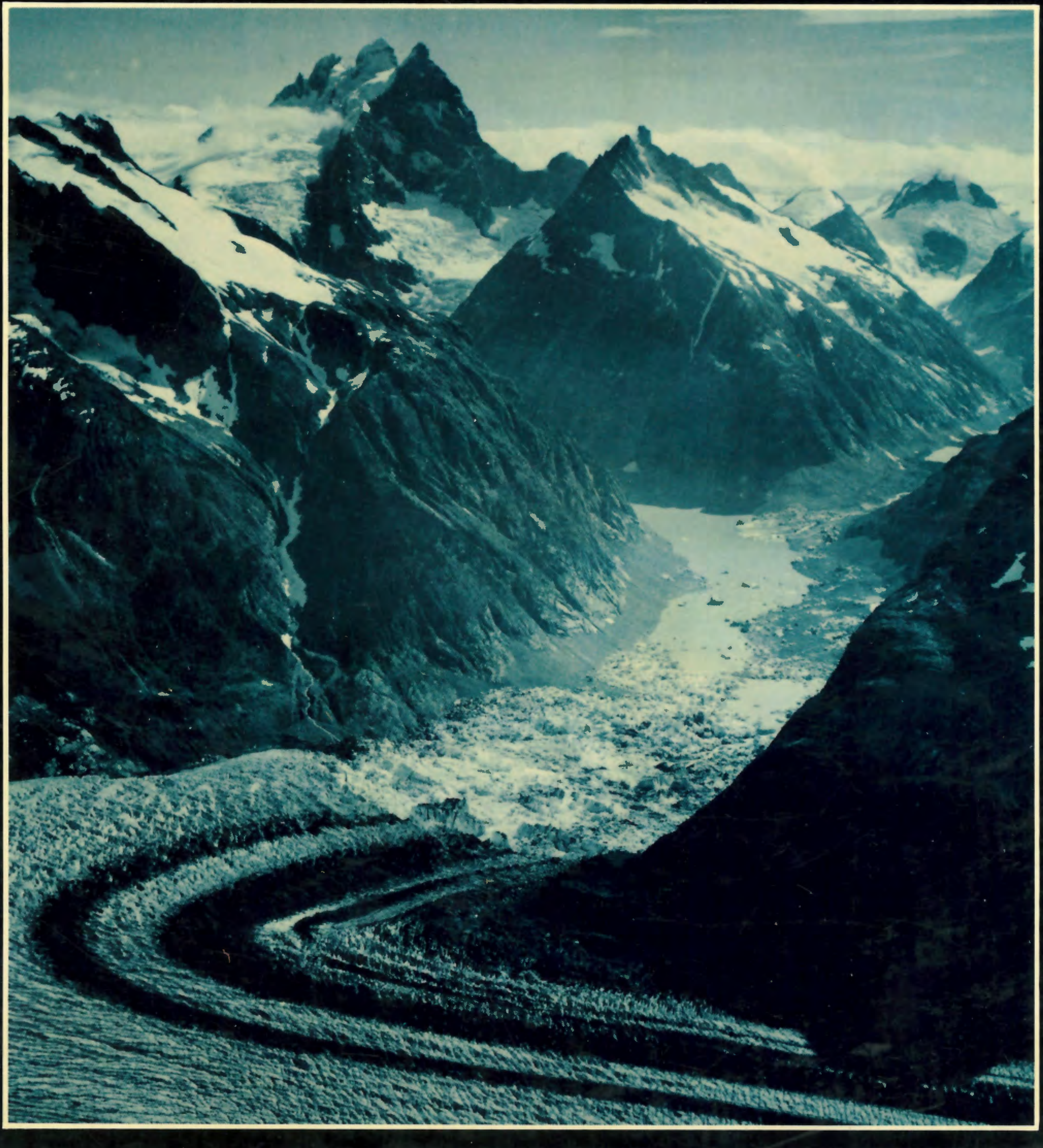
Destructive Mass Movements in High Mountains: Hazard and Management



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COVER

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Destructive Mass Movements in High Mountains: Hazard and Management

Abstract

The remote valleys and high mountain areas of western Canada have experienced accelerated economic development only in recent years. Although there have been occasional mass movements of destructive impact in this region in the past, the short history of human settlement makes it difficult to properly evaluate the potential hazard and appropriate countermeasures during future development of steep mountainsides and narrow valleys. A large body of documented experience from long-inhabited mountain regions, notably the Alps of Europe, suggests that major mistakes can be avoided if this experience is used properly.

Most destructive mass movements are complex and related to mountain torrent systems and instabilities of steep bedrock slopes. They can be conveniently grouped into (1) debris flows from surficial deposits, (2) debris flows from bedrock failures, (3) mass movements on volcanoes, (4) glacier-related mass movements, and (5) rockfalls and rock avalanches. Each of these requires a different type of hazard appraisal and particular set of remedial or preventive measures. The range of possible destructive impacts and appropriate countermeasures are illustrated with 137 case histories from the Alps which are presented in a coherent geological-climatic framework transcending the traditional political boundaries. Analysis of these case histories, combined with insights gained elsewhere, is used to outline the application of active measures (forestry, control works, protective works), passive measures (zoning, planning), monitoring, and acceptance of risk. The decisions regarding which active or passive measures should be applied and what level of risk is acceptable are based on (1) information on recurrence and magnitude of mass movements and (2) a broad social-economic consensus.

Résumé

Dans les vallées éloignées et les hautes régions montagneuses de l'Ouest canadien, l'expansion économique ne se poursuit à un rythme accéléré que depuis ces dernières années. Même si, à l'occasion, des mouvements de masse ont fait sentir leurs effets destructifs dans ces régions à une époque fort ancienne, la courte histoire de la colonisation humaine rend difficile toute évaluation précise des risques éventuels et des mesures préventives appropriées à prévoir au cours des prochaines phases de formation de versants escarpés et de vallées étroites. Cependant, tout un ensemble d'expériences notées, acquises dans les régions montagneuses depuis longtemps peuplées, comme les Alpes d'Europe, permettent de croire que de graves erreurs pourraient être évitées en mettant à profit ces expériences.

Les mouvements de masse les plus destructeurs sont de nature complexe et sont liés aux réseaux de torrents de montagnes et à l'instabilité des pentes abruptes de la roche en place. Ces mouvements peuvent, à toutes fins pratiques, être groupés dans l'une des catégories suivantes: (1) coulées de débris provenant de dépôts de surface, (2) coulées de débris générés par des éboulements rocheux, (3) mouvements de masse sur les volcans, (4) mouvements de masse liés aux glaciers, et (5) glissements d'éboulis rocheux et avalanches de pierres. Chaque catégorie de mouvement nécessite un mode différent d'évaluation des risques et un jeu particulier de mesures correctives ou préventives. La gamme possible des effets destructeurs et les mesures préventives appropriées est illustrée par 137 cas survenus dans les Alpes et est présentée dans un cadre climatique et géologique cohérent qui dépasse les frontières politiques traditionnelles. Une analyse de ces cas ainsi que le savoir acquis ailleurs servent à déterminer l'application de mesures actives (foresterie, travaux de surveillance, travaux de protection), de mesures passives (zonage, planification), de surveillance et d'acceptation des risques. Les décisions quant au choix entre les mesures actives et passives et à la détermination de niveaux de risque acceptables sont prises en fonction (1) de l'information fournie sur la fréquence et l'importance des mouvements de masse et (2) du consensus qui se fait au niveau socio-économique.

SUMMARY

Accelerated development of remote valleys and high mountain areas in recent years has indicated the need for a better understanding of the phenomenon of mass movement. Although there have been destructive slope failures in this region in the past, their significance has not been fully appreciated because of the relatively sparse population, the short history of settlement, and, until recently, the frontier state of development. In contrast, in many other mountain regions, most notably the Alps of Europe, there is a large body of documented experience with slope failures and mass movements. In this paper, we draw upon this body of knowledge and our own experience as field geologists to summarize the major types of destructive mass movements, their potential impact, and the range of remedial and preventive measures that may be employed to reduce hazard.

Most destructive mass movements are complex and are related to mountain torrent systems and instabilities in steep bedrock slopes. They are controlled by unique combinations of geological, topographic, meteorological, and seismic factors and by human activity (e.g. mining, logging, urban and recreational development). The major groups of mass movements considered in this report are debris flows from surficial deposits, debris flows from bedrock failures, mass movements on volcanoes, glacier-related mass movements, and rockfalls and rock avalanches.

Debris flows. A debris flow is a mixture of water and sediment that moves as a viscous fluid, commonly down a torrent channel. Some flows are merely the extreme stage of fluvial bedload transport and can be referred to as 'debris floods'; others are made up mainly of silt and clay and are better termed 'mudflows'; still others consist entirely of boulders or slabs of bedrock and might be called 'boulder flows'. However, most destructive flows are heterogeneous mixtures of sediment ranging in size from clay to boulders. The most important sources of material for these flows are unconsolidated sediments of Quaternary age (mainly glaciofluvial deposits and colluvium) which commonly cover valleys and mountain slopes. Debris flows result from the mobilization of these deposits by water, generally during high-intensity downpours, sustained regional rainstorms, and rapid snowmelt. Several debris-mobilizing mechanisms operate during these storms, including gullying by running water, soil slips and debris avalanches, and reactivation of pre-existing large slumps and slow-moving earthflows. Once mobilized, the debris may travel long distances down a torrent channel, commonly in a highly unsteady, pulsating manner. However, the debris rapidly decelerates as it leaves its steep confined channel and spreads across the cone or fan at the mouth of the torrent basin. Developed cones and fans are the sites of most of the destruction wrought by debris flows.

PRÉCIS

La mise en valeur accélérée des vallées isolées et des régions de hautes montagnes de ces dernières années a mis en évidence le besoin de mieux comprendre le phénomène des mouvements de masse. Bien qu'il y ait déjà eu des ruptures de pente destructives dans les régions en question, leur signification n'a pas été appréciée à sa juste valeur en raison de facteurs comme la population peu répandue, l'histoire relativement brève du peuplement et jusqu'à récemment, la nature pionnière des activités de mise en valeur. Au contraire, dans beaucoup d'autres régions montagneuses, principalement dans les Alpes en Europe, on a réalisé de nombreuses études des ruptures de pente et des mouvements de masse. Dans le présent rapport, les auteurs auront recours à ces connaissances et à leur propre expérience en tant que géologues de terrain pour résumer les principaux types de mouvements de masse destructifs, leur impact possible et les mesures de prévention et de reconstruction qui peuvent être employées pour limiter les risques.

Les mouvements de masse les plus destructifs sont complexes et liés aux systèmes torrentiels des montagnes et à l'instabilité des pentes rocheuses abruptes. Ils sont gouvernés par des combinaisons uniques de facteurs géologiques, topographiques, météorologiques, sismiques et anthropiques (par exemple, l'exploitation minière, l'exploitation forestière, la mise en valeur de nature urbaine et récréative). Les principaux groupes de mouvements de masse considérés dans le rapport sont les coulées de débris provenant des dépôts de surface, les coulées des débris provenant des éboulements rocheux, les mouvements de masse sur les volcans, et les mouvements de masse liés aux glaciers, les glissements d'éboulis rocheux et les avalanches de pierres.

Les coulées de débris. Une coulée de débris est un mélange d'eau et de sédiments qui se déplace comme un fluide visqueux, habituellement dans un canal d'écoulement de torrent. Certains écoulements, auxquels on peut donner le nom de «coulées de débris», ne sont que le stade extrême de transport de la charge de fond fluviale; d'autres consistent principalement en limon et en argile et s'appelleraient plus justement «coulées de boue»; d'autres encore consistent entièrement en gros blocs ou en dalles de roche et peuvent être appelés «coulées de gros blocs». Toutefois, la plupart des coulées destructives sont des mélanges hétérogènes de sédiments dont la taille varie des argiles aux gros blocs. Les sources les plus importantes de matériau pour ces coulées sont des sédiments non consolidés du Quaternaire (principalement des dépôts glaciofluviaux et des matériaux colluviaux) qui habituellement couvrent les vallées et les versants des montagnes. Les coulées de débris résultent de la mobilisation de ces dépôts par l'eau, généralement au cours de très fortes averses, de gros orages régionaux, ou de la fonte rapide de la neige. Plusieurs mécanismes mobilisant les débris se manifestent au cours de ces orages et produisent du ravinement par l'eau courante, des glissements de sol et des avalanches de débris, ainsi que la réactivation d'anciens grands glissements de matériaux et glissements de terrain lents. Une fois mobilisés, les débris peuvent voyager sur de longues distances dans un canal d'écoulement de torrent, habituellement selon un mode pulsé, fortement inégal. Toutefois, les débris décélèrent rapidement au fur et à mesure qu'ils s'éloignent du chenal raide dans lequel ils étaient restreints et qu'ils s'étalent sur le cône de déjection à l'embouchure du bassin torrentiel. Presque toute la destruction

Debris flows also are generated by bedrock failures. Occasionally, a bedrock failure transforms directly into a debris flow through the incorporation of water into the moving mass. More commonly, bedrock failures block or constrict torrent channels; during subsequent periods of intense precipitation or overflow, debris becomes mobilized from the toe of the failed mass and accelerates down the torrent channel in much the same manner as a flow derived from unconsolidated sediments. This type of debris flow is not restricted to any particular type of bedrock failure, although it is particularly common in areas of deep-seated creep ('sagging') involving entire mountainsides.

The appraisal of debris flow hazard is based on: (1) documentation of historical debris flow events and the resulting damage; (2) hazard indicators found on the fan, in the gorge, and in the source area of the torrent (e.g. large blocks, damaged vegetation); and (3) formulas predicting recurrence and magnitude of extreme flows. Once the hazard has been appraised, 'passive' and 'active' countermeasures may be taken to minimize danger to people and property. Passive measures are those which mitigate hazard without modifying the debris flow system; they include land-use zoning and monitoring. In contrast, active measures modify the system so as to lessen the likelihood of destructive flows or to reduce their impact. Active measures are directed towards the control and stabilization of debris in source areas; on the fan they have primarily protective functions.

Mass movements on volcanoes. Although volcanic mass movements share many characteristics with debris flows and rock avalanches, they are discussed separately in this report because of the extreme hazard they pose in areas of recent volcanism in western North America. The potential magnitude of the problem is indicated by the fact that, worldwide in this century alone, thousands of deaths and enormous property destruction have resulted from mass movements on active and dormant volcanoes.

Volcanic mass movements may be broadly subdivided into two groups, one consisting of failures triggered by eruptions and the other consisting of failures unrelated to eruptions. The most common types in both groups are debris flows and debris avalanches comprising pyroclastic material and blocks of crystalline volcanic rocks. Debris flows and avalanches occurring during eruptions are often hot and of very large size, whereas those unrelated to eruptions are cold and generally smaller.

Cold volcanic debris flows may be further subdivided according to their mode of formation. Some result from the failure of a discrete mass of pyroclastic material and/or lava above a well defined rupture surface. As this slab moves down the flank of the volcano, it rapidly disintegrates into a mélange of blocks and smaller particles. Such debris flows commonly are

occasionnée par des coulées de débris se produit à l'endroit de formation des cônes.

Les coulées de débris sont aussi générées par des éboulements rocheux. Parfois, un éboulement rocheux se transforme directement en coulée de débris sous l'effet de l'incorporation de l'eau dans la masse en mouvement. Plus fréquemment, les éboulements rocheux bloquent ou resserrent les chenaux des torrents; au cours des périodes ultérieures de précipitation intense ou suite au déversement d'un lac, les débris deviennent mobilisés à partir du bord de l'éboulis et accélèrent en descendant le chenal du torrent de la même manière qu'une coulée dérivée de sédiments meubles. Ce type de coulée de débris n'est pas limité à un type particulier d'éboulement rocheux, bien qu'il soit particulièrement répandu dans des régions où la reptation en profondeur («affaisement») de flancs entiers de montagne se manifeste.

Les risques liés aux coulées de débris sont évalués à partir de: (1) la documentation sur les anciennes coulées de débris et les dommages consécutifs; (2) les indicateurs de danger trouvés sur le cône, dans la gorge et dans la région de la source du torrent (par exemple, grands blocs, végétation endommagée); (3) les formules prévoyant la récurrence et l'ampleur des coulées extrêmes. Une fois que le danger a été évalué, des mesures préventives à caractère «passif» et «actif» peuvent être prises pour réduire le risque pour la population et les propriétés. Les mesures passives sont celles qui atténuent le danger sans modifier le système d'écoulement de débris; elles comprennent le zonage et le contrôle de l'utilisation des terres. Au contraire, les mesures actives modifient le système de manière à réduire la probabilité des coulées destructives ou leur effets. Les mesures actives visent le contrôle et la stabilisation des débris dans les zones de source; sur le cône, elles remplissent essentiellement des fonctions protectrices.

Les mouvements de masse sur les volcans. Bien que les mouvements de masse sur les volcans partagent de nombreuses caractéristiques avec les coulées de débris et les avalanches de pierres, le présent rapport en discute séparément à cause des dangers extrêmes qu'ils représentent dans les régions de volcanisme récent dans l'ouest de l'Amérique du Nord. L'étendue potentielle du problème est indiquée par le fait que, à l'échelle mondiale et au cours du dernier siècle seulement, des milliers de morts et des pertes matérielles énormes ont été causées par des mouvements de masse survenus sur des volcans actifs et inactifs.

Les mouvements de masse volcaniques peuvent être subdivisés approximativement en deux groupes, l'un consistant en éboulements causés par des éruptions et l'autre consistant en éboulements non liés aux éruptions. Les types les plus fréquents dans les deux groupes sont les coulées de débris et les avalanches de débris comprenant des matériaux pyroclastiques et des blocs de roches volcaniques cristallines. Les coulées de débris et les avalanches se produisant au cours des éruptions sont souvent chaudes et de très grande taille, tandis que celles non liées aux éruptions sont froides et généralement plus petites.

Des coulées de débris froids d'origine volcanique peuvent être redivisées selon leur mode de formation. Certaines proviennent de l'éboulement d'une masse discontinue de matériaux pyroclastiques ou de lave, ou les deux, au-dessus d'une surface de rupture bien définie. Pendant que cette masse descend les flancs du volcan, elle se désagrège rapidement en un mélange de blocs et de particules plus petites. Ces coulées de débris sont fréquemment occasionnées triggered by earthquakes, by heavy rainfall or glacier melt, or by the swelling of the cone due to upward movement of magma. A second major group of volcanic debris flows (as well as many mudflows) results from the erosion of tephra by running water. Such flows are little different from those involving glacial, colluvial, and other types of unconsolidated sediments in nonvolcanic mountain areas.

Cold debris flows are water-bearing and define one end of the spectrum of volcanogenic flows. At the other end of the spectrum are relatively dry flows consisting of incandescent pyroclastic debris mixed with hot air and volcanic gases ('pyroclastic flows'). Where the gaseous phase dominates, such incandescent flows are termed 'nuées ardentes'. Nuées ardentes have tremendous mobility and may sweep across ridges and spread out over extensive areas at the base of a volcano. Hot debris flows formed by the mixing of volcanic ejecta with large volumes of water are also very mobile, although they generally follow valleys or other topographic depressions. Very large flows, which may travel many tens of kilometres, often result from eruptions through a crater lake or through snow and ice.

Many mass movements caused by volcanic eruptions are so large that active countermeasures are of little or no value. In such instances complete protection is provided only by avoiding potentially hazardous areas. In most cases, however, rigid zoning is impractical for economic or social reasons. Protection of people and property in such instances is dependent largely on monitoring, for if the time and place of an eruption can be forecast, people can be evacuated from the threatened area and some property moved to safety. However, even if a volcanic eruption can be forecast, the timing of mass movements resulting from that eruption may be very difficult to predict. Mass movements may occur at the beginning of an eruption or for some time after the eruption has ended.

Some protection against small debris flows may be provided by diversion dams, retention basins, and elevated artificial structures that serve as refuges. Large water and hydroelectric reservoirs may contain debris flows of moderate to large volume, but care must be taken to lower reservoir levels prior to an eruption in order to prevent displacement waves from overtopping the dam. Finally, active countermeasures employed in nonvolcanic areas may be used on extinct volcanoes to protect against rain-induced debris flows, although an appreciation of the large size of such flows is required before protective works are constructed.

Glacier-related mass movements. Many mass movements in high mountain areas are associated with glaciers and with loose, poorly vegetated debris at their margins. Glacier-related mass movements fall into five major groups: (1) debris flows caused by bursts of ice-dammed and ice-marginal lakes; (2) debris flows resulting from bursts of moraine-dammed lakes;

par des tremblements de terre, de fortes pluies ou la fonte des glaciers et par le gonflement du cône dû à la montée du magma. Un deuxième groupe important de coulées de débris volcaniques (ainsi que beaucoup de coulées de boue) résulte de l'érosion du tephra par les eaux courantes. Il existe très peu de différence entre de telles coulées et celles de sédiments glaciaires, colluviaux et d'autres types de sédiments meubles dans des régions montagneuses non volcaniques.

Les coulées de débris froids contiennent de l'eau et représentent une extrémité du spectre des coulées volcanogéniques. À l'autre extrémité du spectre se trouvent les coulées relativement sèches formées de débris pyroclastiques incandescents mélangés à de l'air chaud et à des gaz volcaniques («coulées pyroclastiques»). Là où la phase gazeuse domine, de telles coulées incandescentes sont appelées «nuées ardentes». Les nuées ardentes sont dotées d'une extrême mobilité et peuvent balayer les crêtes et s'étaler sur de vastes régions à la base des volcans. Les coulées de débris chauds formées par le mélange de projections volcaniques à de larges volumes d'eau sont aussi très mobiles, bien qu'elles suivent généralement les vallées ou les autres dépressions topographiques. De très larges coulées, qui peuvent se déplacer sur plusieurs dizaines de kilomètres, sont souvent le résultat d'éruptions à travers un lac de cratère ou de la glace et de la neige.

Beaucoup de mouvements de masse causés par des éruptions volcaniques sont si considérables que des mesures de prévention actives ont peu, sinon aucune valeur. Dans ces cas, une protection complète est assurée seulement en évitant les régions de danger possible. Dans la plupart des cas, toutefois, le zonage rigide s'avère peu pratique pour des raisons économiques ou sociales. La protection des habitants et des propriétés dans de tels cas dépend largement du contrôle de la situation, car si le moment et l'emplacement d'une éruption quelconque peuvent être prévus, il est possible d'évacuer la population de la région menacée et de déplacer certains biens. Toutefois, même si l'on peut prévoir une éruption volcanique, le moment précis des mouvements de masse résultant de cette éruption peut être difficile à prévoir. Les mouvements de masse peuvent se produire au début ou peu après la fin de l'éruption.

