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SURVEY AND PREVENTION OF EXTREME GLACIOLOGICAL HAZARDS IN EUROPEAN MOUNTAINOUS REGIONS

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Compiled by Didier Richard and Michel Gay

1. Jökulhlaups and drainage of glacier dammed lakes

Jökulhlaups are glacier outburst floods caused by the sudden drainage of water reservoirs located in (internal water pockets), on or at the margin of glaciers. During the flood, the discharge at the glacier terminus can increase by more than one order of magnitude within a time period varying between hours and days, as shown by documented past events. Jökulhlaups have caused substantial damage in the past in most glacierized high mountain areas in the world, including Switzerland. Glacier outburst floods are documented for the Alps (Röthlisberger, 1981; Haeberli, 1983; Raymond et al., 2003), Norway (Liestøl, 1956), Iceland (e.g. Björnsson, 1992, 2002), Canada (Clague and Mathews, 1973; Clarke, 1982), Alaska (e.g. Anderson et al., in press) and most other glaciated regions of the world (e.g. Walder and Costa, 1996; Tweed and Russell, 1999). Glacier floods represent in general the highest and most farreaching glacial risk with the highest potential of disaster and damages. Because of accelerated glacier retreat due to climate warming, former glacierdammed lakes e.g. Märjelensee, which is known to have drained periodically with catastrophic consequences in the last century (Raymond et al., 2003) can almost disappear, while new ones can form at places without known historical records (e.g. Belvédère glacier, Monto Rosa, Italy, (Haeberli et al., 2002)). For this reason and because in some areas in the Alps infrastructures and settlements were developed only recently, the hazard potential of existing and newly forming glacierdammed lakes must be considered. For an efficient hazard assessment strategy, an accurate prediction of the timing, duration and magnitude of the glacier flood event is crucial. Furthermore, during megafloods from large icedammed lakes which formed in front of the Laurentide Ice Sheet during the last deglaciation (approx. 812 kyr. BP), huge amounts of fresh water were routed to the Arctic Ocean within a short time period (a few years). According to recent studies, these floods are linked to abrupt but shortlived climatic cooling events (Clarke, 2002; Broecker and Denton, 1990).

Outburst floods from subglacial or icedammed lakes (e.g. Björnsson, 1992) are commonly described to occur when the lake level has risen to a critical level before the hydrostatic water pressure maintained by the lake is equal to the ice overburden pressure of the ice dam. As water begins to leak underneath the dam, flow typically localizes in one or a few channels incised into the ice. Such channels increase rapidly in size due to meltback of the ice walls caused by the dissipation of potential energy (e.g. Röthlisberger, 1972; Nye, 1976). As lake drainage proceeds, water pressure in the channel drops and creepclosure of ice dominates meltenlargement progressively. The rapid closure or even collapse of the channel can stop the flood, even if the lake is not empty. The typical jökulhlauphydrograph has an exponentially ascending limb (because of progressive channelgrowth) and a steep falling limb, reflecting rapid closure of the conduit (e.g. Rist, 1955; Björnsson, 1974).

Röthlisberger (1972) and Shreve (1972) were the first to describe the physics of steady water flow through icewalled conduits. Based on these theories, Nye (1976) published a timedependent theory and successfully simulated ”slow” (several days) jökulhlaup events. In the Nye theory the effects of lake temperature are ignored and several assumptions are made: a simplified form of heat transfer is postulated and water flow is assumed to be driven by the hydraulic potential gradient averaged over the length of the conduit. Clarke (1982) modified the theory to account for the effects of lake temperature and reservoir geometry. Spring (1980) and Spring and Hutter (1981, 1982) developed the most complete theory of the drainage process, including the temporally evolving water velocity, discharge, temperature and crosssectional area over the entire length of the conduit. However, their system of equations was not solvable without major simplifications, so that they could not demonstrate the effective impact of their theoretical innovation. Their system of equations suffers from numerical stiffness, a fact which restricts the applicability of their theory in a numerical model.

Recently, Clarke (in press) removed the source of numerical stiffness that plagued the SpringHutter model and the full system of equations can now be solved numerically. Analyzing the behaviour of the model, Clarke (in press) showed that some of the key assumption of previous models (e.g. Nye, 1976; Clarke, 1982) are not generally satisfied. While previous studies assume that outburst floods are controlled by conditions at a single point along the conduit (bottleneck model), solutions of the more general SpringHutter model indicate, that flow constrictions appear to be highly mobile during the course of an outburst flood, and for some floods the dominant flow constriction is not located close to the water reservoir (as it was assumed in previous works), but near the tunnel outlet.

All theories presented so far are able to explain socalled ”slowly rising” jökulhlaups, in which a balance between melting of ice and release of potential energy is described. Some observations indicate that ”rapidly rising” jökulhlaups can occur (e.g. Björnsson, 2002; Jóhannesson, 2002b), which are governed by a fundamentally different physical mechanism as in the ”slowly rising” case. The Grímsvötn (Vatnajökull, Island) jökulhlaup of 1996, triggered by a subglacial volcanic eruption, is a typical example illustrating the ”rapid rising” case. According to Johannesson (2002b), ”rapidly rising” jökulhlaups can be large or small, and may originate from both, subglacial and marginal lakes at many different locations. They are also not limited to cases induced by volcanic eruptions. It follows that the existence of two main types of jökulhlaups must be assumed, governed by fundamentally different mechanisms and resulting in fundamentally different hydrographs. The example of Grímsvötn shows that both types can occur at a single location and that the trigger mechanism determines the type of lake drainage. When a jökulhlaup is initiated by a water pressure exceeding the ice overburden pressure of the ice dam, a pressure wave outburst will occur, whereas a flood triggered at a lower water level will be characterized by the progressive enlargement of a single drainage channel. These two situations are believed to be determined by the rate at which the lake fills with water. If the water level rises slowly, preexisting hydraulic connections can adjust to the changing conditions and the hydraulic seal that controls the flood initiation migrates closer to the lake. The flood is triggered when the seal reaches the lake position and the flow path increases gradually in size by meltenlargement. In contrast, in the case of a fast rising lake level drainage occurs when the water pressure in the ice dam exceeds the ice overburden pressure. Flow paths are then rapidly enlarged by hydrofracturing and water spreads out laterally along the glacier bed.

