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0012-9402/97/030407-08 $1.50 + 0.20/0
Birkhäuser Verlag, Basel, 1997

Eclogae geol. Helv. 90 (1997) 407–414

Slope stability problems related to glacier shrinkage and permafrost degradation in the Alps

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Key words: Alps, climate change, debris flows, glaciers, high mountains, natural hazards, periglacial, permafrost, rock falls, slope stability

ABSTRACT

Glacier shrinkage in the Alps has been clearly manifest since the middle of the 19th century and could continue beyond the limits of holocene variability in the near future. Changes in Alpine permafrost are less well documented but are likely to take place at various time and depth scales. This development leads to a variety of slope stability problems in bedrock and non-consolidated sediments (moraines and scree slopes). A brief overview, with references to recent literature, is given with regard to characteristic situations and interactions as illustrated by recent events observed in the Alps. The achievement of progress in recognizing and mitigating risks from such slope stability problems in high mountain areas requires improved process understanding from field observations and computer modelling, systematic investigation of natural archives reflecting former slope instability processes and adequate monitoring of potentially critical situations.

ZUSAMMENFASSUNG

Der Gletscherschwund in den Alpen seit der Mitte des 19. Jahrhunderts war markant und könnte in naher Zukunft über die Grenzen der holozänen Variabilität hinausführen. Änderungen des alpinen Permafrostes sind weniger gut dokumentiert und vollziehen sich wahrscheinlich in unterschiedlichen Zeit- und Tiefenbereichen. Diese Entwicklung führt zu einer Anzahl von Hangstabilitäts-Problemen in Fels und unverfestigten Sedimenten (Moränen, Schutthalden). Eine kurze Übersicht mit Hinweisen auf neue Literatur wird hinsichtlich charakteristischer Situationen und Wechselbeziehungen anhand von aktuellen Fallbeispielen aus den Alpen illustriert. Fortschritte bei der Früherkennung und Bewältigung von Hangstabilitäts-Risiken im Hochgebirge erfordern ein verbessertes Prozessverständnis aus Feldbeobachtungen und Computermodellen, eine systematische Analyse von natürlichen Archiven, die ehemalige Instabilitätsprozesse reflektieren, und eine gezielte Beobachtung von potentiell kritischen Situationen.

Introduction

In Chapter 7 dealing with the cryosphere, the 1995 report by the Intergovernmental Panel on Climate Change (IPCC) Working Group II on impacts, adaptations and mitigation of climate change states that cryospheric change will reduce slope stability and increase the incidence of natural hazards for people, structures and communication links in mountain lands and continental permafrost areas (Fitzharris et al. 1995). The IPCC report attributes medium confidence to this statement. Such limited confidence is due to the difficulty connected with the investigation and safe assessment of rare events in remote areas, especially in the case of steep and potentially unstable slopes in high mountain areas such as the Alps (Haeberli 1992, 1996a, 1996b). The following overview is largely based on the experience gained from a research project about the melting of perennial ice and its effects on natural disasters in high mountain areas (Haeberli et al. 1997) carried out within the ongoing Swiss National Research Programme 31 "Climate Change and Natural Catastrophes". Additional information has been obtained from an ongoing research project investigating permafrost effects on rock stability. Blair (1994), Evans & Clague (1993) and O'Connor & Costa (1993) describe comparable observations from other glacierized mountains of the world.

In the chain of processes linking slope stability via preparatory factors to marginal stability and from there via triggering factors to instability with controlling factors, the evolution of ice above and below the ground surface usually represents a preparatory factor, whereas water commonly acts as triggering factor. The resulting instabilities relate to the interactions between

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(1) permafrost and bedrock;
(2) hanging glaciers, permafrost and bedrock;
(3) glacier tongues, permafrost and bedrock;
(4) glacier tongues and bedrock;
(5) glacier tongues and loose sediments (moraines); and
(6) permafrost and loose sediments (scree, rock glacier fronts).

This type of enumeration follows the general sequence of phenomena occurring between the top of the mountains along flowing glaciers down to the margins of the periglacial belt.