Une certaine protection contre de petites coulées de débris peut être assurée par des barrages de diversion, des bassins de rétention et des structures artificielles élevées qui servent de refuges. On doit prendre soin, avant une éruption, de baisser le niveau des vastes réservoirs d'eau et hydro-électriques susceptibles de contenir des coulées de débris d'un volume modéré à important, afin d'éviter le débordement des vagues de déplacement liées au remplissage du barrage par les coulées. Finalement, des mesures de prévention actives employées dans les régions non volcaniques peuvent être utilisées sur les volcans éteints pour se protéger des coulées de débris provoquées par les pluies, bien qu'il soit nécessaire d'évaluer l'ampleur d'une telle coulée avant d'entreprendre des ouvrages de protection.

Les mouvements de masse liés aux glaciers. Beaucoup de mouvements de masse dans les hautes montagnes sont associés aux glaciers et aux débris à faible couverture végétale en bordure des glaciers. Les mouvements de masse liés aux glaciers se répartissent en cinq groupes principaux: (1) les coulées de débris causées par l'éclatement des lacs de barrage glaciaire et des lacs de front glaciaire; (2) les coulées de débris résultant de l'éclatement des

(3) debris flows resulting from bursts of subglacial and englacial water pockets; (4) failures of glacier ice (ice avalanches); and (5) failures of bedrock and superincumbent glacier ice (ice-rock avalanches).

Active countermeasures may be employed to reduce the hazard of some of these types of mass movements. Ice-dammed lakes and in some instances subglacial and englacial water pockets can be drained by tunnelling through rock or ice or by trenching across the surface of the glacier. Where this is impractical, debris retention dams or flood basins can be constructed below the ice dams, although to be effective these must be of large size and therefore are expensive. Alternatively, monitoring of both the lake and the ice dam may provide adequate warning of an impending burst. The most effective active measures against debris flows from a moraine-dammed lake are: (1) the preparation of a deep and stable cut through the moraine prior to the formation of a major lake; and (2) the construction of revetments along the outlet channel to prevent entrenchment and collapse of embankments during periods of enhanced overflow.

Rockfalls and rock avalanches. Rockfalls and rock avalanches are masses of broken rock that move rapidly in a tumbling or streaming fashion. They develop most commonly on steep slopes underlain by fractured, but otherwise competent rock formations, mainly thick-bedded sedimentary rocks and massive igneous and high grade metamorphic rocks. In contrast, they occur rarely on slopes underlain by thin-bedded sedimentary rocks and low grade metamorphics. Rock avalanches result from the rapid detachment of a large, relatively intact mass of rock along a well defined rupture surface controlled by prominent discontinuities such as bedding planes, fractures, and faults. In contrast, rockfalls involve the failure of much smaller rock masses on steep bedrock slopes. Failure as either a rockfall or rock avalanche is preceded by a period during which the cohesive properties of the rock deteriorate due to repeated saturation with water, its freezing and thawing, or due to recurrent seismic activity.

The management of rockfall and rock avalanche hazard, in analogy with management of debris flow hazard, follows three steps: hazard appraisal, application of passive measures, and application of active measures. Hazard may be appraised by determining the age and frequency of past rockfalls and rock avalanches from hazard indicators and historical records and by assessing ground conditions in an area where failure might occur. If such an appraisal shows that an area is too dangerous for permanent habitation, it may be necessary to severely restrict development there. However, it may be impractical to do this because of the difficulty of predicting exactly when, or even if, a failure will occur. Thus in most densely settled mountain regions where land is intensely utilized, the hazard from rare, single, large slope failures generally is

lacs de barrage morainique; (3) les coulées de débris résultant de l'éclatement des poches d'eau sous-glaciaire et intra-glaciaire; (4) l'éboulement de glace de glacier (avalanches de glace); et (5) l'éboulement rocheux et de la glace de glacier qui recouvre cette roche (avalanches de roches et de glace).

Des mesures préventives actives peuvent être employées pour réduire les risques que posent certains de ces types de mouvements de masse. Des lacs de barrage glaciaire et, dans certains cas, des poches d'eau sous-glaciaire et intraglaciaire peuvent être drainés en creusant un tunnel dans la roche ou la glace, ou en creusant des tranchées à travers la surface du glacier. Là où la situation se prête mal à une telle pratique, on peut construire des barrages de rétention de débris ou des bassins d'inondation en contrebas des barrages de glace, bien que, pour être efficaces, ils doivent être de grande taille et donc, coûteux. On peut aussi, en surveillant le lac et le barrage glaciaire, obtenir un avertissement de l'imminence d'un éclatement. Les mesures actives les plus efficaces contre les coulées de débris d'un lac de barrage morainique sont: (1) la préparation d'une tranchée profonde et stable dans la moraine avant la formation d'un lac important; et, (2) la construction de revêtements le long de l'exutoire pour empêcher l'encaissement et l'écroulement des berges au cours des périodes de débordement accru des lacs.

Les glissements d'éboulis rocheux et les avalanches de pierres. Les glissements d'éboulis rocheux et les avalanches de pierres sont des masses de roches brisées qui se déplacent rapidement sous forme de chute ou d'écoulement. Ils ont lieu le plus fréquemment sur des pentes raides reposant sur des formations rocheuses fracturées mais compétentes, composées principalement de roches sédimentaires à couches épaisses et de roches ignées massives, ainsi que de roches fortement métamorphisées. Au contraire, ils se produisent rarement sur des pentes recouvrant des roches sédimentaires à couches minces et des roches faiblement métamorphisées. Les avalanches de pierres sont causées par le détachement rapide d'une masse importante et relativement intacte de roche le long d'un plan de rupture bien défini que contrôle des discontinuités proéminentes comme des plans de stratification, des fractures et des failles. Au contraire, les glissements d'éboulis rocheux ne touchent que de toutes petites masses rocheuses sur des pentes rocheuses raides. Que l'éboulement se produise sous forme de glissement d'éboulis rocheux ou d'avalanche de pierres, il est précédé par une période pendant laquelle les propriétés cohésives de la roche se détériorent à cause de la saturation répétée par l'eau, son gel et son dégel, ou à cause d'une activité sismique récurrente.

La diminution des risques de glissements d'éboulis rocheux et d'avalanches de pierres de même que celle des risques de coulées de débris, suit trois étapes: évaluation du risque, mise en œuvre de mesures passives et mise en œuvre de mesures actives. Le risque peut être évalué en déterminant l'âge et la fréquence des anciens glissements d'éboulis rocheux et des avalanches de pierres à partir des indicateurs de risque et des cas antérieurs et en évaluant les conditions de sol dans une région où ces accidents peuvent se produire. Si une telle évaluation indique qu'une région est trop dangereuse pour un habitat permanent, il peut être nécessaire de restreindre strictement les activités de mise en valeur à cet endroit-là. Toutefois, une telle mesure peut s'avérer peu pratique en raison de la difficulté de prévoir exactement quand, ou même si, un accident se produira. Donc, dans les régions montagneuses les plus accepted. Nevertheless, monitoring of potentially unstable slopes can reduce this accepted risk by making timely evacuation possible. Active measures against small rockfalls are mainly preventive engineering structures placed on and along steep rock faces. Large masses of rock (>10^6 m^3), once in motion, are almost impossible to control and protective works tend to be futile.

Whether or not the risk of mass movements is perceived as acceptable or unacceptable depends to a large extent on the social and economic context in which they occur. In wealthy jurisdictions, a knowledge of past mass movements generally is taken into account during development. However, where scarcity of land, food, and health services are more pressing problems, potential slope hazards are commonly ignored; the ultimate result is considerable loss of life and economic hardship.

Although there are not hard-and-fast rules as to what is an acceptable or unacceptable risk, approaches that may be taken to reduce risk generally depend on two factors: the volume of the potential mass movement, and the probability of its occurrence. In this context, preventive structures and protective forests are most appropriate for mass movements of small size (<10^3 m^3) and high probability (<1 to 10 years average recurrence). More expensive protective and control structures are required for larger (10^3–10^6 m^3) mass movements of moderate to high probability. These can be justified economically only if the value of the property to be protected exceeds the cost of the structures, or if human lives might be in jeopardy. Land-use restrictions generally are necessary to provide protection against large (>10^6 m^3) mass movements that are likely to occur once every few hundred years or less; for rarer events, the risk generally is accepted, although it can be reduced through monitoring.

The Appendix to this report consists of 137 case histories of mass movements from the European Alps, an area with a long history of settlement and a high degree of development. These case histories show how the hazard of destructive mass movements can be managed in mountainous areas. Over the last 2000 years, people in this region have adjusted to recurrent debris flows, rock avalanches, deep-seated creep, and other slope problems through a variety of tactics, including acceptance of risk, avoidance of hazardous areas, monitoring of troublesome slopes, improved forest management, and remedial engineering. This experience obviously is a valuable guide to problems that will be encountered in western Canada as mountain valleys are developed. Except for young volcanic areas, dealt with in the general section, the mountain ranges of Austria, Switzerland, northern Italy, and eastern France display a spectrum of geological conditions broadly similar to those of the Canadian Cordillera.

The literature in German, French, and Italian

densément peuplées où la terre est intensément aménagée, le risque d'éboulement rare de grandes pentes isolées est généralement accepté. Néanmoins, la surveillance des pentes potentiellement instables peut réduire ce risque reconnu en rendant possible une évacuation opportune. Les mesures actives contre de petits glissements d'éboulis rocheux consistent principalement en des structures préventives placées sur les façades rocheuses abruptes ou le long de ces dernières. Une fois mises en mouvement, de grandes masses de roches (>10^6 m^3), s'avèrent presque impossibles à arrêter ou à dévier et les ouvrages de protection érigés se révèlent inutiles.

Que le risque des mouvements de masse soit perçu ou non comme acceptable dépend en grande partie du contexte socio-économique dans lequel il se produit. Dans les circonscriptions riches, on tient généralement compte au cours de la phase de mise en valeur des anciens mouvements de masse. Toutefois, lorsque la rareté du terrain, de la nourriture et des services de santé se révèlent les problèmes les plus pressants, les risques de rupture de pente possibles sont habituellement ignorés; le résultat ultime a entraîné des pertes considérables de vies et d'énormes tribulations économiques.

Bien qu'il n'y ait pas de règle absolue permettant de déterminer ce qui est un risque acceptable, les manières de réduire les risques dépendent généralement de deux facteurs: le volume du mouvement de masse éventuel et la probabilité qu'il se produise. Dans ce contexte, les constructions à caractère préventif et les forêts protectrices sont très appropriées pour les mouvements de masse de petite taille (<10^3 m^3) et à forte probabilité (réapparition moyenne du phénomène <1 à 10 ans). Des ouvrages de protection et de contrôle plus coûteux sont nécessaires pour les mouvements de masse plus importants (10^3 à 10^6 m^3) dont la probabilité de se manifester est de modérée à élevée. Ces ouvrages sont économiquement justifiables seulement si la valeur de la propriété protégée dépasse le coût des constructions proposées ou si des vies humaines peuvent être mises en péril. Des restrictions dans l'aménagement des terres sont habituellement nécessaires pour la protection contre les vastes mouvements de masse (>10^6 m^3) qui peuvent se produire probablement une fois tous les quelques siècles ou moins; pour des événements plus rares, le risque est généralement accepté bien qu'une surveillance suivie permette de la réduire.

L'annexe du présent rapport contient 137 cas ponctuels de mouvements de masse dans les Alpes en Europe, région ayant une longue histoire de peuplement et un haut niveau de mise en valeur. Ces cas ponctuels montrent à quel point les risques de mouvements de masse destructifs peuvent être diminués dans les régions montagneuses. Au cours des derniers 2 000 ans, les populations dans cette région se sont adaptées aux fréquentes coulées de débris, aux avalanches de roches, à la reptation en profondeur et à d'autres problèmes de pente grâce à une variété de mesures, notamment l'acceptation du danger, l'évitement des régions dangereuses, la surveillance des pentes sujettes à ces problèmes, l'amélioration de la gestion forestière et les travaux de reconstruction. Cette expérience sert de guide valable à la solution des problèmes que l'on peut rencontrer dans l'Ouest du Canada lors de la mise en valeur des vallées montagneuses. À l'exception des régions volcaniques d'âge récent, traitées dans la section générale, les chaînes de montagnes d'Autriche, de Suisse, du nord de l'Italie et de l'est de la which covers destructive mass movements in the Alps is extensive, but generally is difficult to retrieve. The Appendix summarizes much of this information in a geological, climatic, seismic, and historical framework that transcends national borders. Every locality described has been visited, and almost all descriptions are accompanied by relevant sketch maps and/or photographs.

For the purpose of slope stability, the Alps can be divided into the South Alpine and Austroalpine zones (metamorphic Paleozoic basement rocks and a cover of massive Mesozoic carbonate successions), the Pennine Zone (low- to high-grade basement and cover rocks), the Helvetic Zone (gneissic Paleozoic basement and a Mesozoic cover consisting of shale, carbonate, and flysch), and the Molasse Zone (Tertiary conglomerate, sandstone, and mudstone). Foliation, bedding, and composite fractures control potential zones of failure and detachment along valleys, which commonly are controlled by regional faults. Parts of some east-trending valleys are seismically active and presumably contain active faults.

Mountain glaciers and ice fields covered most of the Alps at various times during the Pleistocene and have left behind a great variety of coarse unconsolidated sediments. Most of these sediments were deposited during deglaciation at the close of the Pleistocene. Massive slope failures also occurrred at this time. The highest parts of the Alps still host cirque and valley glaciers and small ice fields.

The types of mass movements experienced in Alpine valleys depend to a great extent on the geology, topography, and climate of the region. Deep-seated creep is especially common on dip slopes of low grade metamorphic rocks of the Pennine Zone, but occurs also in metamorphic rocks of the South Alpine and Austroalpine basement complexes. Rock avalanches originate on dip and scarp slopes of carbonate formations in the Helvetic, South Alpine, and Austroalpine cover rocks, on dip slopes of conglomerate in the Molasse Zone, and on steep fracture-controlled faces of high grade metamorphic rocks and granites. Failure of scarp slopes in sedimentary rocks is commonly preceded by slow subsidence of incompetent strata (e.g. shaly flysch and evaporites). Debris flows, the most common and destructive mass movements in the Alps, originate mainly from Pleistocene surficial deposits, from the toe zones of sagging slopes, and from bursting glacial lakes. Ice avalanches are released from the toes of glaciers clinging to steep bedrock slopes.

The Alps are characterized by wet summers and autumns, but relatively dry winters. Precipitation is greatest on the northern and southern mountain fronts, but mid-summer cloudbursts are frequent in many intramontane valleys. Mass movements seem to be most common during summer and autumn months in areas of strong topographic-climatic gradients. However, a disproportionate number of mass movements have

France présentent une variété de conditions géologiques généralement semblables à celles de la Cordillère canadienne.

Il existe une vaste gamme d'ouvrages allemands, français et italiens traitant des mouvements de masse destructifs dans les Alpes, mais ils sont pour la plupart difficiles à trouver. L'annexe résume la plupart de ces renseignements dans un cadre géologique, climatique, sismique et historique qui dépasse les frontières nationales. Chaque localité décrite a été visitée, et presque toutes les descriptions sont accompagnées de croquis appropriés ou de photographies, ou les deux.

Selon le degré de stabilité de leurs pentes, les Alpes peuvent être divisées en une série de zones, soit la zone du sud des Alpes et la zone austroalpine (roches métamorphiques du socle datant du Paléozoïque et couverture de séquences carbonatées massives du Mésozoïque), en une zone pennine (roche du socle et de couverture de degré métamorphique faible à élevé), une zone helvétique (socle gneissique du Paléozoïque et couverture du Mésozoïque formée de schiste argileux, de carbonate et de flysch) et une zone de molasse (conglomérat, grès et pélite du Tertiaire). La schistosité, la stratification et les fractures composées contrôlent les zones sujettes aux éboulements le long des vallées que contrôlent habituellement les failles régionales. Certaines parties des vallées orientées vers l'est ont une activité sismique et contiennent probablement des failles actives.

Des glaciers de montagne et des champs de glace ont couvert la plupart des Alpes à différentes époques ou cours du Pléistocène et ont laissé derrière eux une grande variété de sédiments meubles et grossiers. La plupart de ces sédiments ont été mis en place au cours de la déglaciation vers la fin du Pléistocène. Des ruptures de pente massives se sont produites aussi à cette époque. Les parties les plus élevées des Alpes abritent encore des vallées et cirques glaciaires et de petits champs de glace.

Les types de mouvement de masse survenus dans les vallées alpines dépendent dans une grande mesure de la géologie, de la topographie et du climat de la région. La reptation en profondeur est spécialement fréquente sur les pentes inclinées des roches à faible degré métamorphique de la zone pennine, mais elle se produit aussi dans les roches métamorphiques de la zone du sud des Alpes et des complexes de socle austroalpin. Les avalanches de pierres ont leur origine sur les pentes escarpées des formations carbonatées des zones helvétique du sud des Alpes, et dans les roches de couverture austroalpines, sur les pentes fortes de conglomérats de la zone de molasse et sur les parois abruptes constituées par des fractures de roches très métamorphisées et de granites. Les éboulements sur les pentes abruptes des roches sédimentaires sont généralement précédés par une subsidence lente de la strate incompétente (p. ex., flysch schisteux et évaporites). Les coulées de débris, mouvements de masse les plus fréquents et les plus destructifs des Alpes, prennent leur source principalement dans les dépôts de surface du Pléistocène, les zones bordant les pentes affaisées et le déversement brusque des lacs glaciaires. Les avalanches de glace proviennent de l'extrémité des glaciers accrochés aux parois rocheuses particulièrement abruptes.

Des étés et des automnes humides mais des hivers relativement secs caractérisent les Alpes. Les précipitations se produisent surtout sur les fronts montagneux nord et sud, mais des trombes d'eau sont fréquentes vers le milieu de l'été dans beaucoup de vallées d'entremont. Des mouvements de masse semblent se probeen triggered by a few large regional rainstorms; these storms probably have had the greatest impact on human activity in the Alps.

Deforestation, overgrazing, mining, road construction, and other human activities have had an adverse affect on the stability of many mountainsides since about the 13th century, particularly those covered with thick surficial deposits. Reforestation, torrent control work, hazard zoning, and comprehensive land-use planning during the last century have barely kept pace with an enormous surge of development brought on by tourism and hydroelectric power generation. The Alps are now the most developed mountain region in the world, and expenditures to prevent and alleviate damage from mass movements are high. However, the case histories in the Appendix indicate that there are a variety of approaches to basically similar problems; in any specific situation much depends on the socially acceptable level of risk and the economic means of the community or regional government.

duirent fréquemment au cours des mois d'été et d'automne dans les régions ayant de fortes pentes et de grandes variations de climat. Toutefois, un nombre disproportionné de mouvements de masse ont été déclenchés par quelques grandes averses régionales; ces orages sont le phénomène ayant probablement le plus touché les activités humaines dans les Alpes.

Le déboisement, la surutilisation des pacages, l'exploitation minière, la construction de routes et les autres activités humaines ont eu un effet défavorable sur la stabilité de nombreux flancs de montagne depuis le XIII^{e} siècle environ, particulièrement ceux qui étaient couverts d'épais dépôts de surface. Le reboisement, l'endiguement des torrents, le zonage en fonction des dangers, et une planification globale de l'utilisation des terres au cours du siècle dernier ont tout juste réussi à suivre l'énorme poussée des activités de mise en valeur dues au tourisme et à la production hydroélectrique. Les Alpes sont actuellement la région montagneuse la plus aménagée du monde et les dépenses encourues pour empêcher et réduire les dommages causés par les mouvements de masse y sont élevées. Toutefois, les cas ponctuels à l'annexe indiquent qu'il y a une variété de façons de résoudre des problèmes essentiellement semblables. Toute situation particulière dépend beaucoup du niveau de risque que la collectivité juge acceptable et des moyens économiques dont dispose la communauté et le gouvernement régional.

INTRODUCTION

Mountain ranges have always posed formidable barriers to the development of transportation links and permanent settlements. Unstable valley walls, flood-prone rivers, and harsh climatic conditions present serious obstacles to a full use of mountain terrain. Yet, most mountain regions are rich in timber and mineral resources, and have a high recreational potential; part of most mountain valleys are suited for intensive agriculture as well. However, the profitable and sustained use of these resources requires a thorough understanding of the physical limitations imposed by the stability of steep slopes, the effects of snow and ice, and the strong vertical zonation of vegetaion.

Mining, logging, and trade traditionally have been the pioneer activities in mountain regions requiring construction of permanent roads in river valleys and over mountain passes. Agricultural settlements, hydroelectric reservoirs and generating stations, and tourist centres tend to follow later. As development of mountain areas proceeds, conflicts begin to arise among the various users of the land, presenting a formidable challenge to planners and managers. Eventually mountain land has to be managed so as to accommodate a wide range of activities while at the same time loss of life or property due to catastrophes such as floods, avalanches, and mass movements is minimized.

In the Canadian Cordillera many formerly remote mountain valleys are now being opened to development: mining and mineral exploration are advancing into rugged terrain; logging has encroached onto ever steeper slopes and upland areas; hydroelectric reservoirs have flooded the floors of large valleys; transmission lines, railways, and highways vie for right-of-way along narrow gorges; agricultural land is threatened by urbanization; and tourist facilities have sprung up near major population centres. These activities have brought an increasing number of people in direct contact with mountain hazards, and property losses and casualties resulting from mass movements have increased. This trend has been perceived with concern by land-use managers, politicians, and the general public.

The challenge of planning logging zones, locating townsites safely, and providing unspoiled recreation areas in western Canada will have to be met within the specific geological, climatic, hydrological, and seismic constraints of the region. These constraints are similar to those of other mountain regions, thus benefit can be gained from the land-management experience of other alpine countries. In the mountain regions of Japan and in the Alps of Europe, the transition from pioneer development to highly sophisticated multiple land-use practices has been achieved through more than 300 years of trial-and-error. This experience, paired with new insights gained in other parts of the world is at the core of this study. In this report we review the European experience with mass movements in order to show some of the problems that will be encountered and options for management as development proceeds beyond the pioneer stage in the mountains of western Canada.