Empirical methods to estimate the peak discharge from lake characteristics have been developed by Clague and Mathews (1973) and Walder and Costa (1996). Ng and Björnsson (in press) investigated the ClagueMathews relation in light of the Nyetheory. Fowler and Ng (1996) and Ng (1998) advanced the Nyetheory by considering the flood conduit to be floored by deformable sediment and Fowler (1999) concentrated on the conditions for triggering outburst floods from a lake dammed by a soft-bedded glacier. Walder and Costa (1996) considered a breach drainage mechanism and in analogy to the Nyetheory, Raymond and Nolan (2000) developed a model to describe floods by overtopping of the ice dam. Szilder et al. (1997) analyzed the stability regimes for water flow in a subglacial conduit draining from a reservoir.

The assessment and prevention of natural hazards related to jökulhlaups requires a reliable prediction of the timing, duration and magnitude of the outburst flood. The following limitations and open questions can be identified in previous works and existing theories:

- The empirically derived relation of Clague and Mathews (1973) is helpful to estimate the magnitude of a jökulhlaup from the lake volume. However, the scatter of observed versus predicted values is considerable (Walder and Costa, 1996) and the relation appears to be site-specific (Ng and Björnsson, in press).
- The mechanism of jökulhlaup initiation is poorly understood. Some lakes drain when the water level reaches floatation pressure of the ice dam whereas others release their water at a lower level (e.g. Walder and Costa, 1996; Björnsson, 1992; Jóhannesson, 2002a,b). Fowler (1999) developed a model which explains both types of triggering a jökulhlaup from Grímsvötn. He demonstrated that the seal will break (the flood will be initiated) while the lake level is below floatation pressure when the lake filling rate is low (in the case when the potential gradient toward the lake is low). On the other hand, floatation can be achieved before a flood is initiated when the lake filling rate is rapid.
- To assess the flood duration and peak discharge, the models of Nye (1976) and Clarke (1982) yield reasonable estimates in some cases (Nye, 1976; Clarke, 1982; Björnsson, 1992), whereas in other cases they fail (Walder and Costa, 1996; Björnsson, 2002; Jóhannesson, 2002b).
- Observations of supercooled water or water at the freezing point disagree with model calculations which predict a higher water temperature at the glacier terminus (Johannesson, 2002b; Clarke, in press).
- The roughness coefficient in the hydraulic equations (e.g. Nye, 1976; Spring and Hutter, 1981; Clarke, 1982) is implausibly high throughout all the different studies carried out so far. Clarke (in press) associates this shortcoming with the unsatisfactory formulation of heattransfer that is commonly used. Further, the high value may result from the simplifying assumption of a

straight conduit axis since curvature and irregularities in the geometry of natural conduits contribute additionally to frictional head losses (Schuler, 2002).
- The turbulent friction laws used so far were empirically derived for flow of water through pipes with rigid walls (e.g. Bird et al. (1960)) and not for water flowing through its solid phase.
- The spectacular 1996 jökulhlaup from Grímsvötn (Gudmundsson et al., 1997; Jóhannesson, 2002b; Björnsson, 2002) revealed several unexpected aspects and, thus, highlights the need to improve existing theories. For example, the catastrophic release of water (Björnsson, 2002; Jóhannesson, 2002b) and ice fracturing at the surface (Roberts et al., 2000; Waller et al., 2001) strongly suggest that the flood moved like a pressure wave that propagated spatially distributed over the glacier bed. This governing process differs substantially from the mechanism described in the classical theories (Nye, 1976; Clarke, 1982, 2003) where water flow is confined to a single conduit.

Data about jökulhlaups are sparse and available data are of highly variable quality and detailedness (Walder and Costa, 1996; Anderson et al., in press; Clarke, in press). However, substantial progress related to the above mentioned open questions require better observational data (Clarke, in press), in particular, high quality data of:
- the geometry of the glacier bed and surface between the lake and the glacier snout,
- the strain rates at the glacier surface in the vicinity of the ice dam before, during and after the drainage of the lake,
- the volume and rate of release of water impounded in an icedammed lake,
- the temperature of the impounded water, and
- the timing and hydrograph of the resulting flood at the glacier terminus as well as the water temperature and ice content.

Subglacial drainage system

Closely associated with the outburst of water from glacierdammed lakes is the question of how this water is routed within and beneath the ice to the glacier terminus. The study of subglacial drainage is of importance since it controls glacier dynamics (e.g. Iken and Truffer, 1997; Iken and Bindschadler, 1986; Iverson et al., 1995). Because water moves in response to potential gradients, drainage beneath glaciers is in the simplest case controlled by geometric influences of the ice and bed. Thus, measurements of ice surface and bedrock topography can be used to gain insights into the likely locations of subglacial drainage catchments and preferential drainage axes. This approach requires the derivation of two geometric quantities: a map of the hydraulic potential surface and the distribution of a terrain characteristic called upstream area across the glacier bed (Flowers and Clarke, 1999, and references therein). The total hydraulic potential is the sum of pressure potential and elevation potential (Shreve, 1972), where the water pressure is usually assumed to be spatially uniform and equal to some fraction of the iceoverburden pressure. For a particular gridcell of a hydraulic potential surface, the upstream area is computed as the sum of all gridcell areas that are upstream and connected (Zevenbergen and Thorne, 1987).