Glacier shrinkage and permafrost degradation in the Alps

Glacierized and perennially frozen mountain areas would be among the most heavily affected parts of the world in the event of accelerated future warming. Since the middle of the past century – the end of the "Little Ice Age" – the glacierization of the European Alps has lost about 30 to 40% in glacierized surface area and around half its original ice volume. The estimated total glacier volume in the European Alps was some 130 km^3 for the mid-1970s; strongly negative mass balances have caused an additional loss of about 10 to 20% of this remaining ice volume since 1980 (Haeberli & Hoelzle 1995, Maisch et al. 1997). Average secular lowering of ice surfaces is measured in tens of meters but this surface lowering increases from near zero in the uppermost parts of the glacier towards the tongue where it can reach up to 300 meters in case of the Great Aletsch glacier. Periglacial permafrost in the Alps today occupies an area which is comparable in extent to the glacierized area and must have been affected as well, but its secular evolution is much less well known. Alpine permafrost is typically several decameters to more than 100 m thick and has characteristic mean annual surface temperatures between the melting point and about –3°C (Vonder Mühll & Holub 1992). With continued or even accelerated atmospheric warming, large parts of the Alpine glaciers could disappear within decades and extended permafrost slopes could start thawing (Hoelzle & Haeberli 1995) – first from the permafrost table downwards but later and for extended time periods also from the permafrost base (the interior of mountain slopes) upwards. Such a development would be without historical and perhaps even holocene precedence.

Instability situations

1. Permafrost and bedrock

Field and laboratory experiments (for instance, Pancza & Ozouf 1988) as well as rock glacier analyses (Barsch 1977) document characteristic Alpine cliff recession rates in the order of mm per year. The processes involved (frost shattering, rock fall) have highly variable time and depth scales. With regard to near-surface processes at the decimeter-to-meter scale, freeze/thaw cycles, the degree of saturation of wet rocks and the tensile strength of fractured rocks play a dominant role

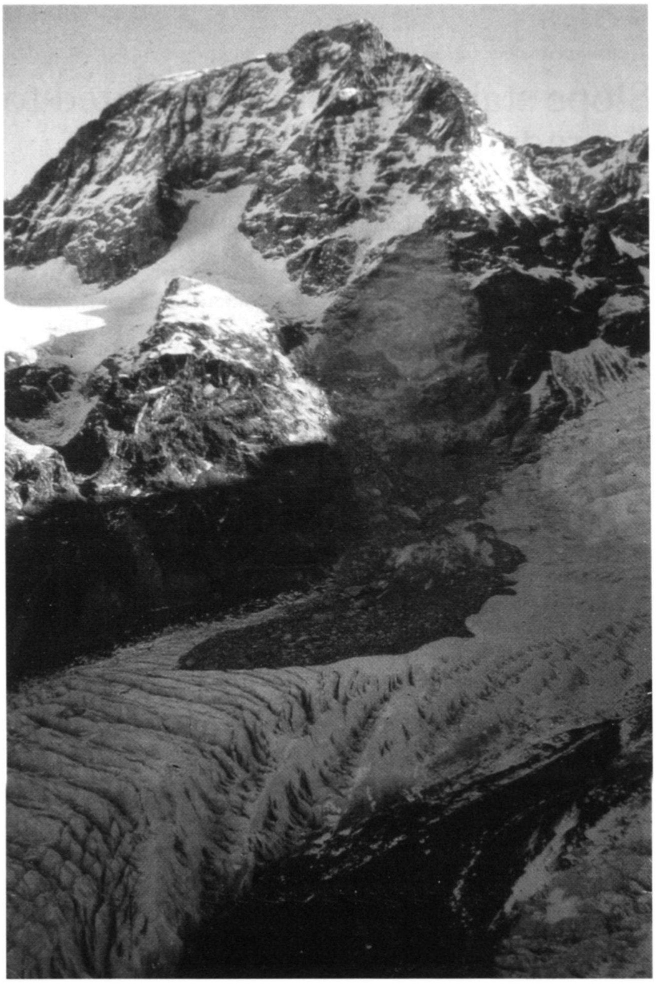

Fig. 1. Rock fall (volume about 300,000 m^3 from permafrost slopes at Piz Morteratsch (Eastern Swiss Alps) onto Tschierva glacier. Foto by J. Schweizer, fall 1988.