We make no attempt to provide an exhaustive review of the landslide phenomenon per se, nor to catalogue all measures that can be used to stabilize slopes; the reader interested in these aspects of mass movements is referred to excellent texts by Schuster and Krizek (1978), Selby (1982), and Záruba and Mencl (1982). Instead, we describe in the first part of this report the five main types of destructive mass movements that affect mountain areas, and review the methods for reducing the hazards posed by them. Geological and meteorological controls of mass movements are discussed in each chapter. The second part of the report (the Appendix) comprises 137 case histories of destructive mass movements from the European Alps presented in a geological framework transcending national boundaries. Case histories in the Appendix are arranged in chronological order. Where more than one destructive mass movement has occurred in a specific area, which commonly is the case, the most destructive event provides the date under which the case history is catalogued. The general section and the Appendix are linked by Table 1 which is an alphabetical list of all case histories and the attributes of each that are significant for management of risk. Individual case histories can be located by using the Index of Case Histories at the end of the Appendix.

Acknowledgments

An interdisciplinary study of the scope encompassed by this report draws on many data sources and relies on the personal communications received from workers in highly diverse fields of endeavour — too numerous to be listed individually. However, we are particularly indebted to N. Skermer who supplied the initial impetus to study the Alpine experience with mass movements and its significance to the impending development of mountain valleys in western Canada. In this effort we were greatly helped by M. Akehurst, Librarian of the Geological Survey of Canada in Vancouver. Discussions with H. Nasmith, D. Lister, J.O. Wheeler, L. Jackson, W.H. Mathews, and with several classes and field trip groups from Simon Fraser University and the University of British Columbia have been most stimulating experiences for us. Our colleague Lionel Jackson also critically read the entire manuscript. One of us (G.H.E.) is grateful for the help he received at the hands of many workers in the Alps, in particular from H. Aulitzky and E. Weiss (Vienna), E. Hanausek, H. Mostler, F. Purtscheller, F. Fliri, R. Stern, and H. Schiechtl (Innsbruck), J.M. Schramm (Salzburg), V. Trommsdorff and H. Röthlisberger (Zurich), and M. Humbert (Orleans).

B. Vanlier coped gallantly with the arduous task of exotic references and names in four languages while processing the manuscript.

Table 1. Key attributes of mass movements listed in the Appendix.

Case History[1]	Type[2]				Trigger[3]				Management[4]			
	DF-Su	DF-Rk	GIMM	RkAv	Eq	Rain	Snow	HumAc	AcM	PaM	Mon	AccRi
Ahrn Valley – A92, G2	•					•	•	•	•			
Aime – A2, B3	•	•				•	•	•	•			•
Airolo – A103, D2				•		•			•			
Altdorf-Spiringen – A97, D2	•	•		•		•	•		•			
Altels – A102, C2			•				•			•		
Antelao – A74, H2				•	•	•						•
Antronapiana – A41, D2				•		•				•		•
Aosta – A27, C3	•					•		•	•	•		•
Ardenno – A24, E2	•	•										•
Bec Rouge – A91, B3		•		•		•	•	•	•			•
Becca de Luseney – A120, C3			•				•					•
Biasca – A23, E2		•		•		•						•
Bilten – A87, E1	•	•					•					•
Blisadona – A100, F1				•			•		•			
Bocca di Brenta – A94, F2				•		•						•
Bourg-St. Maurice – A39, B3		•				•	•	•	•	•		
Bozel – A105, B3	•					•						•
Brannenburg – A85, H1		•				•			•			•
Brentonico – A43, F3		•				•						•
Brienz – A22, D2	•	•				•		•	•	•		
Campo – A90, D2		•				•	•		•		•	•
Cervières – A17, B4	•					•			•	•		•
Chablais – A38, B2		•				•			•	•		•
Clavans – A15, B3				•			•		•			•
Corbeyrier-Yvorne – A32, C2	•			•	•	•	•		•	•		•
Crodo – A82, D2		•				•			•			•
Diablerets – A50, C2				•			•					•
Disentis (Muster) – A48, D2				•								•
Dobratsch – A12, I2				•	•							•
Elm – A93, E2				•		•		•				
Embach – A67, I1	•					•			•			
Felsberg – A80, E2				•		•						•
Férsina – A54, G2	•					•		•	•			
Fidaz-Flims – A116, E2				•				•				•
Fucine – A64, F2	•					•		•				•
Gaishorn – A61, J1	•	•				•			•			
Ganderberg-Passeier Wildsee – A16, G2		•				•			•			
Gastein Valley – A29, I1	•	•				•		•	•			•
Gemmersdorf – A44, J2	•					•						•
Glacier du Tour – A118, B3			•				•			•		
Goldau – A71, D1				•		•				•		•
Grächen – A58, C2				•	•					•		•
Grigno – A55, G2	•					•		•	•			
Illgraben – A126, C2		•				•	•		•			
Inzing – A135, G1	•					•			•	•		
Kalkkögel – A65, G1	•					•		•	•			
Klausen (Chiusa) – A111, G2	•					•			•			
Kollmann (Colma di Barbiano) – A99, G2	•					•			•			
La Chapelle – A18, B3		•								•		
Lago di Alleghe – A63, G2				•		•	•					•
Lake Lucerne (Vierwaldstättersee) – A69, D2				•			•					•
Lake Traun – A45, I1	•					•			•			
La Valle Agordina – A49, H2		•				•			•	•		•
Lavini di Marco (or Slavini di Marco) – A4, G3				•	•							•
Le Châtelard – A114, B3		•					•			•		
Lecco – A136, E3				•		•			•			
Leytron – A30, C2				•				•				•
Lienz – A5, H2	•	•				•			•			•
Linthal – A113, E2				•		•					•	
Martell Ice Lake – A98, F2			•				•		•			
Masière de la Vedana – A6, H2				•	•							•
Massif de Platé – A57, B3		•		•		•	•		•	•	•	•
Matrei – A26, H1		•				•	•		•			
Mattmark – A129, C2			•				•				•	•
Meiringen – A53, D2	•	•		•		•			•			•
Meran (Merano) – A3, G2		•							•			•
Millstatt – A122, I2	•					•						•
Monbiel – A62, E2				•		•	•			•		•
Mont Granier – A10, A3				•								•
Monte Masuccio – A73, F2		•				•			•			•

Case History[1]	Type[2]				Trigger[3]				Management[4]			
	DF-Su	DF-Rk	GIMM	RkAv	Eq	Rain	Snow	HumAc	AcM	PaM	Mon	AccRi
Mottec – A8, C2				•								•
Motto d'Arbino – A112, E2				•		•	•				•	
Mur Valley – A123, K1	•					•			•			•
Neukirchen – A31, H1		•				•			•	•		
Neumarkt (Egna) – A60, G2	•					•			•			
Niedernsill – A68, H1	•					•			•	•		
Nolla – A89, E2		•				•		•	•		•	
Obervellach – Möll Valley – A78, I2	•	•				•	•	•	•			
Ötz Valley Ice Floods – A47, F2			•				•		•			
Peccia – A81, D2		•				•				•		•
Perarolo – A75, H2		•				•			•			
Pettneu – A128, F1		•					•		•			
Piuro (Plurs) – A35, E2				•		•		•				•
Plaine d'Oisans – A9, B3		•				•	•		•	•		
Pra-Lagunàz – A107, G2				•			•			•		•
Putschall – A134, H1		•				•			•		•	
Rackling – A42, H1	•	•				•			•	•		
Radmer an der Hasel – A25, J1				•				•				•
Randa – A40, C2			•				•				•	
Randens – A56, B3		•				•		•				•
Reisskofel – A1, I2		•		•	•	•						•
Rentsch (Rencio) – A11, G2	•					•			•			
Riviera – A125, C5	•	•				•			•			•
Roquebillière – A110, C4		•				•				•		•
Salvan – A37, B2				•			•					•
Salzburg – A46, I1				•		•			•			
Sandling – A109, I1		•				•	•					•
San Giovanni de Crèvola – A124, D2	•	•				•			•	•		•
Schesatobel – A70, E1	•					•		•	•			
Schlanders (Silandro) – A52, F2	•	•				•		•	•			
Schierengrat – A28, D2		•				•	•	•	•			
Schwaz – A72, G1	•					•		•	•			
Semsales – A7, B2		•				•			•	•		
Serrières-en-Chautagne – A115, A3	•						•	•				•
Simplon – A104, D2			•	•			•			•		
Siror – A132, G2	•					•			•	•		
Spriana – A79, E2				•		•						•
St. Gervais-Tête Rousse – A101, B3			•				•		•			•
St. Jean-de-Maurienne – A20, B3		•				•	•	•	•	•		
Steinfeld – A13, I2	•					•		•	•			
Strigno – A133, G2	•					•			•			
Surlej – A66, E2	•					•	•					•
Tagliamento Valley – A137, I2	•	•		•	•	•		•	•			•
Tavernerio – A119, E3		•				•			•			•
Tête Noire – A36, B2		•		•			•			•		•
Toblach (Dobbiaco) – A96, H2	•					•		•	•			
Torrent de St. Barthélemy – A83, B2		•	•			•	•		•	•		
Umhausen – A59, F1	•					•			•	•		
Upper Vintschgau (Val Venosta) – A86, F2	•					•		•	•	•		
Vaiont – A127, H2				•				•			•	
Val Badia (Gader Valley) – A76, G2		•				•			•			•
Val de Bagnes – A34, C3			•				•				•	
Val Ferret-Val Veni – A51, C3			•	•			•					•
Valle di Vanoi – A77, G2	•	•				•		•	•	•		
Vandans – A108, E1		•					•		•	•		
Verrès – A84, C3	•					•						•
Vorder Glärnisch – A33, E1				•	•					•		•
Werfen – A117, I1		•				•			•	•		
Zambana – A121, G2		•				•				•	•	
Zarera – A21, F2				•						•		•
Zell am See – A130, H1	•					•			•			•
Ziano – A14, G2		•				•			•			
Ziller Valley – A106, G1	•					•	•		•	•		
Zug – A19, D1	•					•			•	•	•	

[1]Place name – Appendix number, location on index map (Fig. 40).

[2]DF-Su = debris flow (surficial deposits); DF-Rk = debris flow (bedrock); GIMM = glacier-related mass movement; RkAv = rock avalanche and rockfall.

[3]Eq = earthquake; Rain = rainstorm; Snow = snowmelt or ice melt; HumAc = human activity.

[4]AcM = active measures; PaM = passive measures; Mon = monitoring; AccRi = acceptable risk.

SLOPE FAILURE AND MASS MOVEMENTS

The slope of natural mountainsides depends largely on their internal bedrock structure and composition. Only a few rock types (e.g. igneous rocks of the granitic suite) are sufficiently competent or lack internal structure to form high cliffs and smooth bedrock slopes. More typically, internal discontinuities and pervasive structures such as bedding, cleavage, and foliation, determine the morphology of mountainsides (Fig. 1). With respect to the relationship between bedding (or foliation) and the inclination of a bedrock slope one can differentiate broadly between *dip slopes* which trend parallel to internal bedrock structures and *scarp slopes* which trend across the internal structures. In competent rock formations such as gneisses, carbonates, quartzites, and conglomerates scarp slopes are steeper than dip slopes. However, the shape of slopes in competent bedrock formations is also controlled by fractures and faults whose orientation may bear little relation to the more pervasive internal fabric of the rock mass. Incompetent rock formations such as shale, slate, and thin-bedded sandstone commonly form gently undulating slopes with little expression of internal structure. The slopes of volcanoes are special in that they result from phases of eruption of lava and pyroclastics which produce the constructional conical forms.

The highest parts of many mountain ranges are covered by glacier ice, while the lower parts are mantled by thick surficial deposits which are commonly a legacy of Pleistocene glaciations. Some of the instabilities related to processes of deglaciation have remained sources of significant mass movements (Fig. 2).

Geologically, the failure of slopes is controlled mainly by zones of weakness such as bedding planes, fractures, faults, and the contact zones between bedrock and surficial deposits. Geomorphologically, failure is promoted by undercutting of slopes by rivers and mountain torrents. A great variety of classifications have been proposed for natural slope failures and mass movements; perhaps the most widely used and detailed classification is that of Varnes (1978). We do not intend to review schemes of classification that have been proposed in the past, taking into account that the most destructive, unexpected, and thus problematic slope failures and mass movements in mountain settings are generally composites of several more elementary processes involving three major components: water, rock, and ice. Starting from these end members, mass movements in mountain settings can be grouped broadly into several categories.

Floods are a significant threat along most mountain rivers and by definition consist mainly of water and some sediment. If a flood carries significant amounts of bedload it may be referred to as *debris flood*. In a *debris flow* blocks, mud, trees, and other solid particulate matter predominate. *Debris slump* and *debris avalanche* are rotational or planar failures of generally water-saturated surficial deposits leading to subsequent rapid downhill motion of the detached material. *Bedrock slide* refers to a failure controlled by planar or curved discontinuities such as bedding and fractures. *Slope sagging* (or deepseated creep) is characterized by the large-scale and slow detachment of internally broken bedrock masses along poorly defined rupture surfaces. *Earth flow* is a slow or rapid outward movement along slopes composed of fine grained materials. *Rockfall* results from detachment and rapid tumbling of blocks from steep bedrock cliffs. *Rock avalanche* is a streaming mass of blocky debris which results from bedrock failures in competent rocks; the dry rock mass tends to travel far beyond the foot of the failed slope. *Ice avalanche* is a failure of the frontal portion of a glacier resting on a steep and smooth bedrock slope.

Composite natural mass movements are easily accommodated by a combination of these terms. To facilitate an adequate discussion of the *prevalent* types of mass movements in mountain settings we will discuss them under five separate headings: debris flows from surficial deposits, debris flows from bedrock failures, mass movements on volcanoes, glacier-related mass movements, and rockfalls and

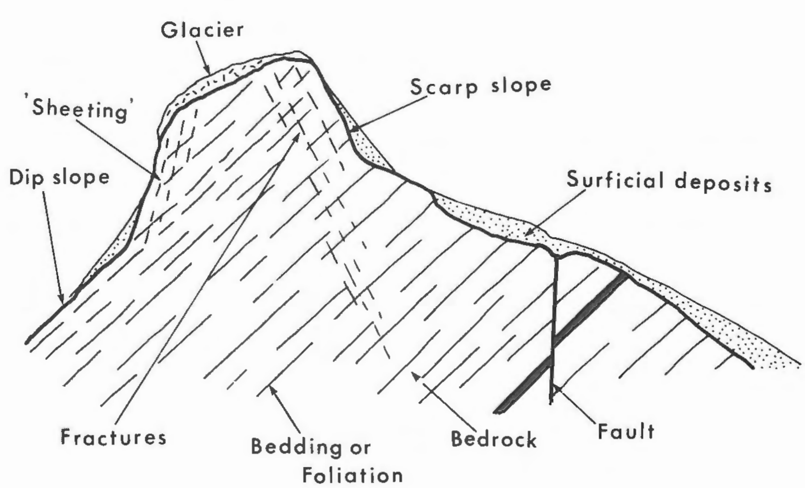


Figure 1: The topographic-geological elements of natural mountainsides.

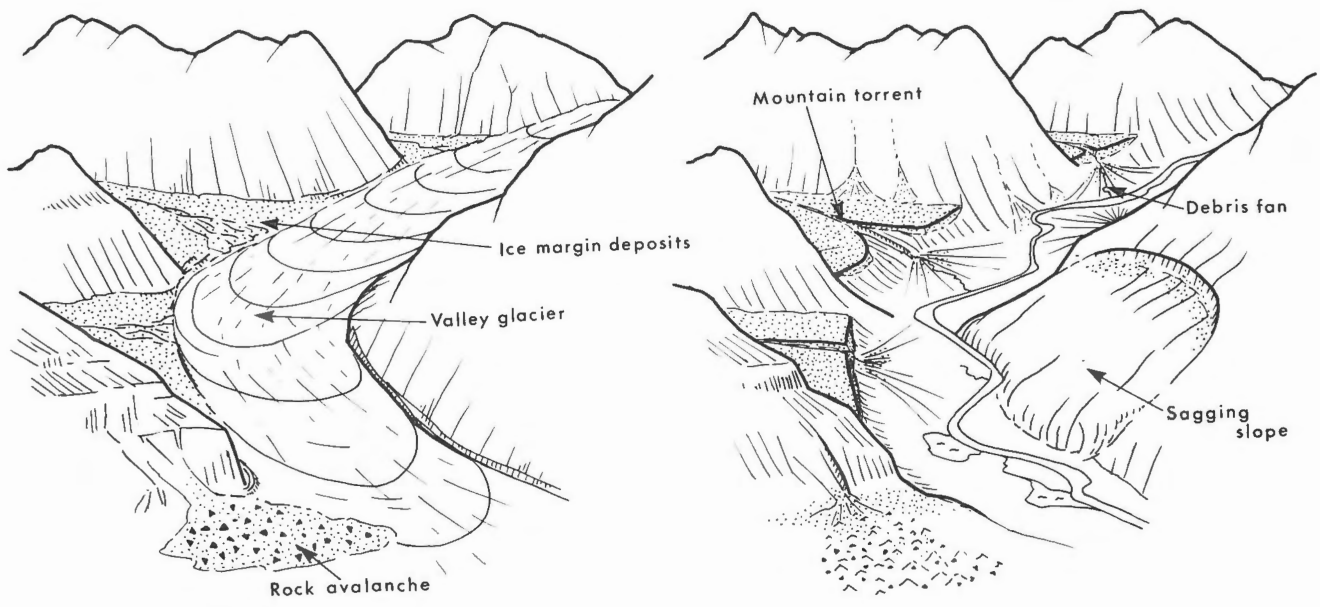


Figure 2: The role of Pleistocene glaciation in shaping valleys and influencing the distribution of surficial deposits, thus controlling areas of postglacial recurrent mass movements.

rock avalanches. These five groups of mass movements require different approaches with respect to analysis and remedial work.

Most destructive slope movements begin as incipient *slope failure* (Fig. 3a). At this stage the failing section of a mountainside passes from the stable to an unstable mechanical state. The volume and geometry of the ensuing slope failure can generally be determined roughly from the shape of the head scarp (and crown cracks), from shear displacement along the flanks, and from contraction or bulging of the toe.

In mechanical terms incipient failure can be approximated by the concept of the *stability of a rigid block* resting on an inclined plane (Fig. 3b). The weight of the block (W) exerts a normal stress σ_n perpendicular to the inclined plane, and a shear stress τ that acts parallel to the dip of the plane. The normal stress acts to hold the block in place, whereas the shear stress acts to move the block down the inclined plane. Movement of the block is restrained by the frictional and cohesional resistance along the contact zone between block and inclined plane.

The critical shear stress τ_{crit} required to overcome the frictional resistance and thus initiate motion can be expressed by the formula

$$\tau_{crit} = (\sigma_n - \sigma_{pw}) \tan \theta + C$$

where σ_{pw} is the pore water pressure along the contact zone; θ is known as the 'angle of friction'; $\tan \theta$ is the coefficient of friction, a material constant which for most natural materials ranges from 0.6 to 0.7; and C is the cohesion across the contact zone. In simple terms this equation says that the downslope shear stress component necessary to overcome

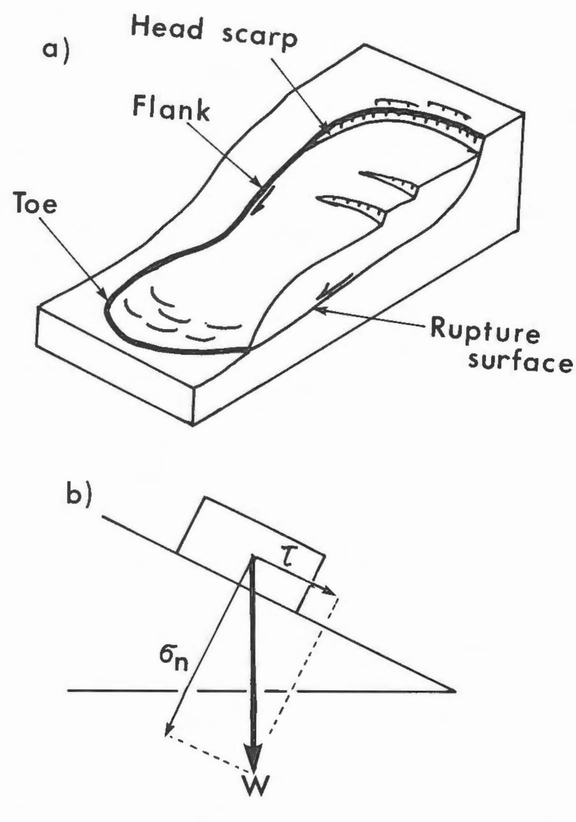


Figure 3: a) Elements of a typical slope failure in its incipient stage. b) The mechanical analogue of a failing slope: the rigid block resting on an inclined plane.

the resistance to sliding decreases with decreasing normal stress (e.g. on a steep potential rupture surface), decreases with increasing pore water pressure (e.g. rising groundwater table, infiltration of rain or meltwater), and decreases with decreasing cohesion (e.g. seismic shaking or blasting of rock, removal of tree roots in surficial deposits).

Although the stability of a slope is mainly controlled by the orientation of internal planar structures, pore water pressure, and the cohesion of the materials forming the slope, it is the topography *below and downslope* from the incipient failure that will determine whether or not an incipient mass movement will accelerate and thus pose a hazard to human works or lives. In this respect the most significant aspects contributing to complex mass movements are embraced by the concept of the torrent system.

MOUNTAIN TORRENTS

High mountain ranges are dominated by narrow flood-prone river valleys, steep tributary basins, and ever more barren bedrock ridges and peaks rising above. The principal valleys are commonly underlain by fault zones in bedrock which, over geological time, have been selectively eroded by running water and glaciers. Fracture zones and faults also control the orientation of many tributary valleys. However, the detailed shape of bedrock massifs and the distribution of surficial deposits can be attributed to the glacial history at the end of the Pleistocene Epoch, processes of weathering, and mass movements.

Although many mass movements are related directly to failure of mountainsides, most arise from the complex interaction of the unstable slopes with the branching network of channels, gorges, ravines, gullies, and chutes that define the upland basins of mountain torrents. The term *torrent*, as used here, denotes a high-gradient ephemeral or perennial water course in montainous terrain which is characterized by sporadic and sudden discharge of debris. In this potential for sudden debris movements a torrent differs from creeks which discharge from basins in more subdued terrain. Mountain torrents are notorious for the fact that discharge of water gives little or no hint as to the quantities of debris released sporadically.