To explore the trigger mechanisms of jökulhlaups and to obtain more insight into the incompletely understood controlling mechanisms combined field and modelling studies are necessary.
In particular the following open questions must be answered:

1. How the position of the seal is controlling the start of the lake drainage ?
2. What is the effect of the rising rate of the lake level on the start of drainage ?
3. Is the currently used heat exchange empirism between the water and the ice in the englacial or subglacial conduits appropriate and what alternatives can be proposed ?

The mechanics of flood initiation, as describes by Fowler (1999), should be analysed with the Spring-Hutter (1982) model. Fowler's analysis is based on the Röthlisberger-Nye model, in which the assumption is made, that the flood is controlled by a "seal" acting as a flow-restricting bottleneck and which is spatially fixed and situated close to the reservoir so that the water temperatur corresponds to that of the reservoir. As Clarke (in press) has shown in his work, this assumption is not generally valid. Fowler's (1999) results indicating that the Röthlisberger-Nye's model provides a robust explanation of the flood initiation is in sharp contrast with Johannesson's (2002) conclusions.

The Spring-Hutter model and earlier works appears to overestimate the temperature of the water exiting the glacier at the snout. The reason is probably that the empirical formulas for heat exchange at the conduits walls are inappropriate. Exact knowledge of the turbulent friction law in a conduit is necessary to test whether the applied heat transfer law is appropriate or not. A thorough calibration of the conduit roughness and an appropriate roughness description (Manning, Darcy-Weissbach or others) for outburst floods is therefore mandatory to improve existing models. It is also necessary to test the theory by laboratory experiments. In particular, it is important to test if the applied laws for the turbulent friction and the heat transfer are appropriate or not. One should also verify if the assumption of constant conduit geometry with time is correct or not. Finally, the question whether the straight-channel theory is appropriate or not waits to be answered.

To achieve the goals the following data should be collected or prepared from case studies:

- Radio-echo-soundings to obtain the glacier bed.
- Photogrammetry of aerial photographs to obtain the present surface topography.
- Photogrammetry of aerial photographs taken before and after lake drainage to derive the lake volume.
- Monitoring the evolution of water-level and temperature in the lake.
- Recording the flood hydrograph and water temperature at the exit from the glacier.
- Monitoring the local horizontal strain rates at the glacier surface in the vicinity of the lake during lake filling and drainage with an automatic theodolite and GPS.
- Monitoring the internal ice deformation with permanently installed tilt sensors.
- Monitoring the vertical strain rates with depth in boreholes.
- Monitoring the englacial water pressure with permanently installed pressure transducers.
- Monitoring deep seismic signals (known as icequakes) during the flood in the ice dam area.
- Conducting tracer experiments before, during and after the flood.

The expertise developed on the glacial lakes of Rocciamelone and of Belvedere Glacier brings to some considerations related to the studies to carry out in order to prevent or mitigate the GLOF hazard. The recommendations for the future are:

- To acquire historical data on the glacier and on possible GLOF of the past (pictures, airphotographs, chronicles, witnessed experiences, unpublished technical reports, scientific publications). All these data are useful to reconstruct the evolutionary trend of the glacier and the lake, if this trend is already present, and to define the occurrence frequency, magnitude, the damage caused by past glacial outburst floods.
  Very important
- To take every year airphotographs of the glaciers affected by epi- or proglacial lakes.
  Useful in order to follow the evolutionary trend of these glaciers and lakes.
  Advisable, expensive.
- To carry out bathimetric survey of the epiglacial lakes in order to define the intensity of the thermokarst processes.
  Advisable, expensive.
- In case of moraine-dammed lakes, to perform geophysical surveys of the moraines to ascertain the presence of dead ice bodies (ice-cored moraines).
  Advisable, expensive.
  To install ablation stakes on the glaciers in order to evaluate the mass balance.
  Advisable, expensive at the beginning, less expensive for the following seasonal measurements.
- To assess the risk due to waves overtopping the dam of the lake in case of ice/rock/snow avalanches impact or of the glacier ice cliffs breaking-off.

Advisable, inexpensive.

In case of emergency is *necessary* to obtain immediate information about the following parameters:

- lake's overall volume and the related level-volumes curve to determine the pumping plant (if this remediation work is settled) and to assess its efficiency;
- the morphology of the lake bottom (by bathimetric survey), to evaluate the optimun position for the pumping station and the presence of possible areas of critical stability within the reservoir;
- the glacier-ice thickness underneath the lake and its immediate areas (by geo-radar survey) to evaluate the stability of the damming ice and the opportunity to undertake a possible drilling of the basin's bottom to promote the subglacial drainage;
- a map of the subglacial discharge route (by tests with tracers), with the aim to investigate the possible connection with streams and springs below the glacier.