(Matsuoka 1990, 1991). At lower frequencies (seasons) and greater depths (meter-to-decameter scale), ice segregation in rock cracks may be the most effective process with respect to rock destruction (Hallet et al. 1991). Optimal fracture propagation in common lithologies thereby appears to take place at temperatures between about –3°C and –6°C and sustained subzero temperatures could be more important than the number of high-frequency freeze/thaw cycles. Rock destruction by ice segregation must be especially intense with spring meltwater from snow penetrating into permafrost (Ødegard & Sollid 1993, cf. Keller & Gubler 1993). The influence of permafrost on the destabilization of rock walls (Fig. 1; Dramis et al. 1995, Schindler et al. 1993), however, has remained until now a virtually untouched field of research. The freezing/thawing front at the permafrost base reaches depths up to hundreds of meters. Typical time scales thereby involved can be many millennia for climatic variations during late Quaternary times but

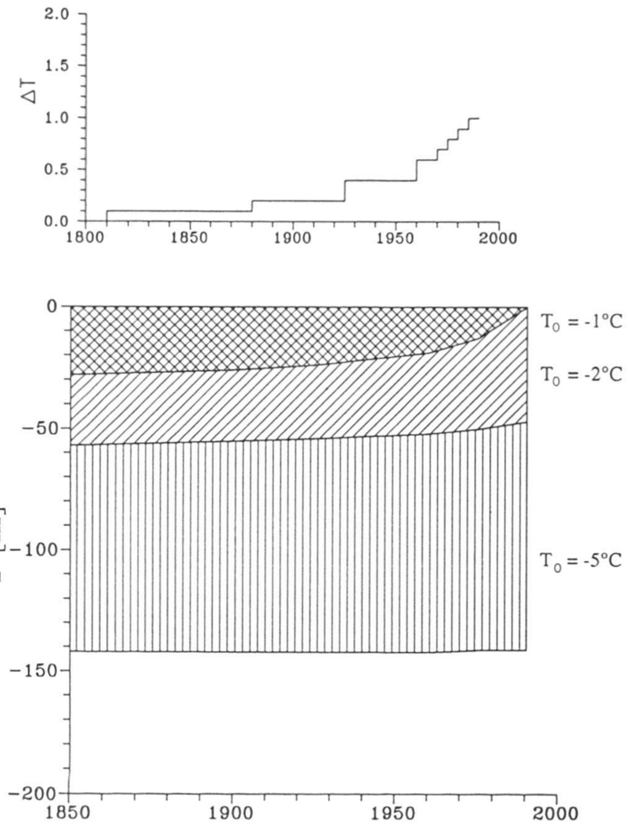

Fig. 2. Changes of permafrost thickness as a consequence of 20th century warming. Model calculation using pure heat conduction for three sets of mean surface temperature and initial permafrost thickness (T_0 = −1°C, −2°C and −5°C for the year 1850).