The *torrent system* (Fig. 4), if viewed within the context of a long period of time (e.g. hundreds or thousands of years), undergoes cycles with deceivingly subdued discharge of water ('dormant torrents'), and short pulses of vigorous debris transport ('active torrents'). Several descriptive geomorphological-geological elements can be identified for both the dormant and active torrent stage (Aulitzky, 1980; Eisbacher, 1982). Every torrent has a *catchment area* (or drainage basin) which is located above the floor of the adjacent river valley and is bordered by a crest, composed of bedrock ridges and peaks. Within the catchment area of the torrent lies the *debris source area* which contains potentially mobile debris within the reach of tributary torrent branches. By definition the debris source area of a torrent system is smaller than the catchment area. When the torrent system changes from its dormant to its active stage debris from the source area enters the torrent via slump scars, chutes,

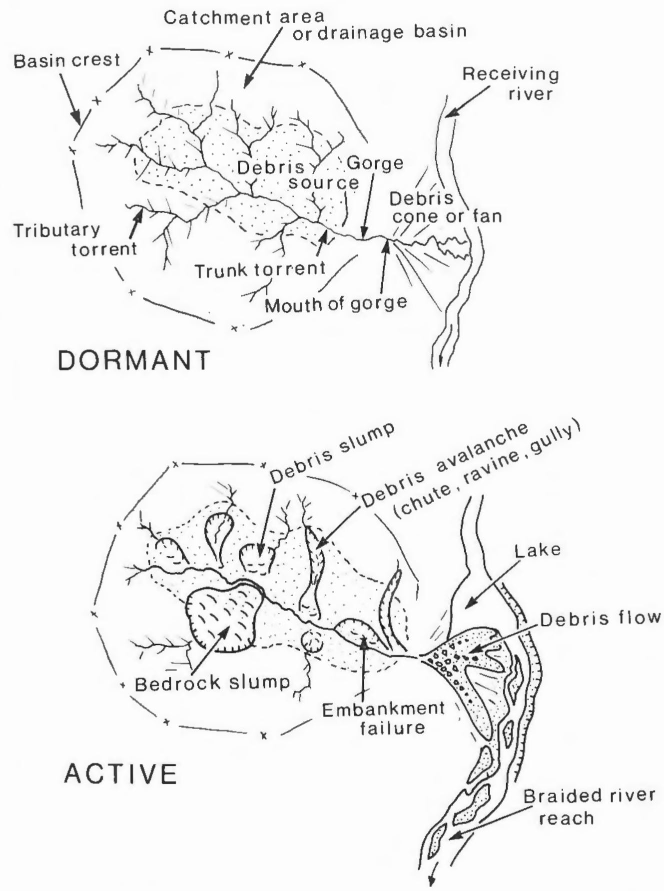


Figure 4: Descriptive elements of dormant and active torrent systems along major mountain valleys; see text for details..

ravines, gullies, embankments, bedrock slides, sagging slopes, morainal dams and other unstable features. Debris from tributary torrent branches collects in the trunk torrent and from there descends into the *gorge*. In the gorge temporary blockages of debris along narrow bedrock reaches create the coherent and pulsating flows characteristic of the active torrent system. From the mouth of the gorge the debris emerges onto the surface of a *fan (or cone)*. Fans are the result of repeated deposition of debris flows and the reworking of debris. In general the term debris fan or alluvial fan is used to denote flat and internally stratified debris accumulations; the term debris cone is generally reserved for steeper (e.g. more than 4°) accumulations of debris. The sporadic additions of debris masses to the surface and along the outer fringe of the fan also modify the geometry and flood potential of the *receiving river* on the valley floor: on the upstream side accumulations of debris create lakes, swamps, or low grade floodplains; at the toe of the fan the river channel is pushed against the far embankment and thus causes erosion; on the downstream side the coarse bedload added by debris flows creates an unruly braided reach. Some debris fans prograde into lakes or fiords, forming *fan deltas*. Fan deltas consist characteristically of a shoreward-tapering wedge of fine

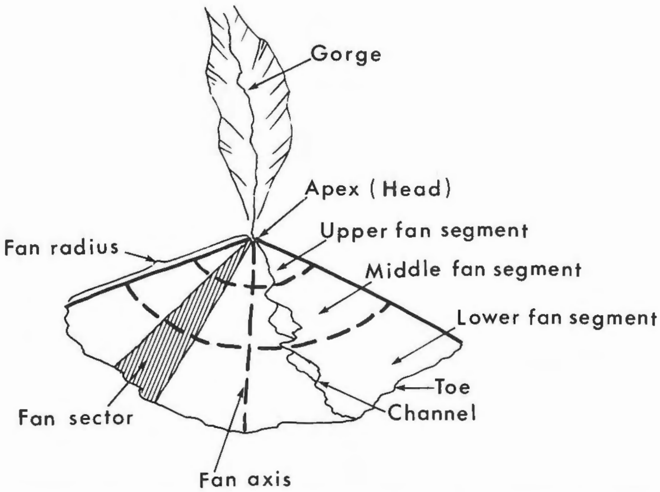


Figure 5: Descriptive elements of a debris fan (or cone).

grained sand and silt overlain by a blanket of alluvial gravels and debris flow deposits. Discharge of massive debris flows onto the fringe of the fan delta or other forms of overloading may result in failure of the delta front and possibly lead to retrogression of the failure onto other parts of the fan.

Debris fans (or cones) are the most conspicuous manifestations of active or dormant torrent systems and generally the first to be developed. Several elements can be used to describe their geometry in detail (Fig. 5). The mouth of the gorge coincides broadly with the *apex* (head) of the fan, which is also its highest point; below the apex upper, middle, and lower *fan segments* can be differentiated; the *toe* of the fan is defined as the lowest line of the fan and coincides with the embankment of the receiving river; the *radius* of the fan is the horizontal distance between apex and receiving river; the *fan axis* divides the area of the fan into two halves of about equal size; *fan sectors* are pie-shaped portions of the fan tapering towards the apex.

The torrent system thus consists of three parts, one which generates, another which translates, and a third which receives debris. The torrent system is intimately linked not only to the generation of debris flows, but also commonly hosts starting zones, tracks, or runout zones of snow avalanches. The understanding of this natural unit is fundamental for mountain hazard appraisal.

DEBRIS FLOWS FROM SURFICIAL DEPOSITS

The term *surficial deposits*, as used here, includes unconsolidated or semiconsolidated materials that overlie bedrock. Surficial deposits comprise Quaternary and older sediments laid down mainly by rivers and glaciers. In young volcanic mountains most of the stratified deposits are transitional in character between bedrock (lava flows) and surficial materials (ash, pyroclastics); mass movements in volcanic terrain are therefore discussed in a separate chapter.

The properties of surficial deposits depend greatly on their origin: some surficial materials are *relicts* of depositional processes which prevailed during Pleistocene glaciations; others are in equilibrium with the present environment and have formed by *in-situ weathering, mass wasting*, and other processes. Relict surficial deposits are generally somewhat better indurated, show internal structures such as bedding and have a well defined base above unweathered bedrock; in-situ soils and colluvium are more homogeneous and tend to grade downward into disintegrating bedrock.

Mountain valleys that were glaciated during the Pleistocene are U-shaped in cross-section, and slopes are mantled by discontinuous and often thin layers of surficial deposits such as talus, till, unsorted ice-contact debris, landslide rubble, alluvial gravel, and lacustrine silt. These deposits are commonly slightly indurated and their irregular external morphology and internal structures are, as a rule, the legacy of processes that operated during or shortly after deglaciation, some 10 000 years ago. They thus reflect a climate and vegetation cover quite different from the ones prevailing today. Only in the immediate vicinity of present-day glaciers do such vigorous depositional and erosional processes still prevail. In most other areas dense vegetation or a dry climate have helped to preserve, as relicts, the morphology of Pleistocene deposits. Only radical changes in the hydrological regime of the slopes, (e.g. during rare storms, after deforestation, or by careless road construction) reactivate the long-dormant ravines, gullies, and incipient slumps in these deposits. Thin soils, which developed in the last 10 000 years on relict surficial deposits, are particularly prone to fail along pre-existing V-shaped ravines (Stini, 1910; Swanston, 1969; Moser, 1973). Once established, such persistent 'scars' may grow into major debris sources and lead to destructive mass movements (Stini, 1931; Schauer, 1975; Bunza et al., 1976).

Mountain slopes in low- and mid-latitude mountain regions that were not covered by Pleistocene glaciers in general fringe roughly V-shaped valleys and are covered by deep soils, colluvial blankets, and disintegrating bedrock. These surficial deposits are distributed quite uniformly, but are thickest near the confluence of tributary branches of upland torrents. A stable vegetation cover plays a major role in maintaining the subtle balance between in-situ generation of debris by weathering and its downslope transport by running water, creep, or slumping. The resulting geomorphology of upward branching networks of gullies and bowl-shaped amphitheatres represents the adjustment of the landscape to recurrent intervals of erosion by running water and slope failure ('gullying' and 'slumping'). Typically, during storm events these two processes act hand in hand (Temple and Rapp, 1972; Brunsden et al., 1981; Caine and Mool, 1982).

Storm-generated debris flows in both relict and in-situ surficial deposits occur more frequently in areas where forest cover has been removed, agricultural terracing abandoned, road construction expanded, and stabilization of stream embankments neglected. What actually constitutes an 'extreme' or 'rare' meteorological trigger event therefore defies precise definition (Starkel, 1976).

Field studies in a great variety of mountain settings have shown that the critical stages in the mobilization of debris and the changeover from the dormant to the active stage of torrents are basically related to intensity and cumulative total of precipitation, exceeding greatly the long-term seasonal averages (Strele, 1932; Starkel, 1972; Campbell, 1975; Fuxjäger, 1975; Rapp and Stromquist, 1976; Bell, 1976; Guidicini and Iwasa, 1977; Ikeya, 1976; Moser, 1980; Crozier et al., 1980; Caine, 1980; Eisbacher and Clague, 1981; Eschner and Patric, 1982; and others). These studies indicate that there are three main types of debris-generating meteorological events: high-intensity downpours, sustained regional rainstorms, and rainstorms associated with rapid snowmelt.

High-intensity downpours (cloudbursts), with more than 40 to 60 mm of rain (and/or hail) falling within one hour, trigger debris flows in high-gradient ravines, torrent channels, and along unprotected escarpments of unconsolidated deposits. Much of the debris is eroded by running water in V-shaped tributary scour ravines and gullies. Bursting logjams accentuate the power of pulsating flows along the trunk channels.

Sustained regional rainstorms with more than 150 to 200 mm of rain in 24 hours and intermittent squalls with 10 to 30 mm of rain per hour cause soil slips, debris avalanches, and embankment slumps, all of which may transform into channelized flows.

Rainstorms associated with rapid snowmelt, with ambient air temperatures rising above 5–10°C, release mixed snow–debris avalanches, cause sudden runoff from frozen terrain, and reactivate large slumps by infiltration of meltwater into crown cracks of incipient bedrock failures and debris slumps.

As indicated in Figure 6, the type of meteorological trigger event thus broadly determines the debris-mobilizing mechanism in the torrent system: short intense storms primarily cause erosion by running water; sustained rainstorms create soil slips, debris avalanches and erosion by running water; long-lasting rains and heavy rains onto snowpacks tend to reactivate pre-existing large slumps and accelerate slow-moving earthflows. Beyond the external meteorological controls debris mobilization in the source areas of torrent systems depends greatly on the texture, vegetation cover, topography, and human activities that modify the slope environment. Most rainstorm-related slope failures in surficial deposits are thus complex and although it may be possible to provide sophisticated after-the-fact explanations for almost any type of mass movement, prediction of the location and timing of storm-related soil slips, debris avalanches, slumps, or flows is still in its infant stages (Moser, 1980). This is partly due to the fact that the mobilization of debris in the source areas of mountain catchments commonly starts with local particle erosion by running water or discrete slope failures where such have not occurred in recent history.

The erodibility of unprotected surficial deposits by running water depends on their texture, state of consolidation, and cementation. In the order of broadly increasing erodibility, surficial deposits can be grouped into till, cemented gravel, clay, uncemented gravel, sandy gravel, sand, and

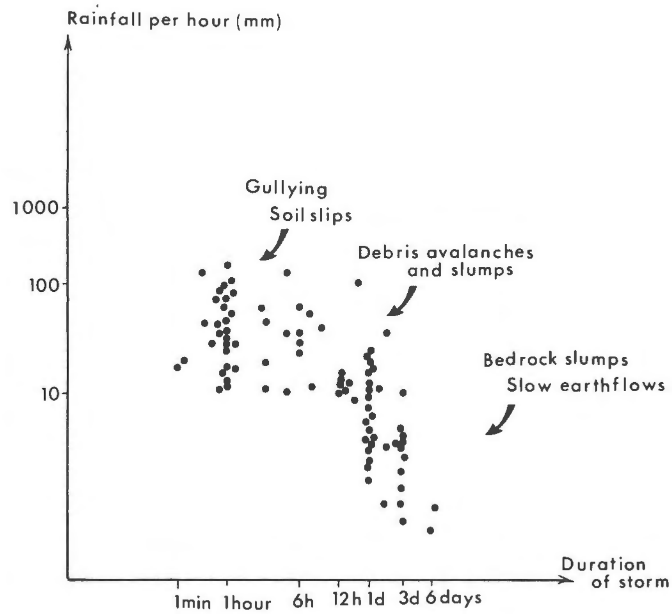


Figure 6: Diagram depicting critical rainfall per hour (in mm) against duration of rainstorms during events which triggered massive release of debris. Dots represent real events reported in the literature (see text for references).

silt. In general, unsorted sandy and bouldery gravels, exposed along high torrent embankments are the most common erosional point sources for potentially destructive debris flows (Stini, 1931). Channel reaches bordered by such bluffs commonly are choked with coarse bedload from previous floods and mass flows (Fig. 7). However, the precise point at which a storm-fed mountain torrent begins to carry bouldery debris is not easily defined. Bradley and Mears (1980), from a review of available theoretical calculations and field studies, suggest that at mean current velocities of 4 to 8 m/s blocks of up to 2 m in diameter begin to move. The potential of a confined torrent to carry coarse bedload also depends greatly on its gradient: it takes very little running water to undercut a steep bouldery embankment along a high-gradient channel and initiate a skidding and rolling motion of the bedload. The most prominent point sources along a channel from which debris can be mobilized during storm runoff can be identified in a general way (Eisbacher, 1980).

Even more difficult than the identification of scars with a potential for debris generation by erosion is the prediction of where discrete debris slumps might occur during storms. Susceptibility to failure depends greatly on slope geometry, type of material, and vegetation cover. Slopes of surficial deposits with inclinations between 20 and 40° are most likely to fail during storms (Moser, 1980). Numerous field observations during and after catastrophic rainstorms indicate that shallow debris slides occur explosively, and that failure of the toe zone is accompanied by the release of significant amounts of water (Jones, 1973; Fuxjäger, 1975; Rogers and Selby, 1980; Pierson, 1983). This suggests that during extreme storms water from surface runoff and throughflow enters unconsolidated materials not only by way of the intergranular pore space but also through open soil 'pipes' which are oriented in a downslope direction along the interlocking network of decaying tree roots. Intergranular and pipe water together create the porewater pressures that reduce the effective internal shear resistance and thus cause failure (e.g. Swanston, 1970; Moser, 1980; Sidle and Swanston, 1982). The coincidence in the timing of debris avalanches with intense squalls during sustained rainstorms (e.g. Ikeya, 1976; Eisbacher and Clague, 1981), indicates that failure is caused mainly by the rapid transfer of water from the surface into subsurface zones 1 to 2 m deep, rather than by the gradual rise of the groundwater table.

Failure of surficial debris commonly occurs along linear downslope re-entrants where surface and subsurface water can be expected to converge during storms (Johnson and Rahn, 1970; Williams and Guy, 1973; Anderson and Burt, 1978; Pomeroy, 1980). Indeed, scars of shallow failures tend to accentuate re-entrant slope contours above springs and near the confluence of subsidiary draws; detachment surfaces have roughly the shape of an inverted spoon (Fig. 8). Stini (1931, p. 58-99), who documented these features in detail termed them 'Muschelanbruch' (=conchoidal failure). He showed that near topographic re-entrants debris generation starts at a single point and then swiftly expands up the slope perpendicular to the contours. In depressions filled by thick surficial deposits this mechanism creates erosional scars, which later are scoured by surface runoff and are thus enlarged until the local debris source is exhausted (or stabilized by control methods).

On relatively planar slopes failure of surficial deposits begins with the opening of downward-concave zones of en-echelon crown cracks that coalesce into a head scarp (Heim, 1932, p. 18-32). The failure commonly expands laterally and assumes the shape of an inverted V, similar in geometry to the starting zone of snow slab avalanches (Fig. 8).

Figure 7: Channel of torrent flanked by bluffs of relict surficial deposits composed of bouldery gravels which serve as source for occasional debris flows in the Muncho Lake region of British Columbia. (GSC 204165)

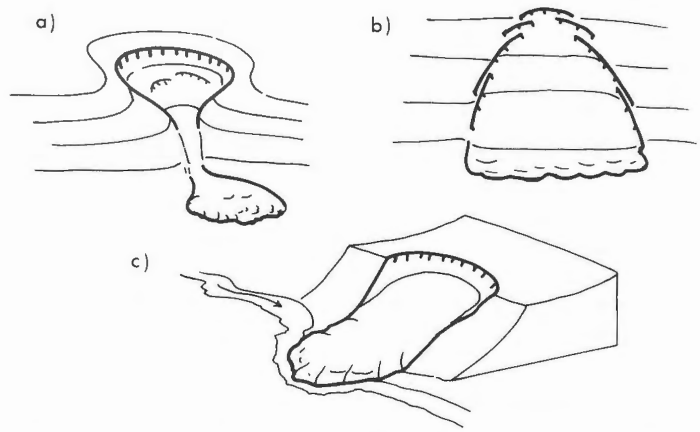


Figure 8: The three principal types of shallow slope failures in surficial deposits: a) spoon-shaped retrogressive failure in topographic re-entrants; b) planar failures; c) embankment slumps.

Along terraces and channel embankments, a rising groundwater table (and increased piezometric pressure) may induce major slumps whose depth and shape depend to a large extent on the thickness and homogeneity of the surficial materials. Oversteepening of channel embankments by erosion also may trigger large debris slumps in the source area of torrents (Fig. 8).

Once failure has occurred the inclination of the slope, ravine, or channel determines whether or not the detached masses of debris combine and accelerate into devastating debris avalanches or flows. Ravines inclined more than 25° are commonly dangerous conduits of failed debris and dislodged boulders; logs, once in rapid motion, become powerful tools of destruction (Fuxjäger, 1975; Eisbacher and Clague, 1981; Smith and Hart, 1982).

Erosion by running water, retrogressive slope failure, debris avalanches, and embankment slumps all act in concert until entire sections of the saturated debris mass begin to move along the torrent channel and a debris flow is born. Debris flows accelerate with increasing water content, volume, and channel gradient; they decelerate on gentle and wide reaches. Driven by gravity, many debris flows travel tens of kilometres, although their advance is highly unsteady and pulsating. In bedrock gorges below the source area individual debris pulses may be stemmed by channel obstructions (log jams, rock spurs), only to grow in volume, and finally break through with explosive violence. The commonly observed change in the shape of channel cross-sections after the passage of flows from erosional V-shaped ravines in the source area to distinctly U-shaped paths farther down indicates that debris flows pick up loose blocks etc. by ploughing and plucking, particularly at the head of the flow (Stini, 1910; Johnson, 1970).

Early visitors to the high ranges of the Alps noted the strange mixture of bouldery debris and water issuing from the mouths of torrent gorges and described them as '... migrating heaps of stones...' (Walcher, 1773) or '... a kind of liquid mud mixed with decomposed slate and rock fragments; the impulsive force of this dense paste is incomprehensible; it incorporates rocks, topples the buildings which happen to be in its way, unroots the tallest trees, and, upon bursting forth from deep ravines, ravages the fields and covers the soil with a considerable thickness of silt, gravel, and rock fragments ...' (de Saussure, 1781, paragraph 485). The difference between debris transport by floods and debris flows was succinctly summarized by Stini (1910, p. 36-37): '... in moving water the largest stones drag behind while smaller debris moves ahead because of its smaller frictional resistance; in a debris flow the large blocks are generally carried in the frontal tongue where they bounce ahead until they are overtaken again by the debris, thus regaining momentum; naturally, debris flows do not develop a crust comparable to lava flows, but up to a degree, one can differentiate a water-rich core from a water-poor crust; the latter gives the debris flow its characteristic outer shape according to whether it consists of stones, trees etc. ...'. The frightening blocky frontal wall, noted by many observers of debris flows, accounts for most of the damage and destruction.

Recent detailed observations, experimental studies, and theoretical work have established that the frontal 'bore' of a moving debris flow is roughly drop-shaped and that it consists of more solids than water. The flow characteristics of debris flows thus vary in both cross-section and longitudinal profile (Fig. 9). In cross-sections perpendicular to the direction of flow the central 'plug' is fringed by a marginal zone with laminar flow; a 'dead zone' characterized by sharp-crested spillover ridges and levees marks the lateral borders of the debris flow along pre-existing torrent channels. In longitudinal section flow characteristics vary from non-Newtonian near the front to partly turbulent Newtonian in the tail. Along steep reaches blocks tumble or avalanche ahead of the main body of the flow, but are soon overtaken and reincorporated into the main debris mass. Emergence of larger blocks at the front, top and margins of the flow is due partly to dispersive forces between grains, partly to buoyancy of large blocks in a finer grained matrix, and partly to the forward push exerted by debris pulses blocked behind large particles. Pushing, ploughing, and plucking at the debris front account for the often dramatic volume increase of flows along the lower parts of their tracks. Peak velocities of debris flows along high-gradient and straight torrent gorges may reach more than 50 m/s; through momentum transfer, individual blocks may be propelled forward as projectiles with velocities greatly in excess of those of the main flow (Bagnold, 1968; Johnson, 1970; Aulitzky, 1970; Ikeya, 1976; Mears, 1977; Hampton, 1979; Okuda et al., 1980; Jian and Defu, 1980; Takahashi, 1981; Xizhi, 1981; Iwamotu and Hirano, 1982).