The high risk of GLOF directed the italian partners' efforts to develop studies and promote factive actions in cooperation with public administrators, National Civil Protection, research organizations.
As a matter of fact these two glaciers became exceptional field laboratories that should be followed in the future, even after the end of the Glaciorisk Project. Moreover the Belvedere Glacier is still extraordinarily dynamic and will likely cause hazard situations in the next years. In the same way the epiglacial Rocciamelone lake seems to become still larger and deeper and therefore will endanger the reservoir located in the valley below.
<u>For these reasons a continue survey of these two glaciers is highly recommandable</u>. Hopefully the gained experience and the historical and experimental data will not get lost.

2. Stability of steep glaciers and ice avalanches

Alpine glaciers can be the source of spectacular ice avalanches when parts of them break off. Although ice avalanches are relatively rare, in some cases they do threaten settlements and other fixed installations. Several ice avalanches consiting of more than 10^6 m^3 of ice have occured in the Alps; the destructive power of an ice avalanche increases during winter when snow may be entrained by the flow. Particularly destructive events may be generated when rock and/or water is entrained into the flow; such mixed-flow avalanches can run long distances.

The most destructive release of ice on record occured in the Peruvian Andes in 1962 and 1970. In 1962, a large snow and ice avalanche from Mount Huascaran traveled 16 km into the Santa Valley, destroying 9 small towns and killing more than 4'000 people. In 1970, an earthquake triggered another ice avalanche from near the summit of Huascaran (Patzelt, 1983, Lliboutry, 1975; Plafker and Eriksen, 1978). Total vertical descent was more than 4'000 m and the estimated volume (10^7-10^8 m^3) was about 10 times more than the 1962 disaster. The city of Yungay was completely buried by the avalanche and more than 18'000 people were killed. In both cases, rock and soil entrained in the avalanche added to the destruction. A similar process was the cause of the recent impressive and catastrophic event in Dzimarai-Khokh (Caucasus), in which many peoples were killed (Kaeaeb et al., 2003). The largest known release of ice (5 10^6 m^3) in the Alps occured on 11 September 1895 from the summit region of Altels (Bernese Alps, Switzerland). It is generally thought that the cause of the release was a weakening of the contact between the ice/rock interface, possibly as a result of a frozen zone near the snout of the glacier which was gradually thawing during several exceptionally warm summers prior to the event. The probable failure process was a reduction of support at the bed, which initiates tensile failure at the top and more complex fracturing at the other boundaries (Röthlisberger and Kasser, 1978; Röthlisberger, 1981; Raymond et al., 2003). The most destructive recent event occured in Valais (Switzerland) on 30 August 1965, when a major portion of the terminus of Allalingletscher (2 10^6 m^3) slide down a rock slope of about 27°, claiming 88 victims at the Mattmark dam construction site. Subsequent investigations showed that the avalanche released during a periof of enhanced bed-slip that started 2-3 weeks earlier. Similar seasonal changes in speed have been observed in the years after the catastrophe, and it is now known that Allalingletscher speeds up regularly every 1-3 years. Enhanced sliding usually starts in summer or late autumn and ends at the beginning of winter but in most cases no large release of ice happens; evedently the speed-up event is necessary but not sufficient to cause breaking off. It is likely that the convex bed topograpphy and a critical mass distribution also contributed to the Mattmark catastrophe.This phenomena leads to the greatest known falling ice volumes.

Snow avalanches or mudflows can be triggered by ice avalanches and devastate areas far away from the glaciers. Experience shows that the larger the ice avalanche, the smaller is the probability of its occurrence. Very large events have a return period of the order of one or two generations. The perception by the local inhabitants of the dangers inherent to these glaciers is often missing. Furthermore, due to climate variations, some hanging glaciers undergo rapid changes leading to isolated catastrophic events or new situations, for which an historical background is missing. For these reasons inhabitants of endangered zones show a psychological barrier to recognize or acknowledge an immediate danger. On the other hand, experts may be wrong with their interpretation of observations, which is largely based on previous experiences and intuition. In this paper we discuss the stability of two types of glaciers: hanging glaciers and steep glacier tongues.

Hanging glaciers are a fascinating class of small glaciers characterized by the contrast of their wild beauty and the dangers inherent to ice fall. The typical feature of hanging glaciers is the absence of ablation replaced by a mechanical release process of ice mass. The second type, steep glacier tongues, concerns alpine glaciers ending in a steep (25 degrees or more) slope.

Direct measurements of a threatening hanging glacier are usually sparse, not representative and difficult to interpret. Most of the time, these measurements are carried out after clear signs of destabilization have been observed. The knowledge of the conditions prevailing before an unstable state is therefore rare. The factors responsible for the destabilization of large ice masses are the mechanical behavior of ice and the applied loads in the zone of fracture. The physics of the ice behavior and the feedback mechanisms between ice deformation and load repartition are complex and yet mostly unknown. Moreover measurements are difficulty to perform on these steep, heavily crevassed and avalanche endangered glaciers. Lack of theory and efficient measurements make an accurate assessment of the stability of hanging glaciers difficult. In this paper a thorough presentation of recent investigations on some hanging glaciers and steep glacier tongues located in the Alps is presented to understand the physical processes governing the dynamics of hanging glacier and their complex relations. Mechanisms leading to rapid changes in the stability of the glaciers are identified. The occurrence of the ice fall are analyzed. Empirical methods are applied to evaluate the danger of ice avalanches and to forecast the failure time.

Classification

Numerous case histories from Switzerland have been used to identify typical morphological features of zones where glacier break off and relationships between the volume, velocity and runout distance of ice avalanches. It is usual to distinguish between two main types of failure processes and topography of the bedrock (Haefeli, 1965): the wedge failure and the slab failure process.