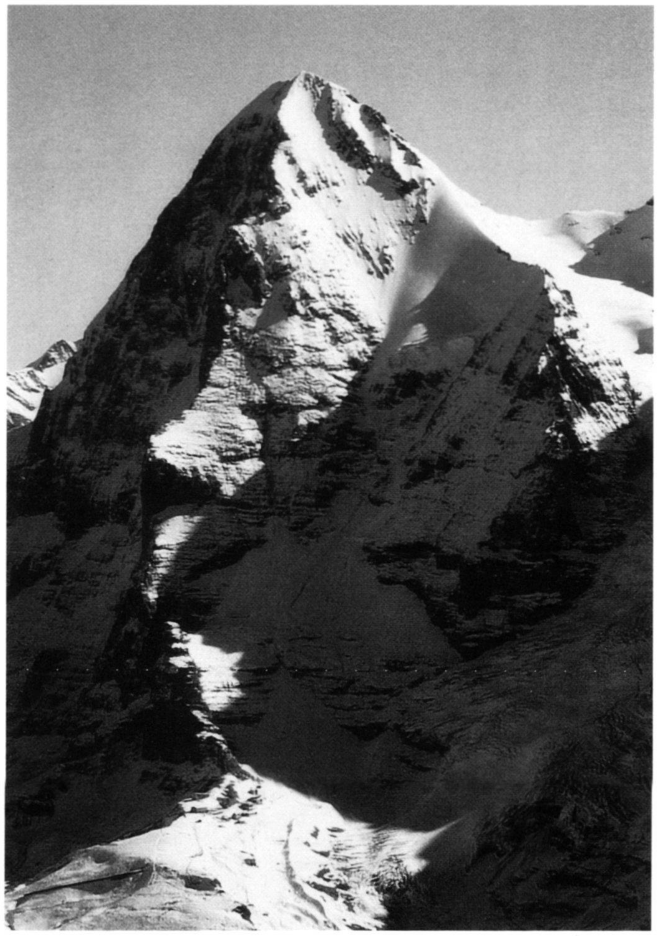

Fig. 3. Hanging glacier on the western slope of Eiger (Bernese Alps). Foto: S. Bader, October 1981.

could, in places, shorten to decades or centuries with accelerated future warming and enhanced convective heat transport by water circulating in fissures and cracks which become ice-free and permeable. The most important processes concern the fracturing of rocks during freezing, the change in hydraulic conductivity and pore water pressure/circulation during freezing/thawing and the change in surface geometry by major rockfalls. Critical parameters are temperature, ice content, permeability, pore pressure, tensile strength and rock deformation. Model calculations based on analytical solutions of the 1-D heat conduction equation (Fig. 2; cf. Carslaw & Jaeger 1986) indicate that the warmest and shallowest parts of mountain permafrost probably have now begun to react to the effects of 20th century warming by raising the permafrost base. Similar calculations taking phase changes into account (Neumann solution) illustrate the strong retarding effects of latent heat exchange with respect to the heat wave penetration at depth. The evolution may nevertheless cause unfavourable changes in hydraulic conductivity (onset of convective heat transfer in opening fissures at depth) and escapes the possibilities of simple/economic direct observation. Colder and thicker permafrost bodies are likely to react in the future with considerable delay and the adjustment to new equilibrium conditions may take centuries if not millennia. A detailed process study is presently being carried out by one of the authors (M. W.).

2. *Hanging glaciers, permafrost and bedrock*

The ablation (regular mass loss) of many glaciers on steep high-altitude mountain slopes is through ice avalanching (Fig. 3). Sometimes very large parts of ice may become detached (Alean 1984, Dutto et al. 1991, Lüthi 1994, Röthlisberger 1981). Coupling of glaciers and permafrost is essential for analyzing the stability of such steep hanging glaciers. At very high altitudes and on the shadow side of mountain peaks, firn and ice temperatures are far below zero. At lower altitudes, however, and on slopes more exposed to the sun, firn temperatures are at the melting point due to percolating meltwater. Such a