Debris flows spread and thin rapidly once they flare across the debris cone at the mouth of a gorge or below a debris chute. Unconfined flows always attempt to maintain a straight course: minor depressions and obstacles are quickly buried or overtopped and do not greatly modify the debris trajectories. However, on the fan or cone frictional forces

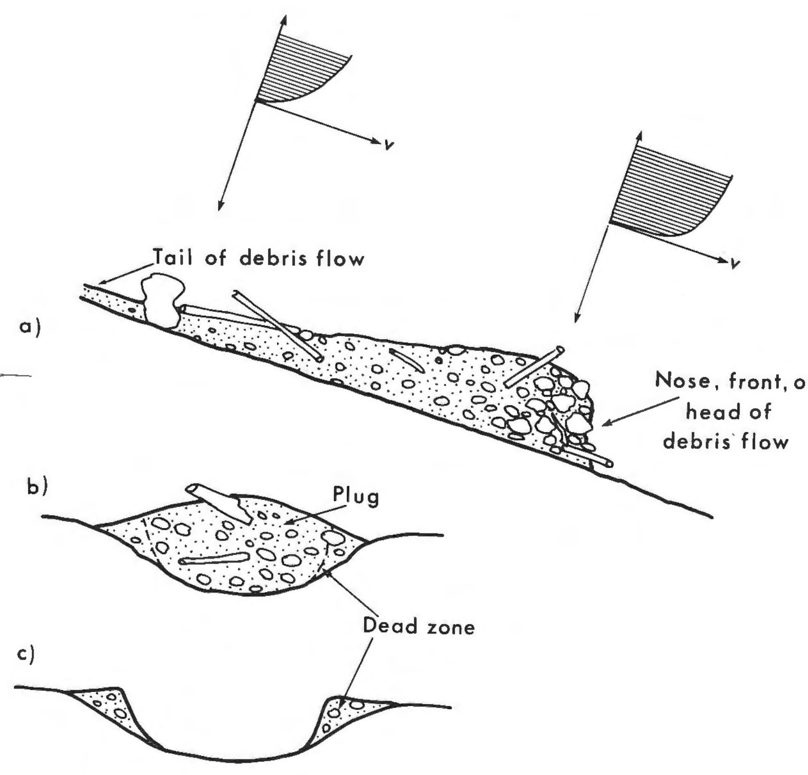


Figure 9: Characteristics of a moving debris flow:
a) longitudinal section showing variable concentration of solids and vertical velocity gradients in different parts of the flow;
b) cross-section through debris flow in motion, indicating central plug and marginal dead zones;
c) cross-section of torrent channel after passage of debris flow; note debris levees left behind by the flow.

within the flow soon dominate, and the flow breaks into finger-like lobes, the coarsest fraction (trees, boulders) settling out rapidly. The lobate geometry of debris flow deposits accounts for the complex ridge-and-mound pattern and the great variety of bedding structures exposed on active fans or cones. Structures tend to be even more complex where fresh debris flow deposits are reworked by normal fluvial processes (*see* Nilsen, 1982, for a review of sedimentary structures in composite alluvial fans).

Because of the variety of geological and topographic parameters influencing the mobilization and transport of debris, it is impossible to identify a single mechanism of debris flows. Debris flows have specific source area mechanics (erosion by running water, debris avalanching,slumping, etc.), transport mechanics (wide channels, narrow gorges, etc.), and depositional mechanics (single lobe, multiple laterally overlapping lobes, etc.). Some flows are merely the extreme stage of fluvial bedload transport and can be referred to as 'debris floods'; others carry predominantly uprooted trees, silt, and clay, and are true 'mudflows'; others consist entirely of boulders or slabs of bedrock and rank as 'boulder flows'.

In the Alps many debris flow disasters have resulted from deep scour by running water and from subsidiary embankment collapse of surficial deposits. Examples in the Appendix include Cervières, Schesatobel, Toblach, Rentsch, Lake Traun, Umhausen, Neumarkt, Kalkkögel, Lienz (Aguntum), Torrent de St. Barthelémy, Millstatt, Riviera, Kollmann, Klausen, Siror, Inzing, Fucine and many other debris flow disasters. Shallow debris avalanches and slumps have created massive flows at Aime, Steinfeld, Obervellach-Möll Valley, Upper Vintschgau, Gastein Valley, Schlanders, Surlej, Niederndsill, Schwaz, Valle di Vanoi, Corbeyrier-Yvorne, Bilten, Bec Rouge, Mur Valley, Zell am See, Biasca etc. Temporary blockages of torrent gorges by debris have contributed significantly to the recurrent catastrophes of Férsina, Grigno, Verrès, Klausen and many others.

DEBRIS FLOWS FROM BEDROCK FAILURES

Every major failure of bedrock embankments along a torrent has the potential for creating a serious downstream debris flow hazard. If the gradient of the basin is sufficiently steep, a bedrock failure may change directly into a channelized debris flow and the distinction between slope failure and debris flow becomes difficult. More often, however, bedrock failures simply block the torrent channel; during subsequent periods of intense precipitation or overflow, debris is mobilized from the slide mass. This type of debris flow and overflow-flood problem is not restricted to any particular type of bedrock failure, although deep-seated creep ('sagging') that involves entire mountainsides is especially troublesome. Bedrock failures with volumes of up to $1 \times 10^6 m^3$ are generally amenable to remedial measures, (*see* Schuster and Krizek, 1978; Záruba and Mencl, 1982), and the control of an incipient bedrock failure may prevent major interference of its toe zone with adjacent stream channels. However, many debris flow problems originate at the toe zone of bedrock slopes which involve entire mountainsides; particularly slopes underlain by incompetent rocks such as mica schists, phyllites, slates, argillaceous limestones, and sandy flysch formations experience deep-seated gravitational creep. Large scale detachment of bedrock slopes by deeply penetrating creep is generally restricted to mountain regions with neotectonic uplift and/or relatively recent deglaciation. Both tectonic uplift and deglaciation cause lowering of local erosional base levels and thus induce substantial gravitational forces in the valley walls. Deep-seated creep or sagging occurs in all young mountain ranges of the world (e.g. Cordillera, Carpathians, Apennines, Alps, Himalayas).

Deep-seated creep does not necessarily lead to accelerated motion or catastrophic failure, although parts of such slopes may fail suddenly (e.g. rockfalls or toe zone slumps). Deep-seated creep may involve as much as $2000 \times 10^6 m^3$ of bedrock and reach depths of several hundred metres. Sagging attains its most spectacular expression and was first recognized on foliation dip slopes of low grade metamorphic rocks. Buxtorf and Wilhelm (1920) were among the first to clearly separate deposits of completely failed bedrock slopes ('Bergsturz' = rockslides or rock avalanches) from semi-detached rock masses undergoing deep-seated creep ('Sackungsvorgänge' = sagging processes).

Deep-seated creep manifests itself by diffuse zones of crown cracks and partly inactive head scarps (commonly referred to as 'pseudo-cirques', 'twin ridges' etc.). In zones of known seismicity and in competent rocks such zones of crown cracks, particularly uphill-facing (antislope) scarplets, may be difficult to distinguish from seismogenic fault scarps (*see* Radbruch-Hall, 1978; Clague, 1979; Bovis, 1982).

Below the broad zone of crown cracks there commonly are gently inclined benches, which, in many mountain regions, have long served as settlement sites (Huder, 1976). Settlements located there move slowly (e.g. a few centimetres per year) often with little relative displacement between buildings. Toe zones of sagging slopes impinge against the opposite valley wall or extend underneath the valley floor, partly buttressed by the toe of the opposite slope (Fig. 10).

During the first half of this century geologists working on large-scale construction projects in mountainous terrain recognized that in contrast to smaller bedrock slumps basal surfaces of sagging slopes are difficult to define. Planar or spoon-shaped zones of rupture merge almost imperceptibly with the semicoherent creeping rock masses above and with the solid bedrock below. The interpretation of stepped head scarps as belonging either to several discrete shallow slides or to a single sagging rock mass is of considerable practical significance and was first seriously discussed by Ampferer (1939) and Stini (1941, 1942).

In central Europe slopes of this type have been termed 'Sackung' (= sagging slope), 'Talzuschub' (= valley closure), 'Bergzerreissung' (= mountain splitting), 'ancien tassement profond' (= ancient deep sagging) and have been

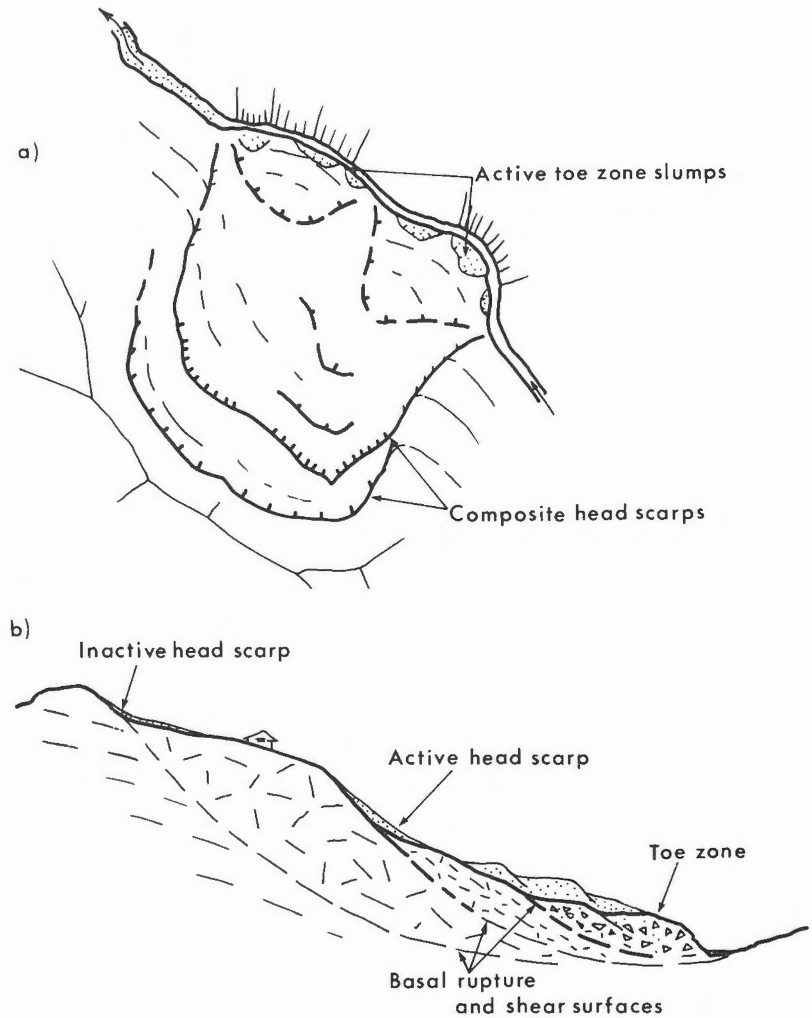


Figure 10: Map view (a) and cross-section (b) of a mountainside undergoing deep-seated creep. Note composite head scarps of inactive and active elements, and toe zone slumps which interfere with the torrent channel along the narrow valley bottom.

shown as such on large-scale geological maps (Heim, 1932; Zeller, 1964; Clar and Weiss, 1965; Nemčok, 1972; Burri and Gruner, 1976; Horninger and Weiss, 1980). Nevertheless, only minor progress in the understanding of the mechanics of large sagging terrains has been achieved until recent years (*see* Terzaghi, 1950). The understanding of sagging terrains came about through detailed mapping of the diffuse headscarp regions, definition of the basal shear and rupture zones by drilling and seismic investigation, geodetic measurements of surface displacement, evaluation of secular and annual hydrological regimes, and an application of theoretical models (Jäckli, 1957; Haefeli, 1967; Zischinsky, 1969; Brückl and Scheidegger, 1972; Robinson and Lee, 1972; Radbruch-Hall et al., 1976; Huder, 1976; Mahr, 1977; Imrie and Patton, 1977; Piteau et al., 1978; Hauswirth et al., 1979, 1982; Kronfellner-Kraus, 1980; Moser and Glumac, 1983).

Creep rates on many sagging slopes are so low (millimetres or centimetres per year) that displacements pose little or no direct threat to human structures placed on them. However, on others creep rates respond dramatically to induced or natural changes in the hydrological regime of the slope (Neuhauser and Schober, 1970). This holds particularly for small (50 to 100 $\times$ $10^6 m^3$) and steep (30 to 40°) sagging slopes with planar detachment zones. There, removal of the forest cover, submergence of the toe zone by a reservoir, enhanced infiltration of snow melt, or an increase in total annual precipitation may cause acceleration and discrete slippage in certain zones of the slope. Creep rates are generally higher at the 'active' toe zone than near the head scarp (Huder, 1976; Imrie and Patton, 1977; Moser and Glumac, 1983). Fresh zones of crown cracks delimit 'incipient' failures above the toe.

Accelerating creep rates may or may not necessitate abandonment of human works on the sagging slope itself. However, they invariably pose problems to structures along the toe zone and particularly to the stability of adjacent torrent channels. If the broken rock mass constrains the flow of a torrent the reach along the toe zone becomes a growing source of debris. The volumes of resulting flows depend largely on the channel gradient below the constriction and on the rate of slope deformation which replenishes the debris source along the channel. Accordingly there are two prototypes of torrent basins with bedrock failures as source areas: small high-gradient basins and large low-gradient basins.

High-gradient torrents (30 to 60%) which drain small funnel-shaped catchments (1 to 10 km^2) underlain by unstable bedrock slopes have the potential for large blocky debris flows (0.1 to 1 $\times$ $10^6 m^3$). These flows are transitional in character between bedrock slides and high-mobility water-saturated flows. In detail, the debris enters the torrent system either as individual rockfalls or as unstable toes of large sagging slopes and bedrock slumps. Debris cones at the mouths of these torrents tend to be steep, their volume seemingly out of proportion with the small area of the catchment basins. The bedload of the dormant torrent stage gives no hint as to the debris potential of the active torrent stage which is initiated entirely by failure of upland slopes. Mapping and monitoring of individual slope failures thus become a significant input for the evaluation of the debris flow potential in high-gradient and unstable torrent basins (Zeller, 1972a; Japanese Society of Landslides, 1972; Malatrait, 1975; Cotecchia, 1978; Azimi et al., 1980; Smith and Hart, 1982).

Along low-gradient torrents (up to about 30%) and rivers, sagging slopes continuously replenish the bedload to the point of changing the torrent channel into a braided track of bouldery debris. If the torrent (or river) has a high mean discharge, the debris load that is being added slowly from the toe zone of a sagging slope will be carried away during annual periods of enhanced runoff. However, if normal runoff is insufficient to carry away the debris the blockage of the channel along the toe zone may create a lake; sporadic bursts across the debris dam during floods may have destructive impact in the valley below.

In practice, a variety of bedrock failures create the conditions for potentially dangerous debris flows: deep-seated rotational failure develops along carbonate-shale-evaporite slopes; collapse and toppling are found mainly along massive gneiss, quartzite, or carbonate cliffs; slumps occur in sandstone-shale terrain; and sagging slopes are most common in low grade metamorphic rocks.

Deep-seated rotational failures in source areas underlain by carbonate-shale-evaporite complexes commonly are the precursors of debris flows which are characterized by a water-saturated clay matrix and large suspended carbonate blocks (for Alpine case histories see La Valle Agordina, Brannenburg, Perarolo, Bourg-St. Maurice, Brienz, St. Jean-de-Maurienne, Reisskofel, Brentonico, Vandans, Sandling, Tavernerio, Aime, Illgraben, Pettneu).

Collapse of steep cliffs of gneiss, carbonate, or quartzite across narrow torrent gorges may produce serious blockage and debris mobilization during the rainy seasons (for Alpine case histories ses Ziano, Zambana, Werfen, Tagliamento, Meran, Plaine d'Oisans, Bec Rouge).

Slope creep in intensely deformed slate or sandstone-shale successions leads to almost continuous embankment instabilities and local blockage of rain-swollen torrents. Without effective drainage and channel maintenance debris flow activity from such torrent systems can be severe (for Alpine case histories see Schlierengrat, Massif de Platé, Semsales, Chablais, Meiringen).

The most common debris-generating slope failures are found along *sagging metamorphic terrain*. There, large volumes of rock and overlying surficial deposits move slowly along shear zones and foliation planes towards torrent channels. The gigantic volume of sagging slopes in metamorphic rocks locally creates an inexhaustible source of debris for recurrent debris flow activity (Neukirchen, Ardenno, Gastein Valley, Ganderberg, La Chapelle, Obervellach-Möll Valley, Schlanders, Upper Vintschgau, Lienz, Matrei, Aosta, Randens, San Giovanni de Crèvola, Motto d'Arbino, Valle di Vanoi, Rackling, Peccia, Crodo, Nolla, Campo, Putschall, Monte Masuccio, Gaishorn).

MANAGEMENT OF DEBRIS FLOW HAZARD

In the foregoing sections we have outlined the main nonvolcanic settings of potentially destructive debris flows; the case histories in the Appendix show how mitigation and acceptance of risk have been used to deal with different debris flow situations. In general, the management of a debris flow hazard can be approached in three steps: hazard appraisal, passive countermeasures (monitoring and evacuation, land-use zoning), and active countermeasures (remedial engineering, reforestation).

Hazard appraisal

The debris flow hazard of any torrent system can be appraised only semiquantitatively. However, the purpose of such an appraisal is to obtain some estimate of magnitude and recurrence potential for destructive flows in order to design preventive measures. Although similar to appraisals of flood potential along rivers, debris flow hazard appraisals are considerably more complicated and less accurate. Crude at best, they are based on (1) documentation of historical debris flow events and the resulting damage, (2) hazard indicators ('silent witnesses' of Aulitzky, 1972) found on the fan, in the gorge, or source area of torrents, and (3) predictive formulas. The cost of an appraisal has to be balanced against the economics of the case in question and will vary with the sophistication of tools and techniques applied (e.g. drilling, trenching, surveys, detailed geological mapping etc.).

Documentation of historical debris flows

The most significant set of data in the appraisal of debris flow hazard, if available, are historical records documenting actual destructive events and the resulting damage (Aulitzky, 1968). To be able to look back at, say, a hundred years of fairly complete damage records in a mountain range or a major valley generally permits a first-order evaluation of the most dangerous torrents, local debris flow recurrence, and the regional meteorological trigger events. Published historical information concerning recurrent debris disasters exists for some mountain valleys of the European Alps, Apennines, Japan, and other regions where mountain settlements exist close to urban centres. For example, Mougin (1914) produced one of the earliest and still most comprehensive regional inventories of historical debris flows in the Savoie district of the French Alps. His painstaking archival studies revealed not only prevailing weather patterns during many debris disasters, but also documented the adverse impact of human activities such as deforestation and overgrazing on the stability of steep upland basins. The detailed analysis of numerous basins ravaged by erosional scour eventually aided in the restoration efforts (Mougin, 1931). In many parts of the world records of debris flow damage are kept by local engineering offices or can be retrieved from newspaper accounts. In central Europe some of the relevant data have been compiled regionally and published in generalized form (e.g. Montandon, 1933; Stini, 1938; Leys, 1977; Stacul, 1979). Nevertheless, such information files are scarce for areas where the development of mountain valleys is of more recent date (e.g. western North America). Thus, in rapidly urbanizing mountain valleys of the circum-Pacific mountain regions systematic observations made during and after major debris disasters will be of economic significance for future development decisions (e.g. Sharp and Nobles, 1953; Weber and Treiman, 1979; Eisbacher and Clague, 1981). Historical damage records may permit the assessment of total damage from major regional storms, the intensity of precipitation required to mobilize debris, the identification of source areas with debris generating potential, and estimates of possible volumes and velocities of destructive flows.

Hazard indicators

Second in usefulness after historical records are hazard indicators, the erosional and depositional features left behind by past debris flow activity. Hazard indicators (or silent witnesses) on debris fans (cones), in the gorge, or in the source areas of torrent systems yield semiquantitative information concerning the recurrence and magnitude of past flows. They are useful only if they can be related to an active or dormant torrent system, i.e. a torrent system along which recurrent mass movements have taken place repeatedly during Holocene time. If blocks strewn across the surface of a debris fan or erosional scars in the source area are relicts of late Pleistocene processes, they reflect a physical environment quite different from the present, and thus the data are of little relevance to the contemporary debris flow potential of the torrent system. The dating of debris fans is thus paramount in the discrimination of relict from active or dormant torrent systems.

On active or dormant fans the most important hazard indicators are: damaged vegetation, large blocks scattered about, thickness and age of debris layers, and channel gradient. The most common vegetation damage includes scarred bark (Fig. 11), indicating the height of a passing flow and permitting the dating of the flow by an analysis of the new bark growth around scars (e.g. Clague and Souther, 1982); broken trees also yield information on the magnitude of impact forces (Mears, 1977); pioneer vegetation fringing active torrent channels on the cone may allow a crude appraisal of the age of the latest debris flow event (e.g. age of willow and alder tracks within conifer forests); buried root horizons in combination with tree-ring or radiocarbon dates can be used to determine thickness and age of debris flow layers (Jackson, 1977); other vegetation remnants such as buried lenses of peat or tree fragments incorporated in debris flow deposits can serve to obtain maximum radiocarbon ages of debris flow events.

Diameters of blocks strewn about on the surface of active fans or cones, if related to datable debris flow deposits, are meaningful hints as to the transporting and destructive potential of past flows. Blocks on the surface of a deeply incised inactive fan do not indicate a current hazard, although they may point to the existence of potential instabilities in the source area.

Thickness, internal structure and texture of debris layers are indicators of the mechanism of deposition (Nilsen, 1982). Structures such as pebble imbrication, crossbedding, and parallel lamination of sandy lenses indicate relatively tranquil processes of alluvial sedimentation; massive unsorted units of blocky debris indicate generally single debris flow events.

