related to the actual process responsible for destabilization make the forecasting of future destabilization very difficult. However, four robust statements can be made: (1) Permafrost affects stability and, therefore, the reliable identification of permafrost in steep bedrock is important; (2) destabilization occurs along fractures and, therefore, knowl-

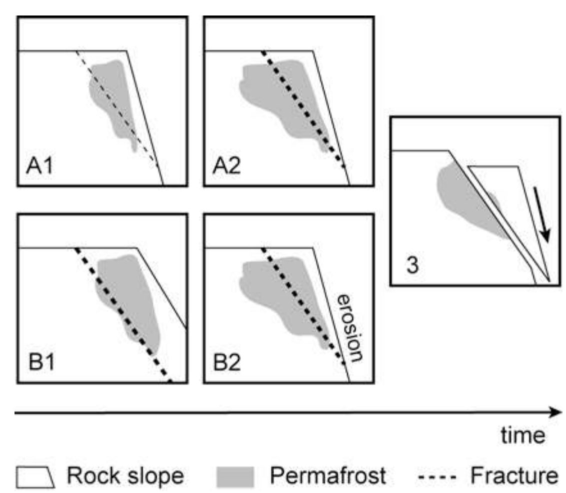

Figure 5. Why does a slope fail during permafrost degradation but not during permafrost formation? Two scenarios that can precede the destabilization of a slope by permafrost warming (3) are proposed: A slope with a stable fracture freezes (scenario A1) that is then widened by ice segregation (scenario A2). A slope with a stable fracture freezes (scenario B1) and is subject to stability-relevant modifications (here geometry, scenario B2).

edge about the fracture characteristics of rock is important; (3) convex topography such as ridges, spurs, and peaks experience faster and deeper thaw than other areas and is likely to be a preferred location of instability; and (4) approximate scenarios for permafrost degradation based on heat conduction can be modeled but advective thaw along fractures must be monitored at critical locations. Although most research on steep bedrock permafrost has been conducted in the Alps, it will likely gain importance in other geographic regions with mountain permafrost. Examples include the Himalayas, where rockfall into the large, newly developing glacial lakes may trigger large floods, and North America and Central Asia, where pipelines and other infrastructure cross mountain permafrost areas.

[36] The destabilization of steep bedrock by permafrost degradation implies a proportion of bedrock slopes in permafrost that is steeper than it would be when thawed. As a consequence, a signal of this climatic control over topography may be detectable in topographic data.

[37] **Acknowledgments.** This work was supported by the Swiss National Science Foundation, grants 21-63678 and 200020-100222, and by a postdoctoral fellowship from the French Ministry of Research. The authors acknowledge the valuable input of one anonymous reviewer and Ivar Berthling, as well as very detailed editorial comments by Fritz Nelson. We are grateful to Alan Rempel for many discussions about the physics of ice segregation.

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