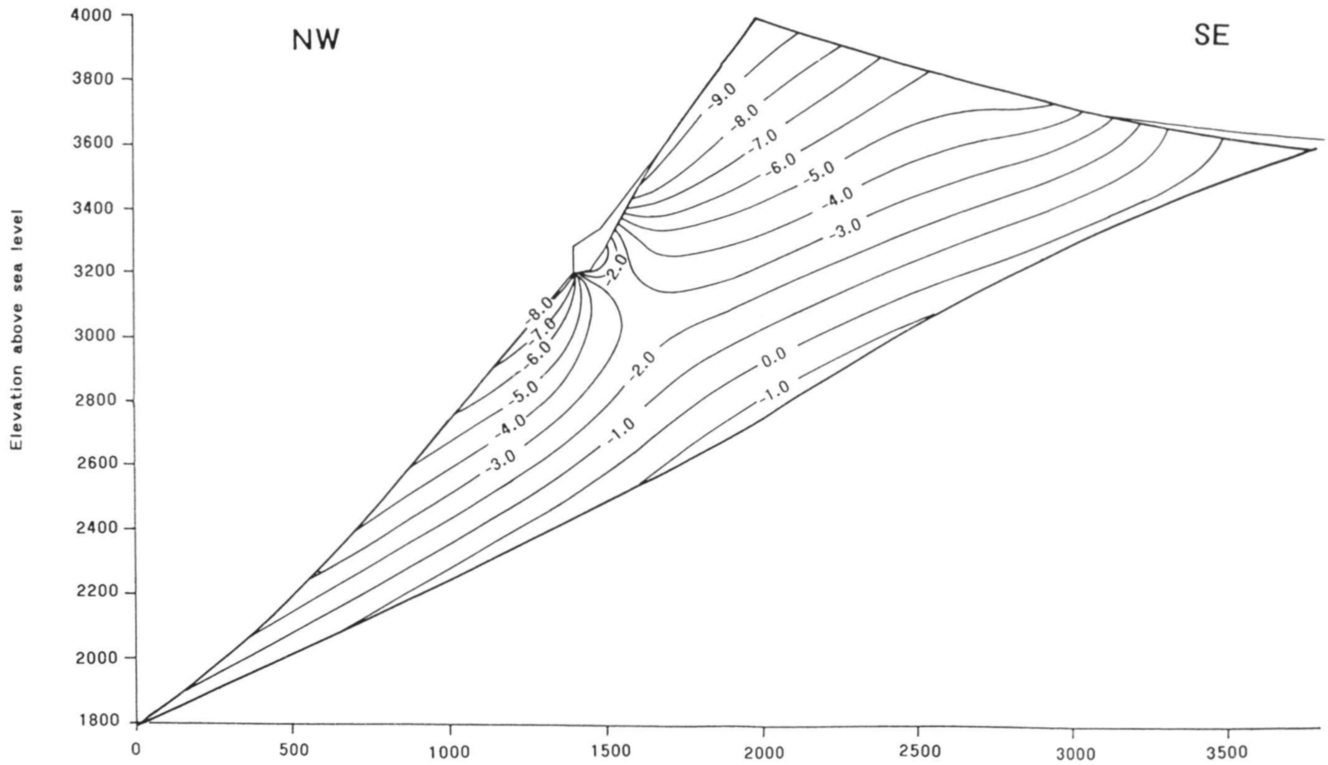

Fig. 4. Two-dimensional temperature distribution calculated on the basis of pure heat conduction for the western slope of Eiger (cf. Fig. 3) illustrating a deep thermal anomaly introduced by the polythermal base of the hanging glacier. The effects are likely to be less pronounced when accounting for the 3-D effects.

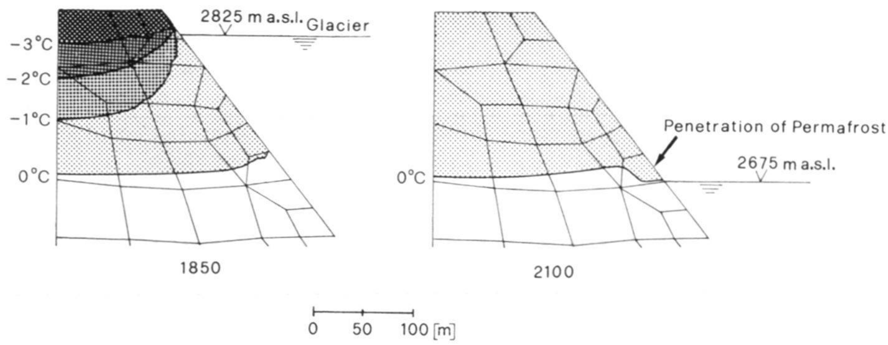

Fig. 5. Calculated temperature distribution (comparison of the situation in the years of 1850 and 2100, respectively) after a modelled 1300 year history of heat transfer forced by surface temperature variations within the rock spur carrying the Konkordia Hut (cf. Fig. 6). Water/ice content in the rock masses is assumed to be 3% by volume and a reasonable history is taken into account with respect to fluctuating glacier surface altitudes and corresponding rock temperature changes. Penetration of permafrost on the shadow side of the rock spur is illustrated.