The gradients of active fans or cones indicate the prevailing depositional processes: flat fans (less than 3°) are mainly the result of extensive reworking of debris by running water along braided channels; steep cones (more than 5°) reflect intermittent deposition of debris flows with little reworking by running water. On fan deltas along lakes or along a mountainous seashore the stability of the lower fan

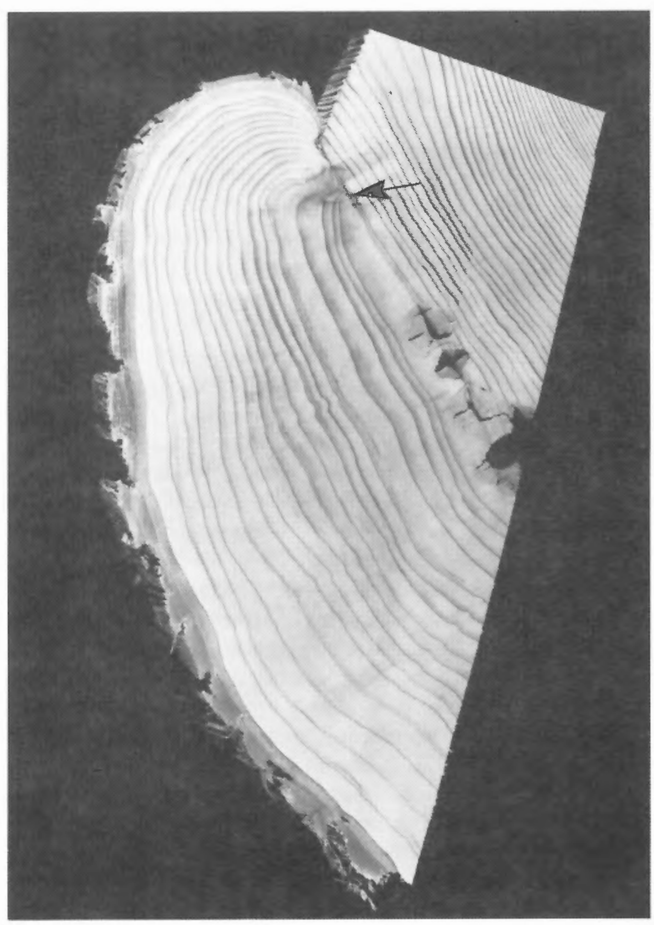


Figure 11: Partial section of a tree damaged by debris flow; arrow indicates time of injury. (GSC 204165-A)

segments has to be appraised as well; existing slump scars along the delta front are certain indicators of inherent instability of shoreline deposits.

Hazard indicators in the gorge of a torrent include patches of debris and scarred vegetation that mark the upper limit of recent debris flows. Sharp vegetation trimlines above the channel often indicate the highest level attained by the moving debris. Along twisting reaches of the gorge debris flows tend to bank and climb the outer bends, a phenomenon known as superelevation. Trimlines on the outside permit determination of superelevation with respect to the position of trimlines on the inside of the bend. A rough estimate of debris flow velocity (V) can be obtained by measuring a) the angle of superelevation perpendicular to the path of the debris flow (θ) and b) the mean radius of curvature of the channel bend at this point (R):

$$V^2 = Rg \sin\theta$$

where g is the constant of gravity.

Hazard indicators in the source areas of torrents are either erosional scars or incipient slope failures. Erosional scars can be recognized by exposed tree roots and debris-charged channel reaches below; incipient embankment failures, bedrock slumps, and active toe zones of sagging slopes can be recognized by irregularly leaning trees, open crown cracks, and sag ponds. Regular surveys which monitor the downhill movement of sagging terrain can establish a value for the volume of debris contributed by the toe zone to the torrent channel. Lakes and ponds with potentially unstable outlets always represent a hazard if they are lcoated above high-gradient torrent systems. Excessive logging and windfall clogging of tributary ravines also increase the debris flow potential of a basin. Similarly, carelessly emplaced sidecast along development roads may cause debris avalanches in a basin.

Hazard indicators, particularly those in the source area, are most appropriately documented on large scale maps (1:25 000 and larger). The geological surveys of several countries have published prototype maps showing relative slope stability for mountain areas which are under development pressures. For example, excellent and readily available maps are those of the Bureau de Recherches Géologiques et Minières of France for the French Alps (Humbert, 1977). A slightly different approach of slope stability mapping has been used to document slope stability by the United States Geological Survey (e.g. Nilsen et al., 1979). Maps of this type are highly relevant for the appraisal of the debris-generating potential of torrent basins (Cotecchia, 1978).

More integrated maps of hazard indicators also show snow avalanche tracks, and existing vegetation patterns (Jeglitsch et al., 1975; Harden, 1976; Kienholz, 1977; Grunder, 1980; Dow et al., 1981; Kienholz et al., 1983).

To be of greatest value hazard indicators should be shown site-specific. The cost of subsequent zoning measures, reforestation, or engineering works can then be evaluated. It is important that the geomorphic-geological evidence derived from hazard indicators is presented in a style comprehensible to all parties involved, including non-experts such as land owners, prospective developers, government agencies, and water resource planners. Experience has shown that where a debris flow hazard is not easily demonstrated, the essential co-operation from the concerned parties may be difficult to obtain.

Predictive formulas

Predictive formulas concerned with recurrence and magnitude of debris flows have been the outgrowth of research into floods and bedload transport from small mountain drainage basins (e.g. Hampel, 1970; Zeller, 1982). However, in many small torrent basins for which historical observations are available the volume of debris flows is difficult to relate to annual storm runoff or bedload transport (Scott and Williams, 1978; Sommer, 1980). This fact is also borne out by the case histories presented in the Appendix. Locally, it may be possible to arrive at predictive quantitative relationships between the magnitude of extreme debris flows and other measurable parameters of the torrent system such as catchment area, forest cover, channel gradients (source area, gorge, fan), fan radius, etc. However, initiation of major debris movements along both high-gradient and low-gradient torrent systems depends on specific geologically controlled slope instabilities and erosional point sources; their size determines the volume of potentially hazardous debris flows. Kronfellner-Kraus (1982) has therefore expressed a warning against the application of oversophisticated prediction formulas after having carried out an analysis of historically documented debris movements from a variety of catchment basins in the Eastern Alps. He found that the volume of extreme flows, although related to the gradient and catchment area of the torrent system, is critically dependent on a factor which reflects the regional geological-physiographic setting. According to his generalized formula (Kronfellner-Kraus, 1982, p. 20)

$$GS = E \times J \times K$$

where GS is the volume of an extreme debris flow (in m^3), E is the area of the catchment basin (in km^2), J is the average gradient of the torrent from source area to the apex of the fan (in %), and K is a regional parameter that varies greatly with the geological characteristics of the source area. The parameter K has been found to range from 500 for large, low-gradient, and well forested basins to 1500 for small, high-gradient, and poorly forested basins. K is also related to the source area mechanics, i.e. it depends on whether bedrock slides, debris slumps, sagging slopes, or erosional scars produce the debris.

Passive measures

Passive measures against debris flow hazard include land use zoning and monitoring. Appropriate land use in mountain valleys is not a precise science and there are no simple rules that apply with equal validity to different cultural domains. But a wise use of information from historical records and hazard indicators can minimize loss of homes, road works, and lives in areas of recurrent debris flow activity. Debris cones and fans have long played a role as preferred sites in the development of mountain valleys and there are indisputable advantages in living on them: the frequent floods of the main valley floor are avoided; building foundations are generally firm and dry; there is a better view of the valley; the mountain torrent supplies clean water for a variety of purposes ranging from irrigation to human use. The apical segments and axial sectors of active or dormant fans, however, are areas of great potential debris flow damage. Only an appraisal of the hazard indicators or historical damage will yield criteria on whether this hazard is acceptable or whether it is unacceptable and thus should be avoided. Prior to settlement such a study may result in recommendations as to the most diligent use of the forest resources of the uplands combined with building restrictions along the torrent and on the cone (e.g. selective logging from generally protected forests, rigorous maintenance of access roads, minimum culvert diameters and bridge freeboards, etc.). Protective forest belts may also be preserved along the apex and the axial sector of the fan (cone), where their protective function is one of retaining the force of minor flows.

In general, passive measures (i.e. zoning) for certain land use must be simple in order to be effective. A relatively simple approach, followed in some countries, is the threefold traffic-light system: a *red zone* of unacceptable risk (e.g. no logging, no road building, or no residential construction), a *yellow zone* with carefully balanced risk (required safety features such as dams, selective logging, full-cut roads, channel revetments or check dams, etc.), and a *green zone* of acceptable risk. The terms 'acceptable' and 'unacceptable' risk are clearly preferable to the terms 'safe' and 'unsafe' and depend entirely on the social-economic framework: near a densely populated housing development an average debris flow recurrence of once in a hundred years may be unacceptable, while the same recurrence interval may be acceptable for a temporarily occupied recreation area (see farther on).

What represents an acceptable or unacceptable hazard has to be understood prior to development of a torrent system. In the broad yellow zone, which bridges areas of unacceptable with those of acceptable risk the land use decisions cannot be objective (Zeller, 1972b; Antoine, 1978). The legal expressions 'right' or 'wrong' for the type of development chosen have little significance when the data base is simply inadequate (Zollinger, 1976).

A special problem is posed by debris flow damage in areas previously considered to have acceptable risk. Another common source of difficulties is a change in land use of the source area after the cone has been developed. If such change in land use creates previously unknown debris movements the only recourse is the implementation of costly active measures. Simple financial compensation for losses to property owners without active countermeasures are bound to result in even greater damage and costs during the next catastrophe. Physical relocation of communities is both expensive and often socially unacceptable.

In summary, passive measures (zoning) are economical if undertaken *prior* to development; more often, however, they have to be balanced against existing historical rights, local traditions, political pressure, and economic considerations, all of which may blunt the eventual zoning plan for an area exposed to debris flow hazards. At best, passive measures may improve forest practices, road construction, and ski area developments in the catchment areas, and inhibit further development of debris fans (cones); at worst they may lead to protracted lawsuits between landowners and zoning authority, or between users of the fan and users of the source area.

Monitoring is a relatively inexpensive, although not entirely satisfactory, passive measure for protecting people against debris flows. Regular surveys of sagging slopes and other bedrock failures in the source area of torrents may provide hints of renewed debris flow activity. Warnings may be issued to inhabitants of fans and temporary evacuation may be required during intense rainstorms. Wire detectors, strung across the torrent channels in the gorge above the fan provide little time between warning and evacuation during debris flow events of great velocity. The manning of observation points along the gorge during storms is another traditional monitoring technique. Monitoring activities only help to save lives; removal of fixed property prior to the impact of the debris flows is possible only in cases where flows move exceptionally slowly.

Active measures

Active (remedial) measures against debris flows are undertaken where human infringement on the fan or debris source area poses an unacceptable risk, but where passive measures (zoning) are precluded by cost of resettlement, value of land, or vested traditional rights. Active measures against debris flows are generally designed according to source area characteristics, debris composition, and fan geometry. In the source area active measures are directed towards *control and stabilization* of debris, on the fan they have primarily *protective* functions.

From an economic point of view, the cost of active measures should be considerably less than the total value of the property to be protected. However, where human lives are threatened, economic considerations may have to be abandoned. According to political preferences the cost of remedial measures may have to be borne by individuals, entire communities, or the public at large.

Historically, active measures against debris flows were first taken near densely settled medieval towns of central Europe and eastern Asia (Strele, 1932; Ikeya, 1976; Stacul, 1979): there, debris disasters often prompted the reinforcement of existing flood dykes or 'water walls' along torrent channels. Aggradation of debris in these channels and continued dyke reinforcement often led to spectacular elevation of the axial fan sector over the rest of the built-over fan.

Since many debris disasters in the medieval towns of the Alps were triggered by deforestation and overgrazing of upland areas, some communities began to implement measures designed to check the erosional source area processes as well: certain forests were declared off-limits or received special protected status ('Bannwald'), often serving a combined function against snow avalanches, rockfalls, and erosion. Locally, steep erosional ravines were stabilized by primitive masonry check dams as early as the late 17th century (Stacul, 1979, p. 84-85). In central Europe the debris

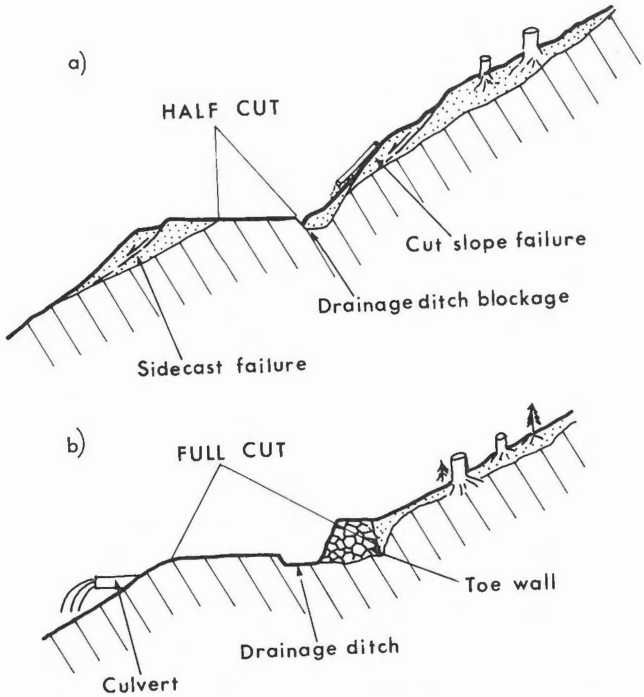


Figure 12: Schematic cross-section through a logging road showing: a) debris generation along poorly designed road cuts and road beds; and b) well designed access road.

flow problem was clearly perceived after the widespread collapse of agriculture and depopulation of mountain valleys after the Napoleonic wars. An early scientific literature soon pleaded for better forest practices (e.g. Aretin, 1808; Surell, 1841), and a pioneering treatise on the function of check dams in controlling unruly torrents above endangered settlements was published about the same time (Duile, 1834). Following a series of debris disasters in the Alps in 1859, 1868, and 1882 active measures against debris flows became part of the work carried out by regional forest services (Salis, 1892; Mougin, 1931; Strele, 1932). Active measures also have a long tradition in other parts of the world where terrace agriculture and runoff control are as old as civilization (Ikeya, 1976; Johnson et al., 1982). Today expensive engineering works are more and more complemented by less expensive and environmentally pleasing bioengineering techniques (Schiechtl, 1980; Gray and Leiser, 1982). Active measures fall broadly into three groups: (1) improvement of forests and reforestation in debris source areas, (2) construction of control works, and (3) erection of protective structures on and above the cones.

Forest practices and reforestation

The beneficial function of an intact forest on the stability of slopes stems not only from the ability of trees to remove water and thus lower pore pressure in the slope by evapotranspiration and by interception of rainfall, but also from a reinforcement of surficial deposits by tap roots and secondary sinker roots of individual trees (Swanson and Dyrness, 1975; Rice, 1977; Wu et al., 1979; Gray and Leiser, 1982, p. 37-65). Complete removal of forest thus lowers the threshold values of rainfall intensity and rate of snowmelt necessary to mobilize unconsolidated surficial deposits and other debris on steep slopes. By how much these thresholds are lowered depends on the removed tree species, aspect of the slope, geology of the site and other parameters which have to be determined in detail. During truly catastrophic rainstorms shallow slope failures will occur regardless of the vegetation cover.

In logging zones most debris is generated along unmanaged or poorly managed access roads. The most common debris-generating mechanisms along forest roads are sidecast failures, retrogressive cut-slope failures, and erosional scour near plugged culverts (Swanston and Swanson, 1976; Megahan, 1977). These three mechanisms commonly act together where forest roads cross ravines, gullies, tributary torrents, or other topographic re-entrants. Here construction, maintenance, and runoff control are of critical importance (Burroughs et al., 1976). Full cuts and complete removal of the excavated material are advisable. Culverts should be of large diameter, unobstructed, and oriented parallel to the pre-existing drainage: culverts at angles to natural channels generally are soon plugged by debris and large erosional gashes cross the roads. Gabion toe-walls may have to be used locally to keep cut-slopes from ravelling into drainage ditches (Fig. 12). Reforestation of potential sources of debris may require artificial terracing and temporary debris retaining structures (*see* Schiechtl, 1980, for details).

A most significant aspect of proper forest practice is the maintenance of access roads, roadside ditches, and culverts after logging has ceased. Removal of logging debris from tributary ravines and torrent channels is also advisable; this reduces the potential volume of debris that might be set in motion during rare storm events. Improved logging practices, proper construction of access roads, and reforestation will, over time, reduce the potential of debris generation from logged slopes. However, one has to keep in mind that even careful logging disturbs the natural setting and that economic considerations always enter into the long-range development plan for a logging zone. The economics of logging practices therefore have to be balanced against inevitable losses to fisheries or tourism. The desired balance should be decided for every case; there are no fixed rules that can be applied for every natural situation or social context.

Control works

These are designed to stabilize major potential debris sources and thus prevent the mobilization of debris into coherent flows. In detail, the type of debris source will dictate the type of remedial measures chosen. The principal debris sources of mountain torrent systems are erosional scars, debris slumps, bedrock slides, and sagging slopes.

Erosional scars (e.g. along high gravel embankments) are commonly reactivated during local downpours by scour of running water. Torrent reaches bordered by erosional scars are most effectively stabilized by stacked arrays of check dams (Fig. 13). The art of check dam design is old and the principles simple (Duile, 1834; *see* Stacul, 1979, for

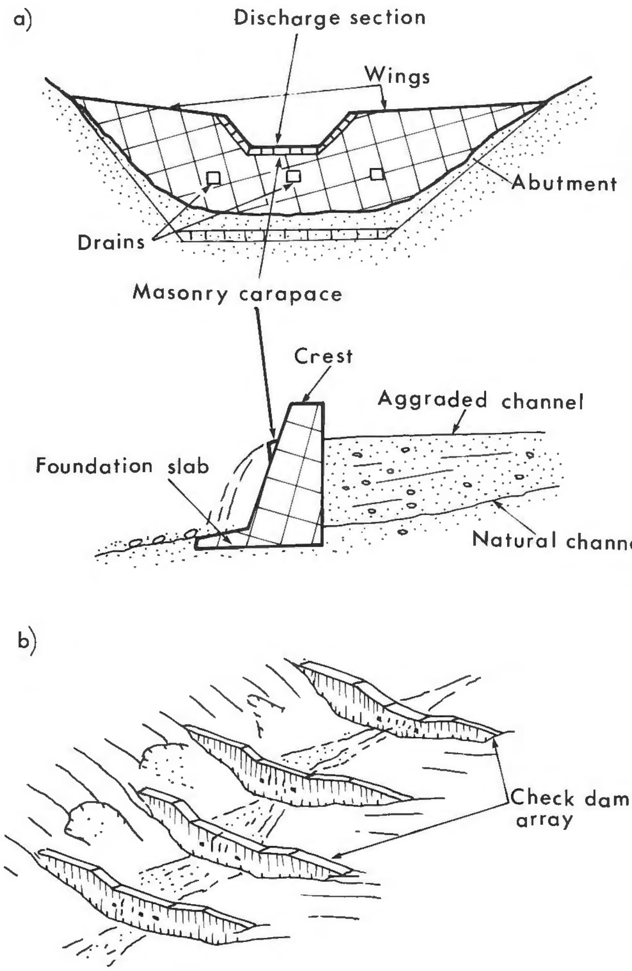


Figure 13: a) Diagrammatic views of the elements of an individual check dam; b) The arrangement of check dams in stacked arrays along torrent reaches bordered by erodible and unstable surficial deposits.

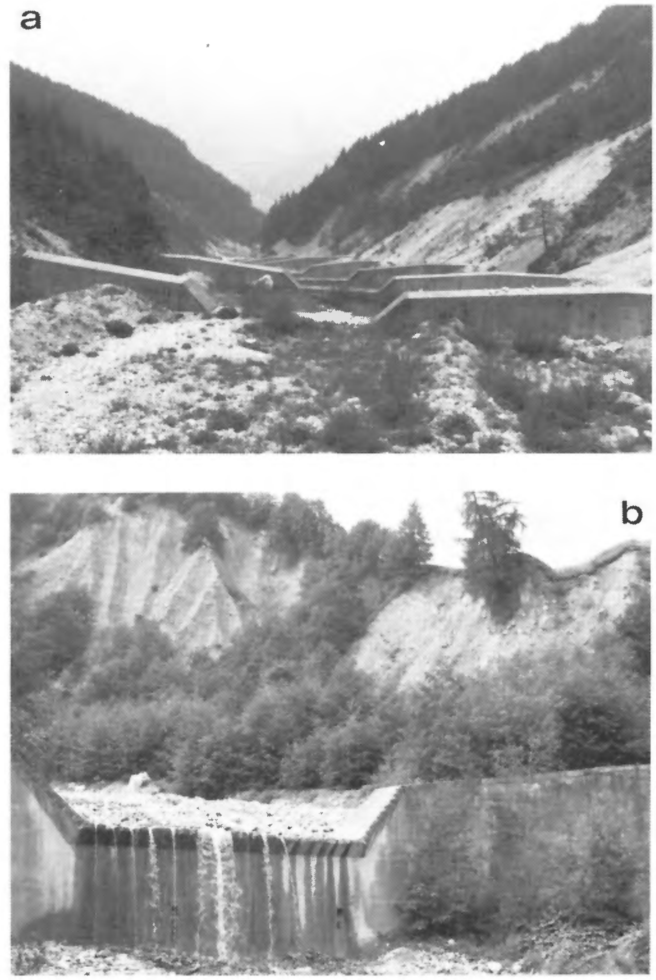


Figure 14: Check dams along torrent reaches with unstable debris-generating embankments: a) near Predazzo, Italy (GSC 204165-B); b) near Bludenz, Austria (GSC 204165-C).

review): the rising wings of check dams keep the flow of the torrent to the central axis of the channel and thus prevent lateral and vertical erosion of the unstable bed; aggradation behind check dams adds stability to the toes of embankment slopes on the upstream side; the rush of water is broken by the stepped channel profile; and the gently curved discharge sections of the check dams facilitate unobstructed passage for minor debris floods and flows without endangering the stability of dam abutments (Fig. 14). Depending on the required strength of check dams the material used in their construction may be steel-concrete, masonry, metal-concrete cribs, gabions, or timber (Strele, 1950; Stacul, 1979; Heede, 1980). The construction of check dams is generally supplemented by revegetation of scarred embankments with fast-growing, deep-rooted trees (e.g. *Alnus, Salix*). In built-over areas erosional reaches may have to be stabilized by the complete training of torrent channels by revetments of concrete-masonry or rip-rap.

Debris slumps and minor bedrock failures in the source area of torrents are stabilized most economically by drainage or retaining structures, again supplemented by revegetation. Numerous technical details for the stabilization of natural debris slumps and slides are described by Záruba and Mencl (1982) and by Gedney and Weber (1978). Considering that many source areas of mountain torrents are poorly accessible, the most common technique for the stabilization of slumps is the use of gabion retaining walls placed against the toe of the unstable debris mass. However, such stabilizing berms may only be of temporary value and may have to be replaced by more substantial structures.

Bedrock slides and toe zones of sagging bedrock slopes are most economically improved by meticulous rerouting of surface runoff away from the crown cracks, by draining sag ponds, and by removing water from the slope by external (ditch) or internal (pipe) drains. Toe retaining walls, concrete berms, or dams may be feasible locally, but tend to be expensive because of the required size and strength.