- Hanging glaciers typically flow over a moderately inclined bed ($< 10^{o}$) and end in a frontal ice cliff near where the slope angle increases. Wedges of ice tend to break off when the cliff steepens or becomes overhanging. The basal ice temperature is either temperate or cold. The unstable ice mass (lamella) separates progressively from the glacier by the formation of a frontal crevasse due to traction forces acting at the terminus part of the glacier. The lamella tilts progressively downstream. When the ice cliff becomes too steep or even overhanging, the lamella breaks off. The typical volume amounts to 10^3-10^5 m^3 .
- In contrast, much larger volume release (typically 10^5-10^6 m^3) from ramp-type glaciers. In these cases there is no significant change in slope and the ice fails as a slab. No release has yet been observed where the bed slope is less than 25°. Three types of ramp glaciers can be found:
    1. Ramp hanging glaciers (e.g. Weisshorn),
    2. Ramp glaciers (e.g. Altels) and

3. Steep glacier tongues (e.g. Allalin).

The unstable ice mass can be the totality of the glacier, or a large part of the glacier tongue. The instability is primarily due to the loss of shearing forces at the glacier base and is therefore strongly related to the bed slope. In case of temperate basal ice, the unstable ice slips over the bed. . For cold basal ice, a shearing fracture propagates parallel to the bed. Here the bedrock slope is steeper but does not exceed 45°. Alean (1985) observed a dependence of the critical slope (the slope for which the glacier is unstable) with the altitude: the higher the altitude, the steeper is the critical slope. This dependence is related to the correlation of the ice temperature (especially the basal ice temperature, i.e., the basal adhesion of the glacier) with altitude.

This classification is based on a two dimensional representation of the glacier morphology. In reality, subprocesses of failure related to the third dimension exist. The major ones are the arched structure of the glacier front and the concavity or convexity of the bedrock (Ott, 1985). These subprocesses can interfere in the main fracture process.

In the Alps, only a limited number of hanging glaciers have been investigated so far, some of them after a catastrophic event, others within the context of a consulting work or a research program. Some of them, which have been more carefully investigated, are the following:

Eiger

This hanging glacier is situated between 3'200 and 3'500 m above see level on the west-facing slope of Eiger (Bernese Alps, Switzerland). In 1993, Lüthi and Funk (1997) determined the bed topography and the ice thickness by carrying out radio echo soundings along several profiles across the glacier. Boreholes were drilled to the glacier bed and the englacial temperature distribution was measured with thermistors, and surface velocities were measured with stakes. They found that the glacier is temperate except near the front and concluded that the ice temperature is an important factor for the global stability of the glacier. In 1990 a large crevasse behind the glacier front could be observed. Three stakes surrounded with prismas were installed on the lamella and their position regularly surveyed using a theodolite laser-distometer. By applying an hyperbolic function to fit the measurements, the forecasted time for breaking off $t\infty$ was estimated. The ice fall of 10^5 m^3 occurred on August 20, three days after the prediction. In 2001, the position of a smaller lamella was measured until failure with a higher temporal resolution. Because of disaggregation of the unstable ice mass, ice blocks began to fall some days before the predicted time estimated on August 18. The failure process covered an entire week.

Gutzgletscher

In September 1996 an important ice fall (2 10^5 m^3) occured on Gutzgletscher (Bernese Alps) and dropped down a 1000 m high rock face. The avalanche debris blocked a road and the air pressure injured three people (Margreth and Funk, 1999). Three years later, a dangerous situation could be recognized at time and a stake network was installed on the glacier to perform velocity mesurements. A breaking off could be predicted one week in advance within one day accuracy. The event occurred on August 14.

Mönch

The hanging glacier of the Mönch south face is also located in the Bernese Alps. It is situated between 4100 m and 3600 m. In 2001 the glacier bed was derived from the depth of boreholes drilled to the bed. Temperature records in the borholes showed that the glacier is temperate. The glacier is reposing on a large flat bedrock terrace. The detachment of two lamellas (in 2000 and 2001) were monitored by repeated surveying of stakes. In both cases the lamella broke off in several pieces. Note the abrupt velocity decrease after a partial fracture. Three partial breaks where observed in the year 2000. The volume of each ice fall amounted to 45 10^3 m^3 , 35 10^3 m^3 and 30 10^3 m^3 . The remaining part of the lamella (60 10^3 m^3) desegregated progressively (VAW, 1997b and VAW, 2000). The ice volume of the two recorded ice falls in 2001 were of the order of magnitude of 10^3 m^3 . They broke off from a a 10^5

m^3 large lamella, which disaggregated progressively during one year. According to observations over the past 20 years, only a few lamella fell in one major ice fall (VAW, 1997b) and further observations by the authors). The majority of the lamella disaggregated in many successive partial ice falls. No observations allowing to identify the evolution of the fracture process.