situation induces a complex polythermal structure with only marginal parts of the hanging glaciers – especially the vertical front – consisting of cold ice frozen to the underlying bedrock. This basal temperature pattern probably introduces a deep-reaching thermal anomaly within the underlying permafrost (Fig. 4), enabling strong lateral heat flow through the base of the hanging glacier front. The geometry and thermal condition of this ice front, where shear stresses can reach values close to the strength of ice, appears to constitute the key factor controlling the stability of entire ice bodies (Lüthi 1994, Röthlisberger 1981). Climate changes may introduce highly complex feedback mechanisms involving surface geometry, firn accumulation, en-/subglacial temperatures and stress distribution. Long-term monitoring of ice geometries using aerial photography may help to detect unfavourable developments. Independently of such difficult attempts to forecast the time of instability, the

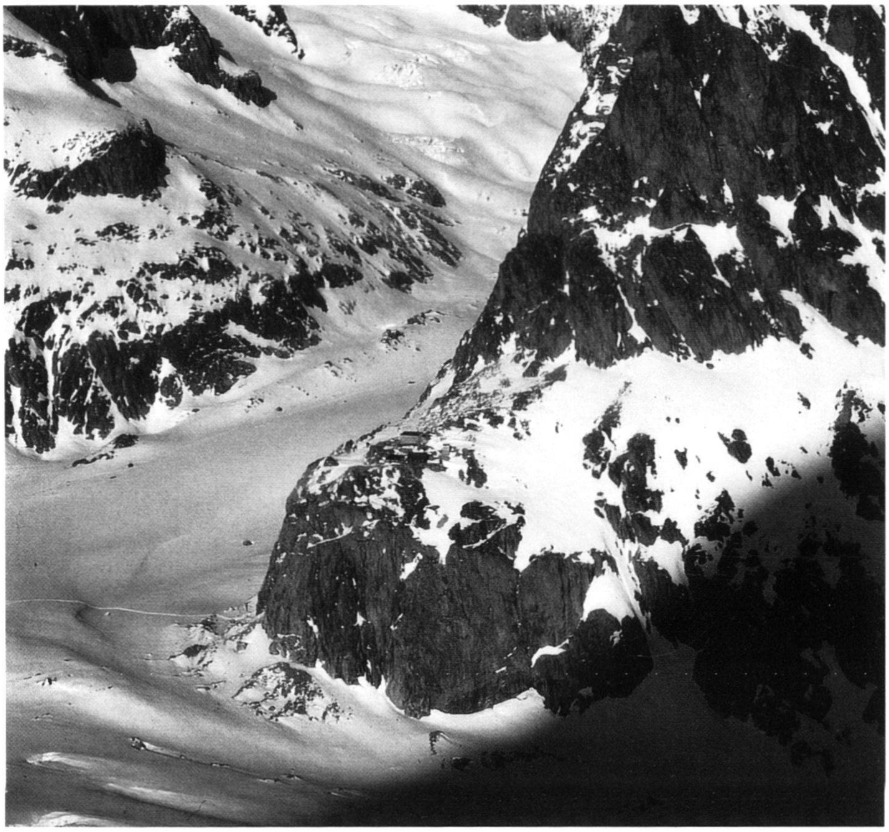

Fig. 6. Rock spur carrying the Konkordia Hut of the Swiss Alpine Club at the confluence of the main tributaries to the Great Aletsch glacier. Surface lowering at the rock spur exceeds 100 meters since the middle of the last century. Foto: W. Haeberli, February 1993.

run-out distance of potential ice avalanches can be quite realistically assessed (Alean 1984, Haeberli et al. 1989, cf. also the hazard map prepared by Bieri 1996 including an avalanche from Gutz glacier near Grindelwald which – in fact – happened a few months later).