Emplacement of check dams at actively eroding toe zones of large debris slumps or sagging bedrock slopes is only partly effective: over time such transverse channel structures are either buried or crushed by the encroaching slide mass; recently, check dams with compressible gabion-type discharge sections have been found suitable in the control of erosion along bedrock-debris slumps (Kronfellner-Kraus, 1982). Where infiltration of water into unstable slopes is caused by the torrent itself, rerouting of the torrent by diversion tunnels or cuts may be necessary. A dense mantle of vegetation may improve the hydrological balance of the slope and thus enhance its stability.

Protective works

These are built along the lower reaches of torrents, particularly at the apex of debris fans or cones; their function is to reduce the direct impact of debris flows on human structures. The 'design events', which serve as a basis for the choice of protective works, are derived from a careful hazard appraisal. However, costs for protective works are relatively high, and the economics of certain situations may preclude their use. To avoid a false sense of security, the magnitude of the 'design event' and the uncertainty of obtaining the projected volume and velocity therefore should be spelled out clearly prior to the onset of construction. No protective structure can be considered *absolutely* safe, and there always remains a residual acceptable hazard.

Because of their momentum and their non-Newtonian behaviour, debris flows in natural settings debouch along straight and flaring paths close to the axial sector of fans. Protective structures should be emplaced in consideration of this peculiarity: sharp bends or depressions in the path of a debris flow are filled quickly by early pulses of debris and are thus easily overtopped by later pulses. The simplest design criterion therefore is to provide a straight, uninhibited, and flaring course across the fan towards the receiving river. To avoid a potentially serious buildup of debris immediately above the apex of the fan it may even be necessary to remove blocking bedrock spurs above the mouth of the gorge.

Protective works include, in their order of increasing cost, forest belts and dykes along torrent channels, deflection dams, transverse dams, debris retention basins, and galleries (sheds).

Forest belts along a dyked torrent channel may offer satisfactory protection for a sparsely inhabited fan of low gradient; rows of closely spaced trees may impede the lateral breakout of coarse debris from the axial sector of the fan during minor flows. However, it is important that both channel and forest belt are entrenched somewhat into the surface of the fan. This may require widening and lowering of the axial sector of the fan prior to dyking and planting of tree belts.

Deflection dams and *guiding walls* are generally built at the apex of debris fans whose axial sectors have been built over, but whose flanks or axial sector still provide sufficient space for a wide and straight torrent track (Fig. 15). A deflection dam should be sufficiently strong and high so that it is not breached or overtopped by debris flows; if curved its radius of curvature should be as large as possible to reduce the impact of the flow on the structure. In general, some of the material used in the construction of the deflection dams can be derived from the excavated blocky debris found along

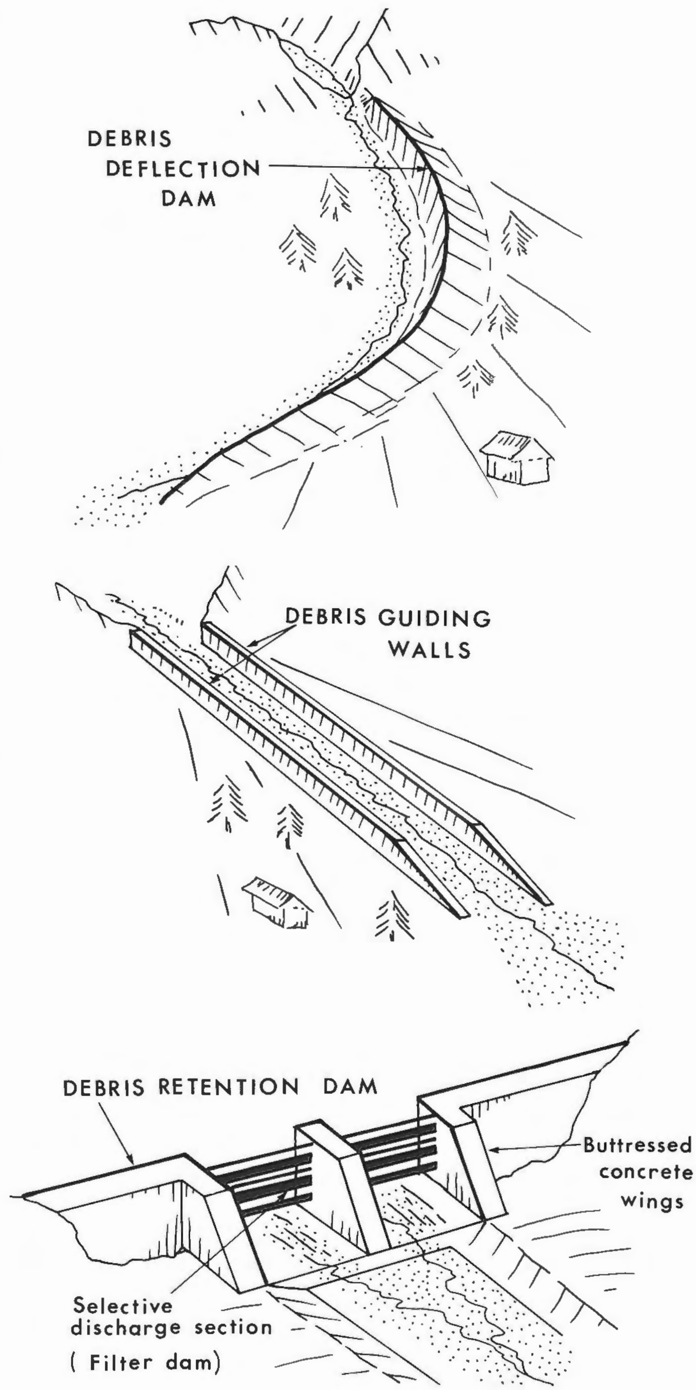


Figure 15: Schematic illustrations of the three principal types of protective engineering works against debris flows.

Figure 16: Debris deflection dam with strong stone-block armour (near Chiavenna, Italy). (GSC 204165-D)

the torrent channel. The curved side of the dam facing the torrent generally has to be reinforced with a stone-block armour (Fig. 16); tree belts may supplement the structure to soften its visual impression and to allow for the inevitable spillover of fines during flows.

Transverse debris retention dams are erected across the torrent at or immediately above the apex of a debris cone or fan (Fig. 17). Modern transverse protective structures incorporate steel, concrete, and masonry. The foundation sills and abutments are preferably emplaced on bedrock or are at least anchored to the channel substratum by sheet piles. Since they have to withstand the direct thrust of the moving frontal debris wall the design has to be safe against shear and toppling of the structure. The concrete wings are therefore generally provided with downstream-pointing steel or concrete buttresses (Fig. 17c). The central discharge section of a protective dam consists either of steel or steel-concrete beams. The slits or openings between the beams of the discharge section are wide enough to permit passage of normal bedload through the dam, thus preventing premature aggradation behind it. However, they are narrow enough to block a potentially destructive debris flow. Screens, rakes, and braced steel grids may serve as central discharge sections where design of the dam is directed mainly against organic (timber) debris. The crest of the dam is commonly provided with a strong masonry armour and downstream-projecting carapaces that prevent damage to the structure due to spillover of debris when, during debris flow events, the space behind the dam fills up. Road access to the retention space on the upstream side of the dam permits maintenance and cleanup after debris flows. Available space at the apex of debris cones and/or the economics rarely permit construction of protective dams against design debris flows in excess of 300 000 m^3.

Galleries are built over highways which cross steep debris cones; they are commonly supplemented by debris guiding walls or retention basins (Fig. 18).

Figure 17: Examples of protective debris retention dams: a) masonry-lined steel concrete dam with vertical selective discharge slits under construction (near Innsbruck, Austria) (GSC 204165-E); b) steel-concrete dam; note drains in the lower half and selective discharge section of steel beams above (near Caoria, Italy (GSC 204165-F); c) buttressed selective discharge section of an earth-rockfill dam (near Innsbruck, Austria) (GSC 204165-G).

Figure 18: Combination of debris retention basin and gallery designed to protect major mountain highway from recurrent debris flows (Arlberg Highway, Austria). (GSC 204165-H)

MASS MOVEMENTS ON VOLCANOES

Severe slope stability problems occur on Quaternary volcanoes, especially those that are active or dormant. Worldwide in this century alone, thousands of deaths and enormous property destruction have resulted from volcanic debris flows. For example, in 1919 a single debris flow from Kelut volcano in Indonesia destroyed or damaged 104 villages, killed over 5000 people and ruined thousands of hectares of cultivated land. In this chapter we summarize the slope stability problems associated with young volcanoes and discuss the premonitory and remedial measures that may be employed to reduce the risk of destructive mass movements in these areas.

It must be emphasized at the outset that the mass movements of concern here are restricted to a relatively small portion of the Earth's surface; well over half of the roughly 10 000 recently active volcanoes of the world are found in less than 0.3% of the Earth's land areas (Fig. 19; Williams and McBirney, 1979). Almost all active and recently active volcanoes are located at or near lithospheric plate boundaries. The majority are associated with convergent plate boundaries and occur above subduction zones. For example, the "Ring of Fire", a necklace of active volcanoes girdling the Pacific Ocean, closely follows several subduction zones encompassing New Zealand, New Guinea, the Philippines, Japan, Kamchatka, the Aleutians, western North America, Central America, and western South America. Likewise, the volcanoes of the Indonesian archipelago are associated with a major subduction zone in the Indian Ocean that connects eastward with the Pacific volcanic belt. The volcanoes of the Lesser Antilles and southern Europe also are located along convergent plate boundaries. Most of the volcanoes occurring at convergent plate boundaries are stratovolcanoes formed by repeated explosive eruptions of silica-rich magmas.

Volcanoes also occur at divergent plate boundaries where new crust is created by the upwelling of magma along spreading ridges, and in the middle of plates where magma is generated above "hot spots" in the Earth's mantle. The volcanoes of Iceland on the Mid-Atlantic Ridge exemplify the former group, and Hawaiian volcanoes are examples of the latter. Volcanoes located at divergent plate boundaries characteristically are silica-poor and relatively nonexplosive; they consist mainly of lava flows and have less pyroclastics than volcanoes at convergent margins.

Mass movements tend to be very common on Quaternary volcanoes at convergent plate boundaries for several reasons. Firstly, the volcanic materials associated with silica- and volatile-rich eruptions are weak and prone to failure on moderate and steep slopes. Pyroclastic deposits ("tephra"), the dominant constituents of most stratovolcanoes, are loose and poorly indurated volcanic sediments that are easily eroded by running water and glacier ice and also are susceptible to slope failure. Lavas, dykes, sills, and plugs commonly are intensely fractured and jointed, and consequently also have relatively low strengths. Soils separating eruptive units on most volcanoes are themselves potential failure surfaces.

Secondly, young volcanoes at convergent margins generally have moderate to high local relief, a requisite for large mass movements. A volcano, of course, becomes elevated above the surrounding terrain by the eruption of flows and pyroclastics from vents and fissures. The morphology of a volcano, the steepness of its flanks, and the shape of its summit vary in a complex fashion as a function of eruptive history and magma chemistry. In general, relatively steep depositional slopes are found on stratovolcanoes and pyroclastic cones formed from silica-rich magmas. High local relief on volcanoes also may result from erosion by streams, glaciers, and eruptive processes. Erosional ravines and valleys on the flanks of volcanoes commonly have steep walls that are susceptible to failure.

Thirdly, many volcanic mountain ranges receive abundant precipitation which reduces the stability of slopes. Torrential rains commonly trigger debris flows in loose pyroclastic materials. In other instances, eruptions melt summit glaciers or empty crater lakes to produce enormous flows that travel many tens of kilometres.

Fourthly, young volcanoes generally are located in zones of high seismicity (Fig. 19). Earthquates which either accompany eruptions or are unrelated to them may initiate a wide variety of mass movements on the flanks of volcanoes.

Finally, eruptions are themselves direct causes of landslides. The upward movement of magma in a vent may deform and weaken the volcano and eventually lead to flank failures. In addition, volcanic ejecta falling back on the cone may become mobile and flow downslope.

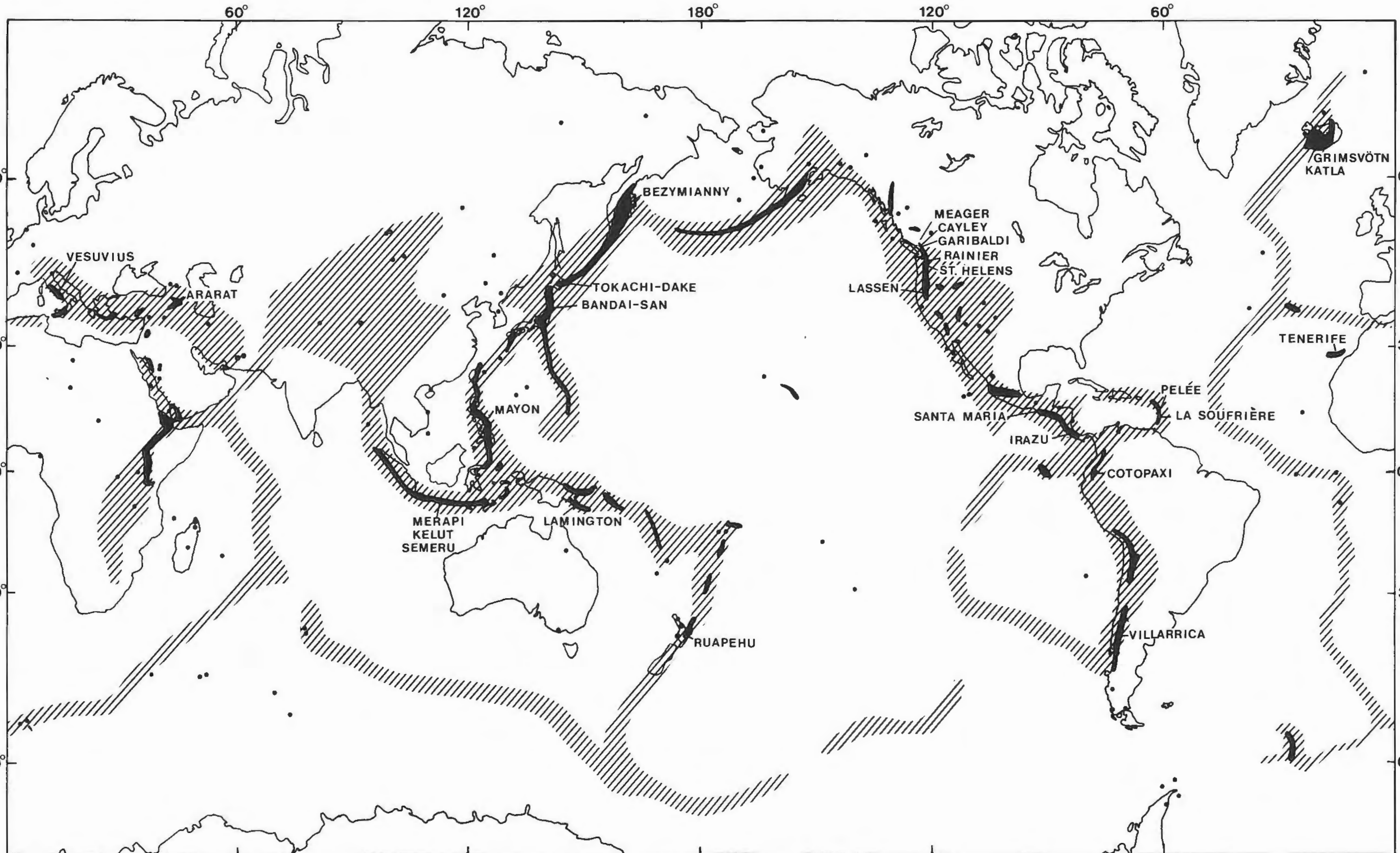


Figure 19: Worldwide distribution of Holocene volcanoes (darkened areas) and major earthquake zones (lined pattern). Individual volcanoes outside of major volcanic belts are shown as dots. Volcanoes mentioned in the text are named and located on the map. (Volcano data from Simkin et al., 1981; earthquake data from United States National Oceanic and Atmospheric Administration, Environmental Data and Information Service).

Types of volcanic mass movements

Mass movements on volcanoes may be broadly subdivided into two groups, one consisting of failures triggered by eruptions and the other comprising failures unrelated to eruptions (Table 2). The most common types of mass movements in both groups are debris flows and debris avalanches consisting of pyroclastic material and blocks of crystalline volcanic rocks. Debris flows generally have higher water contents than debris avalanches and flow in the manner of a viscous fluid, but the two types of slope failures are otherwise similar. Many debris avalanches begin as slides or slumps and transform into debris flows while moving downslope. Volcanic "mudflows" also are similar to debris flows, but consist dominantly of dust, ash, and lapilli, with little coarse particulate matter. Debris avalanches, debris flows, and mudflows occurring during eruptions are often hot and of very large size ($>10^7$ m^3); avalanches and flows unrelated to eruptions are cold and range widely in size (most are 10^5–10^7 m^3).

Cold volcanic debris flows may be subdivided according to their mode of formation. Some flows result from the failure of a discrete mass of pyroclastic material and/or lava above a well defined rupture surface. As this slab moves down the flank of the volcano, it rapidly disintegrates into a melange of dust, ash, lapilli, and blocks. Such debris flows commonly are triggered by earthquakes, by heavy rainfall or glacier melt, and by swelling of the cone due to the upward movement of magma. A second major group of cold volcanic debris flows (as well as many mudflows) results from the erosion of tephra by running water. Loose pyroclastic sediments may be scoured by runoff during heavy rainstorms and thus mobilized into debris flows. Such flows are little different from those involving glacial, colluvial, and other types of unconsolidated sediments in nonvolcanic mountain ranges.

Debris flows and mudflows are water-bearing and define one end of the spectrum of volcanogenic flows. At the other end of this spectrum are relatively dry flows consisting of incandescent pyroclastic debris mixed with hot air and volcanic gases (i.e. "pyroclastic flows"). Where the gaseous phase dominates, such incandescent flows are termed nuées ardentes.

Nuées ardentes and other pyroclastic flows

When viscous, gas-rich magma erupts vertically under fairly low pressure, incandescent volcanic ejecta may fall back on the volcano and flow downhill as a fiery gaseous avalanche (Fig. 20). Similar superheated clouds of gas and particulate matter also may result from a directed lateral blast. Such swiftly flowing and turbulent hot tephra-laden clouds are known as nuées ardentes, literally "glowing clouds". Most of the solid material in a nuée ardente flows near the ground, often in a confined channel such as a valley. Above this, and commonly covering a much larger area, is an ash-laden hot

Table 2. Types and characteristics of volcanic mass movements

Type	Temperature (°C)	Water content	Gas content	Solid constituents	Relationship to eruption
Pyroclastic flow (including nuée ardente)	>100 <1000	Low	High	Pyroclastics	During
Hot debris flow and mudflow	30–100	High	Low	Pyroclastics, crystalline volcanics[1]	During
Cold debris flow and mudflow	0–30	High	Low	Pyroclastics, crystalline volcanics[1]	During or unrelated
Rock avalanche	0–30	Low	Low	Crystalline volcanics	During or unrelated
Debris avalanche	0–30	Low	Low	Pyroclastics	During or unrelated
Rockfall	0–30	Low	Low	Crystalline volcanics	During or unrelated

[1]The solid constituents of mudflows are mainly dust, ash, and lapilli; in contrast, debris flows consist of both fine and coarse pyroclastic material, or a mixture of such material and fragments of lava flows, domes, plugs, dykes, or sills.

cloud. This cloud has tremendous mobility and may sweep across ridges and spread out over extensive areas at the base of the volcano. Much of the death and destruction resulting from nuées ardentes are due to these superheated clouds rather than the dense channelized pyroclastic flows that accompany them. The velocities of nuées ardentes are controlled largely by the direction of the eruption blast. If the blast is vertical, the speed of the cloud is determined by gravity and rarely exceeds 45 m/s; in a lateral blast, the initial velocity may reach 300 m/s. Although individual nuées ardentes vary tremendously in temperature, debris concentration, and volume depending on the size and character of the eruption and the morphology of the volcano, all share the characteristics of great mobility and resultant high velocity.

The best known and most destructive nuée ardente is that which destroyed the city of St. Pierre, Martinique, on 8 May 1902 (Fig. 21; Hovey, 1902; Heilprin, 1903; Lacroix, 1904; Bullard, 1976). This catastrophe resulted from an eruption of Mount Pelée, located about 8 km from the city. During this eruption, a great cloud of ash rose many kilometres above the volcano, and simultaneously another cloud shot laterally southward through a notch in the crater wall towards St. Pierre. In less than two minutes it engulfed the city, killing its 30 000 inhabitants. The cloud, travelling about 45 m/s, struck St. Pierre with tremendous force. Stone and cement walls 1 m thick were knocked over and torn apart, big trees were uprooted, and a 3 tonne statue was carried 12 m from its base. Much of the city caught fire, and most of the ships in the harbour sank or were destroyed by fire. The temperature of the blast can be estimated from its effects on objects that did not burn. Glass was softened (650–700°C), but the melting point of copper (1058°C) was not reached. Death apparently resulted from inhalation of highly heated gases. The May 8th catastrophe was followed by sporadic nuée ardente eruptions for several months. Most affected the same area as the May 8th nuée ardente, but one on August 30th extended somewhat to the east of the previous ones and destroyed five villages, adding 2000 victims to the death toll of Mount Pelée (Fig. 21).

Other well documented eruptions that have produced spectacular nuées ardentes are La Soufrière on the island of St. Vincent, Lesser Antilles, in 1902 (Anderson and Flett, 1903; Anderson, 1908); Mount Lamington, Papua-New Guinea, in 1951 (Taylor, 1958); Bezymianny volcano, Kam-

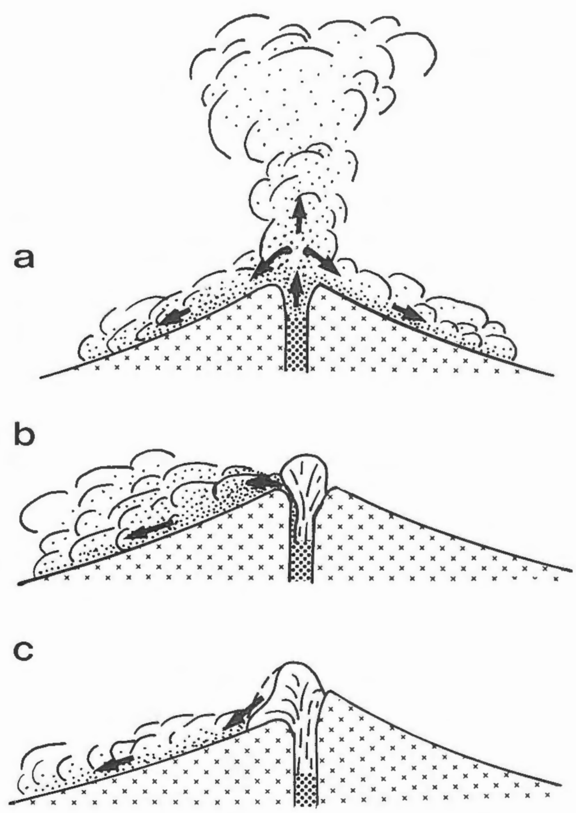


Figure 20: Mechanisms of nuées ardentes formation: a) fall of volcanic ejecta onto the flanks of a volcano from a vertical eruption cloud (e.g. La Soufrière); b) laterally directed blast resulting from the plugging of a vent by a lava dome (e.g. Pelée); c) collapse of an incandescent lava dome (e.g. Merapi). (Adapted from Macdonald, 1972, Fig. 8-2).