Allalin

The Allalin glacier is a temperate glacier in the Vallese alps (Switzerland). Its frontal part is formed of a tongue located on a 26° steep (middle value) sleek bedrock (Fig. 8a). In 1965 the destabilization of the tongue killed 88 people, who were working on a construction site. According to Röthlisberger (1981), this event was primary due to an important decrease of the friction forces at the glacier base. This modification of the friction conditions led to a rapid advance of the tongue and the overlapping of a terrace bedrock, which contributed to the stability of the glacier front. Then the abutment on the margin of the glacier and the retain on the upper stable glacier zone progressively failed and led to the collapse of the frontal part of the glacier tongue. No important precursor signs were observed before the breaking off. Only the fall of some small ice blocks have been reported the day of the drama. The volume of the ice avalanche amounted to approximatively 1 million m^3. Until present the global behavior (velocity, geometry) of the glacier has been surveyed. Only in summer but not every years an acceleration (active phase) of the glacier tongue has been observed. Röthlisberger (1981) concluded that this acceleration is not only due to the increase of the lubrication due to melt water, but also to a dynamic effect of mass repartition along the longitudinal profile of the glacier. During a summer, if an acceleration occurs, the ice thickness at the upper part of the unstable zone is decreasing, since the glacier is longitudinally stretched. The following summer, if the frontal part is stable (e.g. leaning on a terrace bedrock), it retains the glacier tongue and the thin zone is filled with the upper flowing ice. This filling process occurs without acceleration and ends when the zone, who was thinner, becomes heavy enough to move the whole tongue (typically after one or two years). The active phase can be accompanied by ice falls. During an active phase, the velocity of the unstable ice mass develops gradually. The velocity ratio between active and quiescent phases is of the order of hundred. In 1999, since the glacier configuration was similar to 1965, before the ice fall, the hazard zone was closed during summer. An ice volume of $1{,}6\ 10^5\ m^3$ fell, but did not cause any damage.

The first category concerns **steep** (20^{o} or more) **glacier tongues**, like Allalingletscher. The breaking off of such large ice masses (1 million m^3 or more) is very rare. But it has been recognized that such events occur during an active phase with enhanced sliding motion, which usually lasts for 1-2 weeks prior to the event. Such an active phase only occurs during the melt season or shortly after its end. The experience on Allalingletscher showed that the breaking off of large ice masses only occurs in very rare cases during such active phases.

The mechanisms of such break-off of large ice masses remain unclear. It is also not known, why some steep glacier tongues switsch in a so-called active phase (e.g. Allalingletscher) during the summer season and other glacier tongues do not (e.g. glacier de Giétro).

To obtain more insights in the processes governing such events, field studies on such steep glacier tongues are necessary:

1. Monitoring the surface velocity field of the tongue with remote sensing techniques (photogrammetry, laser-scanning).
2. Monitoring the water discharge exiting the glacier tongue.
3. Monitoring the water turbidity of the proglacial water stream.

The progressive increase of the flow velocities due to enhanced sliding motion has to be analyzed in detail and explained. For future hazard assessment strategies, the following questions have to be answered :

1. Why do certain glacier tongues switch in an active phase during the summer season and others do not ?
2. What are the causes for the sudden detachement of large ice masses during an active phase and why this happens only in very rare occasions ?

3. Reliable observable criterion have to be defined, which allow to predict or at least indicate that a large breaking off event must be expected in a particular case.

The second category concerns **hanging glaciers**. Observations show that ice masses become detached from such glaciers by progressive fracture at englacial interfaces. Such events occur all around the year. The main difference between hanging glaciers and steep glacier tongues is that for hanging glaciers a progressive motion increase of an instable ice mass always leads to a major break off. For steep glacier tongues this happens only in very rare situations. Results from combined field and modelling studies on hanging glaciers show that a forecast of a major breaking off event is possible in some situations. However, it is still very difficult to predict in advance the breaking off volume. In most cases, an instable ice mass breaks off in many smaller chunks of ice. But even a small volume of falling ice can trigger a huge combined ice/snow avalanche if a thick and unstable snow pack exists around a hanging glacier. Such events can be relevant in the winter season and should be seriously considered by people responsible for the security of roads and railways in regions where hanging glaciers exist.
To improve forecasting possibilities of breaking off of instable ice masses, it is necessary to monitor the regular formation and detachment of ice chunks on hanging glaciers by:

1. Monitoring the breaking off activities at the front of hanging glaciers with regular high quality photographs, and
2. Monitoring the evolution of the motion of the unstable ice mass from the beginning of the instability to the time of breaking off.

For hanging glaciers located in regions where the basal ice temperature is close to the melting point, their stability can change in near future because of global warming. This is because the transition of a previously cold to a temperate glacier base, basal sliding can suddenly be induced, leading to a major destabilisation of the hanging glacier. Where such an evolution is expected to happen and if human activities are endangered, a monitoring program for these type of hanging glaciers is extremely important for hazard assessments.

Also the Croce Rossa Glacier <u>would require a permanent monitoring</u> in order to evaluate the glacier stability and future scenarios in case of outburst floods and of ice avalanches.
The glacier overhangs the Lago della Rossa reservoir. Because flood waves generated by an ice fall may overtop the dam, the lake level was kept very low in the last years. A complete survey has been carried out since 1998. A detailed analysis of the stability conditions and a permanent monitoring of the relevant parameters (e.g. basal ice temperatrures, surface velocities, crevasse formation) would be a usefull contribution for using the reservoir being aware of the hazard.

References

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

Anderson, S. P., Walder, J. S., Anderson, R. S., Kraal, E. R., Cunico, M., Fountain, A. G., and Trabant, D. (2003). Realtime hydrologic observations of Hidden Creek Lake jökulhlaups, Kennicott Glacier, Alaska. *J. Geophys. Res.* in press.

Bird, R. B., Stewart, W. E., and Lightfoot (1960). *Transport Penomena*. John Wiley, New York.

Björnsson, H. (1974). Explanation of jökulhlaups from Grímsvötn, Vatnajökull, Iceland. *Jökull*, 24:1–24.

Björnsson, H. (1992). Jökulhlaups in Iceland: prediction, characteristics and simulation. *Annals Glaciol.*, 16:95–106.