3. Glacier tongues, permafrost and bedrock

In the ablation area and especially towards the snout of valley glaciers, the lowering of ice surfaces in the course of the past century can easily exceed 100 meters. This vertical loss in valley filling induces a change in the stress field inside the confining mountain walls. On slopes protected against direct solar radiation, the lowering of glacier surfaces can enable the penetration of negative temperatures (permafrost) into and the formation of ice within rock walls originally covered by temperate ice. This process is illustrated by model calculations of transient heat transfer including latent heat effects (Fig. 5) for the rock spur carrying the Konkordia Hut (Fig. 6) at the confluence of the main tributaries to the Great Aletsch glacier (Keusen & Wegmann 1996). The penetration of the freezing front into previously thawed material has the potential of intensifying rock destruction through ice formation in cracks and fissures. Such ice formation, in turn, reduces the near-surface permeability of the rock walls involved and affects hydraulic pressures inside the still open (non-frozen) fissured rocks. The general lowering of water pressures in lateral rock walls accompanying the disappearance of temperate glaciers may thus be counteracted and the rock-wall stability altered.

4. Glacier tongues and bedrock

In the absence of permafrost – i.e., at low altitudes and on sunny slopes – the reorientation of the stress field within lateral valley walls evolves in parallel to a general head reduction of water circulating in fissures and cracks, due to the reduction of water pressure with decreasing ice thickness. Slope stability may become critical in many instances (Blair 1994, Evans & Clague 1993). Figure 7 shows a large slope instability developing on the orographic right side of the Great Aletsch glacier. At this site, ice surface lowering amounts to about 200 meters since the past century and recently accelerated to about 50 meters during the past decade alone. Horizontal displacements sharply increased since the 1970s to characteristic values of a few decimeters per year (A. Kääb, written communication).

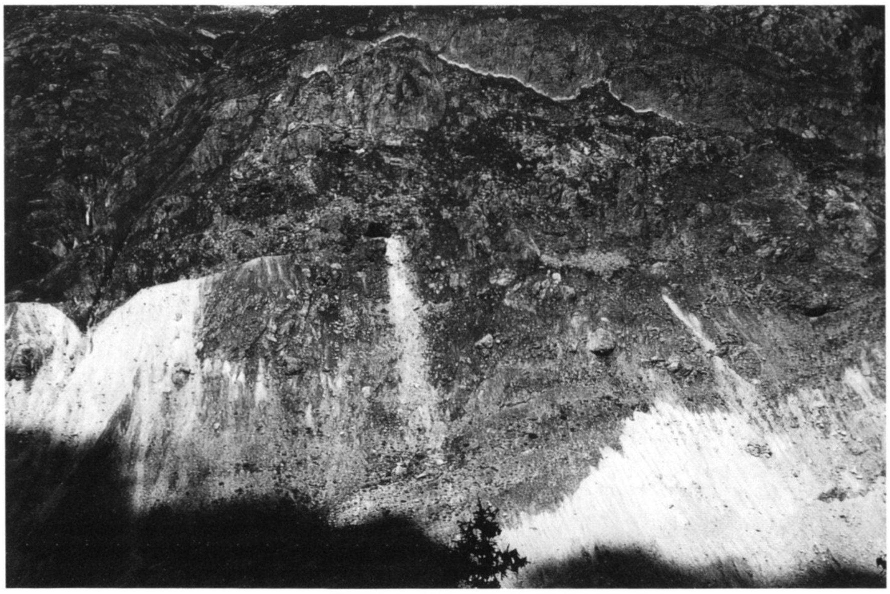

Fig. 7. Large-scale slope instability on the orographic right side of Great Aletsch glacier (Valais). The upper limit of the unstable slope is marked by a bright, lichen-free rock-band. Surface lowering of the glacier and slope movement have markedly accelerated during recent years (A. Kääb, written communication). Foto: W. Haeberli, June 1994.

Fig. 8. Breach and debris-flow starting zone at Bodmer glacier (Valais). The erosion of the roughly 10 to 20 m deep gully was triggered by heavy precipitation on 24 September 1994. Foto: W. Haeberli, October 1994.