Björnsson, H. (2002). Subglacial lakes and jökulhlaups in Iceland. *Global and Planetary Change*, 35:255–271.

Broecker, W. S. and Denton, G. H. (1990). What drives glacial cycles? *Scientific American*, 262:43–50.

Clague, J. J. and Mathews, W. H. (1973). The magnitude of Jökulhlaups. *J. Glaciol.*, 12(66):501–504.

Clarke, G. C. K. (2002). Hydraulics of iceage megafloods. *Geophysical Research Abstracts, contributions to the 27th General assembly of the EGS*, 4. Abstract.

Clarke, G. K. C. (1982). Glacier outburst floods from “Hazard Lake”, Yukon Territory, and the problem of flood magnitude prediction. *J. Glaciol.*, 28(98):3–21.

Clarke, G. K. C. (2003). Hydraulics of subglacial outburst floods: new insights from the SpringHutter formu-

lation. *J. Glaciol.*
Fowler, A. C. (1999). Breaking the seal at Grímsvötn. *J. Glaciol.*, 45(151):506–516.
Fowler, A. C. and Larson, D. A. (1978). On the flow of polythermal glaciers: I. Model and preliminary analysis. *Proc. R. Soc. London, Ser. A*, 363(1713):217–242.
Fowler, A. C. and Ng, F. S. L. (1996). The role of sediment transport in the mechanics of jökulhlaups. *Annals Glaciol.*, 22:255–259.
Gudmundsson, M. T., Sigmundsson, F., and Björnsson, H. (1997). Icevolcano interaction of the 1996 Gjálp subglacial eruption, Vatnajökull, Iceland. *Nature*, 389(6654):954–957.
Haeberli, W. (1983). Frequency and characteristics of glacier floods in the Swiss Alps. *Annals Glaciol.*, 4:85–90.
Haeberli, W., Kääb, A., Paul, F., Chiarle, M., Mortara, G., Mazza, A., Deline, P., and Richardson, S. (2002). Surgetype movement at Ghiacciaio del Belvédère and a developing slope instability in the east face of Monte Rosa, Macugnaga, Italian Alps. *Norwegian Journal of Geography*, 56:104–111.
Jóhannesson, T. (2002a). The initiation of the 1996 jökulhlaup from Lake Grímsvötn, Vatnajökull, Iceland. In *The Extremes of the Extremes: Extraordinary Floods*, volume 271 of *IAHS Publ.*, pages 57–64. International Association of Hydrological Sciences.
Jóhannesson, T. (2002b). Propagation of a subglacial flood wave during the initiation of a jökulhlaup. *Hydrological SciencesJournaldes Sciences Hydrologiques*, 47(3):417–434.
Liestøl, O. (1956). Glacier dammed lakes in Norway. *Norsk Geografisk Tidsskrift*, 15:122–149.
Haefeli, R. (1965). Note sur la classification, le mécanisme et le contrôle des avalanches de glaces et des crues glaciaires extraordinaires. In *Extrait de la publication de l'A.I.H.S no. 69*, pages 316–325, Davos. Symposium International sur les Aspects Scientifiques des Avalanches de Neige.
Iken, A. (1977). Movement of a large ice mass before breaking off. *Journal of Glaciology*, 19(81):565–605.
Kääb, A., Wessels, R., Haeberli, W., Huggel, C., Kargel, J. S., and Khalsa, S. J. S. (2003). Rapid ASTER imaging facilitates timely assessment of glacier hazards and disasters. *EOS*, 84(13):117 and 121.
Lliboutry, L. (1975). La catastrophe de Yungay (Péru). *IAHS-AISH Publikation*, 104:353–363. Snow and Ice Symposium (Proceedings of the Moscow Symposium, August 1971).
Lüthi, M. (1994). Stabilität steiler Gletscher: Eine Studie über den Einfluss möglicher Klimaänderungen; Untersuchungen am Beispiel eines Hängegletschers in der Westflanke des Eigers, Diplomarbeit an der VAW/ETH-Zürich (unveröffentlicht).
Lüthi, M. and Funk, M. (1997). Wie stabil ist der Hängegletscher am Eiger ? *Spektrum der Wissenschaft*, 5:21–24.
Maag, H. U. (1972). Icedammed lakes on Axel Heiberg Island, with special reference to the geomorphological effect of the outflowing lake water. In *Axel Heiberg Island Research Reports, Miscellaneous papers*, pages 39–48. McGill University, Montreal.
Margreth, S. and Funk, M. (1999). Hazard mapping for ice and combined snow/ice avalanches -two case studies from the Swiss and Italian Alps. *Cold Regions Science and Technology*, 30:159–173.
Moore, J. C., Pälli, A., Ludwig, F., Blatter, H., Jania, J., G¸adek, B., Głowacki, P., Mochnacki, D., and E, I. (1999). High resolution hydrothermal structure of Hansbreen, Spitsbergen mapped by ground penetrating radar. *J. Glaciol.*, 45(151):524–532.
Ng, F. (1998). *Mathematical modelling of subglacial drainage and erosion*. PhD thesis, St Catherin's College, Oxford.
Ng, F. S. L. and Björnsson, H. (2003). On the ClagueMathews relation for jökulhlaups. *J. Glaciol.*
Nye, J. F. (1976). Water flow in glaciers: jökulhlaups, tunnels and veins. *J. Glaciol.*, 17:181–207.
Ohmura, A. (2001). Physical basis for the temperaturebased meltindex method. *Journal of Applied Meteorology*, 40:753–761.
Ott, B. (1985). Effets de voûte dans les glaciers. Mitteilung 80, Versuchsanstalt für Wasserbau, Hydrologie und Glaziologie der ETH Zürich.
Patzelt, G. (1983). *Die Berg-und Gletscherstürtze vom Huascaran, Cordillera Blanca, Peru*. Universitätsverlag Wagner, Innsbruck. Mit Beiträgen von W. Hofmann, H. Körner, E. Schneider, J. Stadelmann und W. Welsch.
Plafker, G. and Ericksen, G. E. (1978). Nevados huascarán avalanches, peru. In Voight, B., editor, *Rockslides and Avalanches, 1, Natural Phenomena*, chapter 8, pages 277–314. Elsevier, Amsterdam.
Pralong, A., Funk, M., and Lüthi, M. P. (2003). A description of crevasse formation using continuum damage mechanics. *Annals of Glaciology*, 37. in press.
Raymond, C. F. and Nolan, M. (2000). Drainage of a glacial lake through an ice spillway. In *Debriscovered glaciers*, volume 264, pages 199–207.
Raymond, M., Wegmann, M., and Funk, M. (2003). Inventar gefährlicher Gletscher in der Schweiz. Mitteilung 182, Versuchsanstalt für Wasserbau, Hydrologie und Glaziologie der ETH Zürich.
Rist, S. (1955). Skeiðarárhlaup 1954, The jökulhlaup of Skeidara. *Jökull*, 5:30–36.
Roberts, M. J., Russell, A. J., Tweed, F. S., and Knudsen, O. (2000). Ice fracturing during jökulhlaups: implications for englacial floodwater routing and outlet development. *Earth Surface Processes and Landforms*,