5. Glacier tongues and loose sediments (moraines)

The largest debris flows in the Alps usually relate to glacier floods from water pocket ruptures, from outbursts of moraine- or ice-dammed lakes or simply from heavy precipitation (Fig. 8). Breaching of moraine dams involves piping (progressive groundwater flow) within the morainic material, liquefied flow/slippage on steep slopes and overtopping with retrogressive incision (Haeberli 1996a, 1996b, cf. Clague & Evans 1994, Jackson et al. 1989). In order to check the stability of moraine dams and lakes with respect to piping, the gradient between the lake and the spring from groundwater can be compared with the critical hydraulic gradient for progressive groundwater flow. The risk of slippage in cohesionless material is usually judged by defining a safety factor for slope stability where the vertical extent of water-saturated material and the depth to the assumed slip surface are the most important factors. Formation of a breach within morainic material with highly variable grain size leads to the development of block pavements, the destruction of which requires a critical water depth. Vonder Mühll et al. (1996) describe a combination of geophysical

Fig. 9. Rock glacier front and debris-flow starting zone in the Kintole/Chessi area above the Valley of Zermatt (Valais). Foto: W. Haeberli, August 1992.

soundings (seismic refraction, geoelectrical resistivity and gravimetry) which enables stability assessments with respect to the large moraine dams built up by glaciers with excess debris production from the surrounding rock walls.

6. Permafrost and loose sediments (scree, rock glacier fronts)

As a consequence of intense snowmelt and/or heavy precipitation, debris flows of highly variable size may also form at marginal permafrost sites in scree of debris cones or rock glacier fronts (Fig. 9; cf. Dikau et al. 1996, Rebetez et al. 1997, Zimmermann & Haeberli 1992). The transition – in space as well as in time – from the frozen to the unfrozen state involves a loss of cohesion with simultaneous build-up of internal friction in originally ice-supersaturated materials and strong losses in resistance against erosion by running water. Especially dangerous are transitional conditions of water-saturated fine material remaining on steeply inclined permafrost tables of thawing permafrost and large caverns originating from the disappearance of massive ground ice bodies and leading to extreme hydraulic heterogeneity in non-consolidated materials. Physical and mathematical modelling with respect to debris flow dynamics (for instance, Davies 1988, Rickenmann 1991) reveal the limitations of the present understanding regarding such important processes as the transportation of largest grains at the surface, limitation of depth erosion along the flow path, or the sudden halt of flow in the runout zone. Empirical approaches to quantifying the most important trigger conditions and flow parameters are discussed by Rickenmann & Zimmermann (1993) and Rickenmann (1995) based on a great number of Alpine cases. Maximum volume evacuated per unit channel length, for instance, usually remains below some 500 to 700 m^3 and the fact that debris flow trajectories have an overall slope greater than about 10° (cf. Clague & Evans 1994) can be used to estimate maximum runout distances.

Perspectives and recommendations

Glacier shrinkage and permafrost degradation induce complex problems of slope stability in bedrock as well as in non-consolidated sediments (moraines, scree). Such problems may become more acute in the future and develop beyond existing historical experience if atmospheric warming indeed continues or even accelerates. Progress in recognizing and mitigating risks from such slope stability problems in high mountain areas requires improved process understanding from field observations and computer modelling with respect to stress distribution, hydraulic conditions and thaw destabilization in rock walls and moraines/scree slopes. Debris-covered glacier tongues, rock glacier surfaces, moraines and debris flow cones constitute natural archives reflecting former slope instability processes and should be systematically investigated in order to quantify corresponding mass fluxes. Heavily fissured rock walls in relatively warm and shallow permafrost, thick moraine dams on steep slopes and periglacial lakes should be mapped as presenting potentially critical situations. Repeated aerial photography with photogrammetric analysis should be used to follow the development of recorded events, lowering glacier and permafrost surfaces, moving or degrading slopes and growing periglacial lakes. Based on this type of evidence, hazard assessments with respect to potential catastrophes may be improved (cf. Kääb & Haeberli 1996).

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Manuscript received February 14, 1997
Revision accepted May 28, 1997