25:1429–1446.
Röthlisberger, H. (1972). Water pressure in intra– and subglacial channels. *J. Glaciol.*, 11(62):177–203.
Röthlisberger, H. (1977). Ice avalanches. *Journal of Glaciology*, 19(81):669–671.
Röthlisberger, H. (1981). Eislawinen und Ausbrüche von Gletscherseen. In *P. Kasser (Ed.), Gletscher und Klima glaciers et climat, Jahrbuch der Schweizerischen Naturforschenden Gesellschaft, wissenschaftlicher Teil 1978*, pages 170–212. Birkhäuser Verlag Basel, Boston, Stuttgart.
Röthlisberger, H. (1987). Sliding phenomena in a steep section of Balmhorngletscher, Switzerland. *Journal of Geophysical Research*, 92(B9):8999–9014.
Röthlisberger, H. and Kasser, P. (1978). The readvance of the Allalingletscher after the ice avalanche of 1965. In *Proc. Int. Workshop on Mechanism of Glacier Variations, 30.9.-11.10.1976*, Alma-Ata. Materialy Glyatsiologicheskikh Issledovaniy, Khronika, Osuzhdeniya, 33:142-164.
Schuler, T. (2002). *Investigation of water drainage through an alpine glacier by tracer experiments and numerical modeling*. PhD thesis, ETH Zürich.
Shreve, R. L. (1972). Movement of water in glaciers. *J. Glaciol.*, 11(62):205–214.
Spring, U. (1980). Intraglazialer Wasserabfluss: Theorie und Modellrechnung. Mitteilung 48, Versuchsanstalt für Wasserbau, Hydrologie und Glaziologie der ETH Zürich, Gloriastrasse 3739, ETHZentrum, CH8092 Zürich.
Spring, U. and Hutter, K. (1981). Numerical studies of Jökulhlaups. *Cold Regions Science and Technology*, 4:221–244.
Spring, U. and Hutter, K. (1982). Conduit flow of a fluid through its solid phase and its application to intraglacial channel flow. *International Journal of Engineering Sciences*, 20:327–363.
Szilder, K., Lozowski, E. P., and Sharp, M. J. (1997). Glacial lake drainage: a stability analysis. *Annals Glaciol.*, 24:175–180.
Tweed, F. S. and Russell, A. J. (1999). Controls on the formation and sudden drainage of glacierimpounded lakes: implications for jökulhlaup characteristics. *Progress in Physical Geography*, 23(1):79–110.
VAW (1997b). Hängegletscher Mönch-Süd, Gutachten zur Eislawinenproblematik. Im Auftrag der Jungfraubahnen, (M. Funk, unveröffentlicht).
VAW (2000). Hängegletscher Mönch-Süd, Bericht über die Messungen und Beobachtungen im Sommer 2000. Im Auftrag der Jungfraubahnen, (M. Funk unveröffentlicht).
Walder, J. and Costa, J. (1996). Outburst Floods from GlacierDammed Lakes: The Effect of Mode of Lake Drainage on Flood Magnitude. *Earth Surface Processes and Landforms*, 21(8):701–723.
Waller, R. I., Russel, A. J., van Dijk, T. A. G. P., and Knudsen, Ó. (2001). Jökulhlauprelated ice fracture and supraglacial water release during the November 1996 jökulhlaup, Skeidarárjökull, Iceland. *Geografiska Annaler*, 83A(1–2):29–37.
Weigand, B. and Beer, H. (1993a). Iceformation phenomena for water flow inside a cooled parallel plate channel: An experimentaland theoretical investigation of wavy ice layers. *Int. J. Heat Mass Transfer*, 36:685–693.
Weigand, B. and Beer, H. (1993b). A theoretical and experimental investigation of smooth and wavy ice layers in laminar and turbulent flow inside an asymmetrically cooled parallel plate channel. *Wärmeund Stoffübertragung*, 29:27–36.