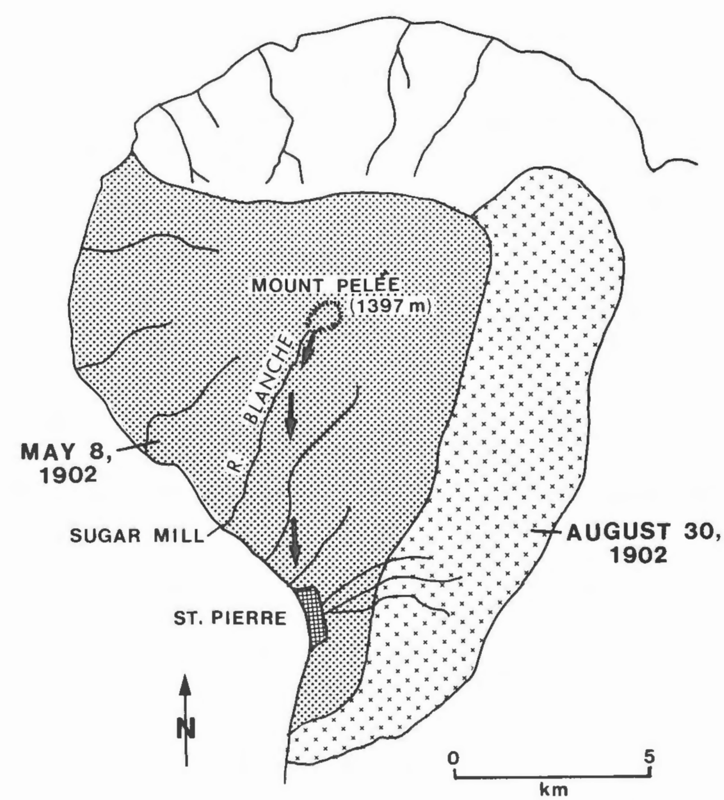


Figure 21: Map of the northern end of Martinique showing areas devastated by eruptions of Mount Pelée on 8 May and 30 August 1902. Arrows indicate the path of the glowing cloud that destroyed St. Pierre. (Adapted from Bullard, 1976, Fig. 18; 8 May area after Hovey, 1902; 30 August area after Heilprin, 1903).

chatka, in 1955-1956 (Gorshkov, 1959); and Mayon volcano, Philippines, in 1968 (Moore and Melson, 1969).

During the eruption of Mount St. Helens, Washington, on 18 May 1980, a hot (ca. 500°C) ash cloud, somewhat different from a classic nuée ardente, spread over a fan-shaped area north of the volcano, devastating an area of almost 600 km^2 (Lipman and Mullineaux, 1981). This cloud was generated by massive explosions that occurred when an enormous landslide released the confining pressure on a shallow magma body and its associated hydrothermal system. Propelled by expanding gases and gravity, a mixture of gas, rock, and ice moved off the volcano as a hot, ground-hugging, turbulent pyroclastic cloud at velocities of as much as 300 m/s. Within a few minutes, the directed blast had extended to about 25 km from its source and carried off or knocked down all trees in its path. The cloud remained low, extending only a few kilometres above ground level, until forward motion diminished, whereupon convecting ash clouds began to billow up. The deposit emplaced directly by the blast cloud was thin, varying from about 1 m near the volcano to only about 1 cm near the margins of the affected area. About half of the material consisted of fresh dacitic debris; the remainder was chiefly mixed lithic fragments from the north flank of the volcano and entrained soil and organic detritus.

Finally, the volcano Merapi in Indonesia has produced many highly mobile flows of incandescent volcanic debris accompanied by lighter clouds of hot air, volcanic gases, and dust. These flows differ from the more common explosion-generated nuées ardentes in that they originate from the collapse of a growing lava dome extending over the rim of the crater (Fig. 20; Neumann van Padang, 1933; van Bemmelen, 1970; Macdonald, 1972).

Many flows of incandescent tephra or lava blocks are not accompanied by glowing clouds. There is, in fact, a continuum between pyroclastic flows lacking hot ash clouds and true glowing clouds with a basal layer of fiery pyroclastic debris. Pyroclastic flows commonly have tremendous mobility due to the presence of abundant gas which buoys up and separates solid particles. Much of this gas is incorporated at the time the flow is initiated, but some is added later by degassing of tephra and entrapment of air beneath the onrushing avalanche (Macdonald, 1972).

Pyroclastic flows probably have a turbulent regime in their upper gas-rich parts and a laminar regime close to the base where the density and effective viscosity are higher (Williams and McBirney, 1979). They behave as fluidized systems in which solid particles are suspended in hot expanding gases. These flows decelerate and begin to deposit debris along their paths as basal parts lose gas by upward escape, thus increasing frictional drag along the ground. As gases rise, the upper parts of a flow become increasingly fluidized, so that they tend to travel farther and faster than the lower part. Each higher part of a flow therefore tends to override the part below, and each in turn becomes a basal layer that loses velocity and deposits fragments of all sizes together.

Mass movements related to volcanic eruptions

Debris flows and debris avalanches triggered by volcanic eruptions typically are heterogeneous mixtures of tephra and lava fragments ranging widely in size; in some instances the debris includes vegetal matter and blocks of ice and snow. Mudflows, although lacking significant detritus coarser than lapilli-size, form in the same manner as debris flows and have similar dynamics.

Volcanic debris flows are generated on the flanks of volcanoes and move downhill at a velocity dependent on the steepness of the slope, the freedom from obstructions, and the viscosity of the flowing mass. Speeds of 10 to 15 m/s are common, and some of more than 30 m/s have been recorded (Macdonald, 1972). The distance travelled also depends to a large extent on the nature of the terrain traversed, although the volume of the flow also is an important variable in this respect. Commonly, debris avalanches and flows originating on the steep slopes of a volcano come to rest near the base of the cone. Others, however, travel great distances from the volcano: many extend 10 to 20 km from their sources, and some more than 200 km. Individual flows may have huge volumes and cover large areas. For example, the pre-historical Osceola debris flow, which originated on Mount Rainier, Washington, had a volume of about 500 $\times$ 10^6 m^3 and covered an area of over 300 km^2 (Fig. 22; Crandell and Waldron, 1956; Crandell and Mullineaux, 1967; Crandell, 1971). Flows that follow valleys are long narrow tongues, but on reaching surfaces of low relief, they spread out as broad sheets.

Debris flows and debris avalanches triggered by volcanic eruptions may be either hot or cold depending on their mode of formation. They originate in a variety of ways; six of the more common formative mechanisms are discussed below and illustrated with historical examples.

Emptying of crater lakes

The Indonesian volcano Kelut is notorious for causing debris flows by violently expelling its crater lake. In 1919, an eruption through Kelut's crater lake spilled about 38 $\times$ 10^6 m^3 of water down the south and southwest slopes of the volcano. The water picked up great quantities of loose volcanic debris to form a series of debris flows that travelled as much as 38 km from the crater (Fig. 23; Kemmerling, 1921). The first flows were cold, but as activity continued and the proportion of water diminished, they became hotter and finally gave way to pyroclastic flows. Though the flows lasted only 45 minutes, they buried about 131 km^2 of farmland, partly or completely destroyed 104 villages, and killed over 5000 people. Debris flows of similar origin occurred during the early phase of the devastating eruption of La Soufrière in 1902 (Anderson and Flett, 1903; Anderson, 1908).

Sometimes, the release of a crater lake is only indirectly related to volcanism. For example, three days before the cataclysmic eruption of Mount Pelée on 8 May 1902, a steaming lake in the crater of the volcano breached a plug of tephra and emptied into the gorge of the Rivière Blanche (Fig. 21; Hovey, 1902). The near-boiling waters swept towards the coast, 7 km away and almost 1000 m below, quickly transforming into a debris flow up to several hundred metres wide and 30 m deep. Upon reaching the coast, it engulfed a sugar mill and entombed about 30 workers beneath debris so deep that only the top of the mill's tall chimney remained uncovered.

Some crater lakes empty by massive failure of enclosing walls. This may occur either during eruptions, as for example on Semeru volcano, Indonesia (Baak, 1949), or during periods of volcanic inactivity, as for example in 1953 on Mount Ruapehu, New Zealand (O'Shea, 1954).

Rapid melting of snow or ice

In 1877, a large eruption of Cotopaxi volcano in Equador produced pyroclastic flows that cut 30 to 60 m deep channels in the glaciers at the 5897 m summit of the mountain (Wolf, 1878; Miller et al., 1978). The liberated water changed the pyroclastic flows into mammoth debris flows that swept down all sides of the mountain at speeds up to 22 m/s. From the base of the volcano, these flows were funnelled into river valleys where they caused extensive destruction. Hundreds of people and thousands of livestock were killed; the lower part of the town of Latacunga, about 30 km southwest of the volcano, was destroyed. One debris flow from the north side of the mountain travelled a distance of over 250 km in 18 hours and reached the Pacific Ocean via the Río Pita and Río Guayllabamba.

During the 1955 eruption of Bezymianny volcano, Kamchatka, pyroclastic flows spread over a large area on the east flank of the cone. The hot pyroclastics melted snow and thus provided the mobility necessary for debris flows to travel more than 85 km (Gorshkov, 1959; Williams and McBirney, 1979).

Destructive mudflows and debris flows developed within minutes of the beginning of the eruption of Mount St. Helens, Washington, on 18 May 1980 (Fig. 24). Hot pyroclastic debris melted snow and ice on the upper slopes of the cone, producing flows that travelled down the major valleys on the east and west flanks of the mountain, locally attaining velocities of as much as 45 m/s (Janda et al., 1981). However, the largest flow developed many hours later by remobilization of water-saturated landslide deposits in the North Fork Toutle Valley.

Lava flows coming into contact with ice and snow also can generate debris flows. For example, in 1915 lava issuing from the summit crater of Mount Lassen, California, came into contact with snow on the upper east flank of the peak. Water released from the melting snow scoured nearby unconsolidated volcanic debris to create a debris flow that moved down two valleys for a distance of more than 30 km (Day and Allen, 1925; Finch, 1930; Harris, 1976). More recently, in 1963, a lava flow from the crater of Villarrica volcano in Chile melted snow and ice near the summit, thus triggering a debris flow that rushed down the mountain and onto the plains to the west. This flow greatly damaged fields and destroyed a village on the lower flank of the volcano.

Many of the more than 55 Holocene debris and mudflows originating on Mount Rainier in Washington probably have resulted from rapid melting of summit snow and ice

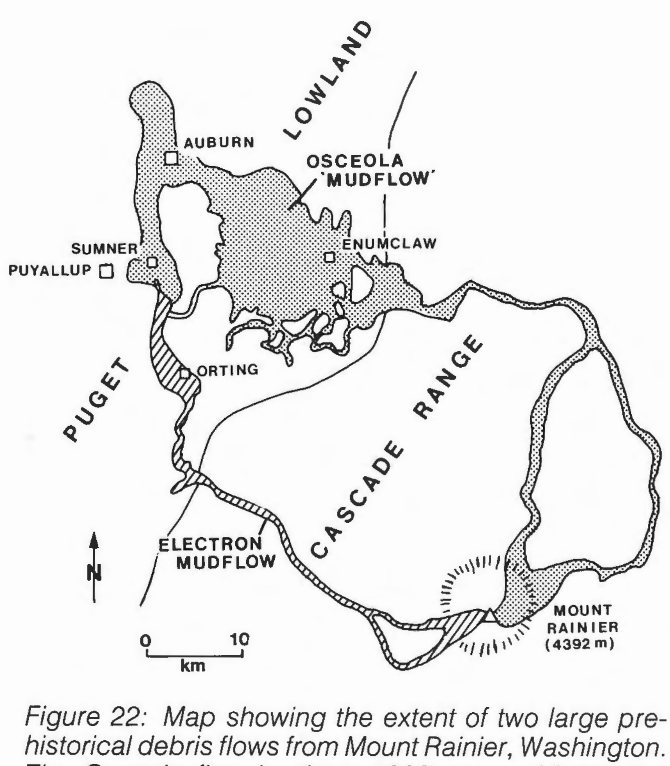


Figure 22: Map showing the extent of two large prehistorical debris flows from Mount Rainier, Washington. The Osceola flow is about 5000 years old and the Electron flow about 500 years old. (Adapted from Crandell et al., 1979, Fig. 7.5).

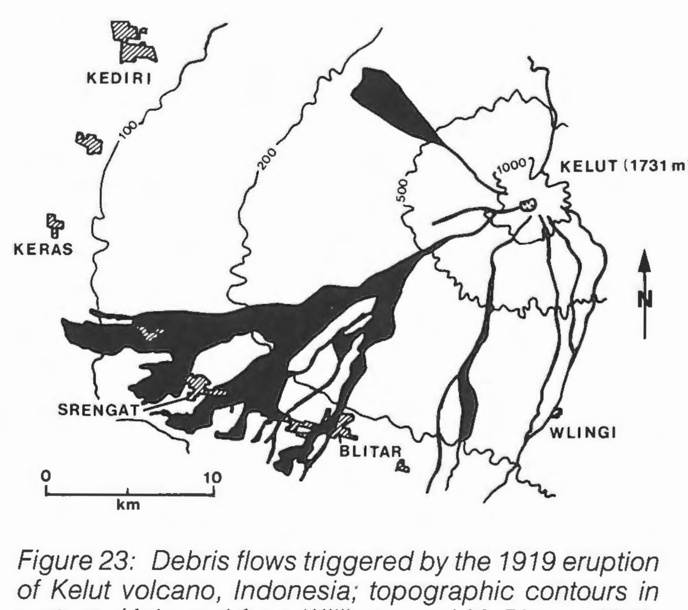


Figure 23: Debris flows triggered by the 1919 eruption of Kelut volcano, Indonesia; topographic contours in metres. (Adapted from Williams and McBirney, 1979, Fig. 7-7; original source Kemmerling, 1921).

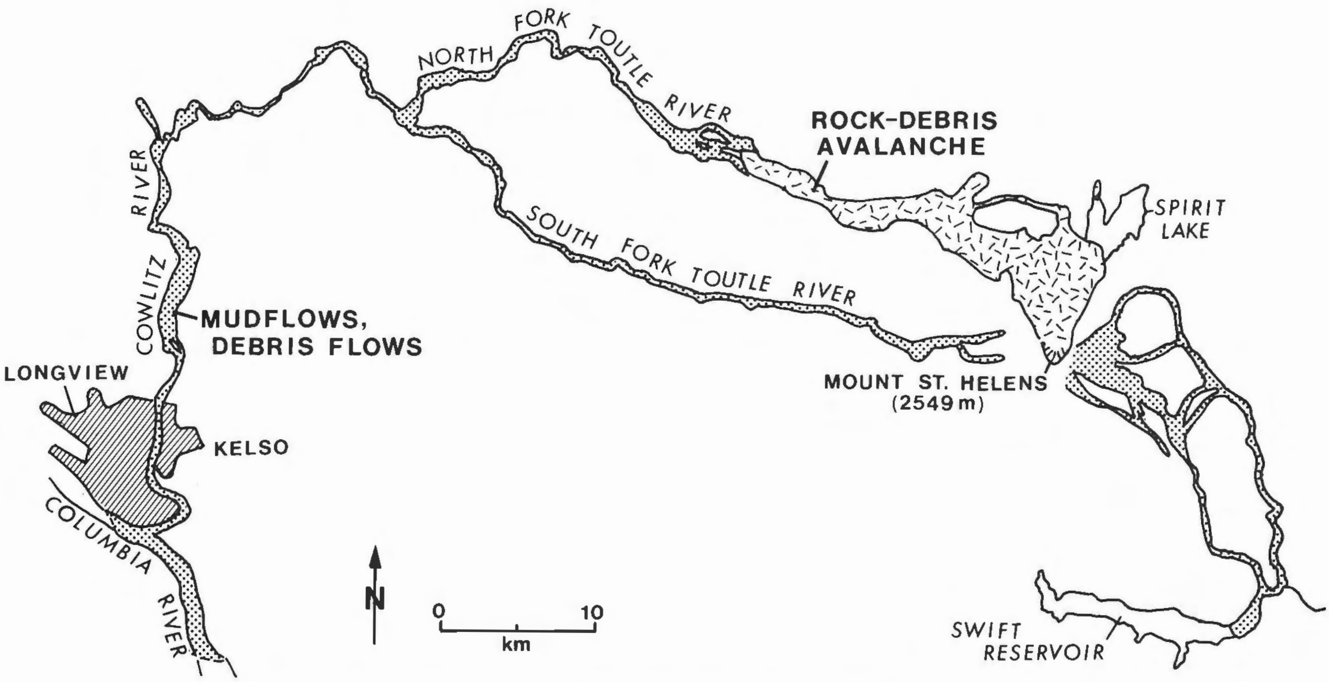


Figure 24: Map showing the rock-debris avalanche, mudflows and debris flows resulting from the eruption of Mount St. Helens, Washington, on 18 May 1980. (Adapted from Janda et al., 1981, Fig. 271).

during eruptions (Crandell, 1971). Similar processes have occurred on all Cascade volcanoes in the western United States, most recently in May 1980 during the climactic eruption of Mount St. Helens.

Some of the most spectacular debris movements associated with the rapid melting of snow and ice result from eruptions beneath glaciers. Many of Iceland's active volcanoes, for example, are buried by glaciers. During eruptions of these volcanoes, large amounts of water accumulate in subglacial chambers. This water may be released suddenly to produce tremendous floods, or jökulhlaups, that transport large amounts of sediment (Wadell, 1935). Often, true debris flows are generated by these floods, as for example during the 1934 eruption of Grimsvötn and the 1948 eruption of Katla. The Grimsvötn flows, which formed over a period of 3 days, were up to 8 km wide and had a total volume of 8.3×10^9 m^3 (Williams and McBirney, 1979).

Explosion-induced avalanches of volcanic debris into streams

Avalanches of volcanic material can cause debris flows by temporarily damming streams and rivers. The build-up of water behind a debris dam may eventually cause it to fail. An event of this kind occurred during an eruption of Asama volcano in Japan in 1783 when a pyroclastic flow dammed a nearby river. Within an hour the river overtopped the dam and rapidly cut down through it, creating a flow of hot debris that travelled downvalley over 80 km and killed more than 1300 people (Aramaki, 1956).

Movement of pyroclastic flows along river valleys

As mobile pyroclastic flows move down or across valleys, they may transform into debris flows due to cooling and addition of river water. For example, during the 1929 eruption of Santa Maria volcano in Guatemala, pyroclastic flows travelled down the valleys of the Río Tamblor and Río Concepcion and changed into boiling-hot debris flows, one of which overflowed a channel more than 90 m deep and 75 m wide (Sapper and Termer, 1930).

Failure of the flank or summit of a volcano

Volcanic explosions may cause portions of the flank or summit of a volcano to slide away and transform into debris avalanches and flows. This can occur if part of the cone is undermined, if steep brecciated domes are built above the walls of a crater, or if large quantities of unstable tephra accumulate on a steep slope. Also, prior to an eruption, a portion of the cone may collapse due to the development of excess pressures in the magma chamber below. Lastly, flank and summit failures may be triggered by earthquakes that accompany an eruption. Large flank and summit failures are common occurrences during volcanic eruptions; the following are typical examples.

In July 1888 after a week of earthquakes, a phreatic explosion blasted away part of the flank of Bandai-san volcano in Japan. As a consequence, a large part of the upper cone began to move downslope breaking up and gathering speed as it went. In all, about 1200×10^6 m^3 of volcanic debris avalanched off the volcano and spread out over an area of

70 km^2. The northern flank of the mountain was replaced by a horseshoe-shaped amphitheatre; thousands of conical, domal, and pyramidal mounds as much as tens of metres across stood out from the avalanche surface. The debris consisted of angular blocks as much as 10 m in diameter in a matrix of finer debris. The avalanche entered stream valleys where it mixed with water to form mobile debris flows that destroyed farmland and killed more than 400 people (Sekiya and Kikuchi, 1889).

An explosive eruption of Tokachi-dake volcano in Japan in 1962 caused about 4×10^6 m^3 of hydrothermally altered rock to slide from one side of the cone. The slide mass transformed into a debris flow as it moved downward (Murai, 1963).

The eruption of Mount St. Helens on 18 May 1980 also produced a large landslide remarkably similar to that at Bandai-san nearly a century earlier. Following two months of intense slope deformation associated with magmatic intrusion, seismicity, and gravitational creep, an enormous mass of volcanic material with an estimated volume of 2300×10^6 m^3 became detached on the north slope of Mount St. Helens during a magnitude 5+ earthquake and accelerated towards the base of the mountain (Voight et al., 1981, 1983). After 26 seconds, the first of a retrogressive series of blocks had moved 700 m and was travelling at about 50 m/s; a second block had moved 100 m. These slide masses rapidly disintegrated and evolved into an enormous pulsing hot avalanche. Within 10 minutes, avalanche lobes had travelled as far as 8 km northward and 23 km westward, and an area of about 60 km^2 of the North Fork Toutle River valley system was choked with 2800×10^6 m^3 of hummocky surfaced, poorly sorted debris to an average depth of 45 m. One avalanche lobe entered a lake and generated a wave that surged up to 260 m higher on adjacent slopes.

When this avalanche came to rest on the morning of May 18, it still was very unstable because much of the debris was saturated with water derived from snow and ice of the summit cone. Flows probably soon developed on various parts of the hummocky landslide surface, but it was not until late on the afternoon of May 18 that a large integrated mudflow moved off the avalanche and down the North Fork Toutle Valley. This flow destroyed trees, homes, and bridges, and wiped out a mid-valley logging camp. Peak velocities probably ranged from about 12 m/s near the source to less than 1.5 m/s in downstream reaches (Janda et al., 1981). Thinning with distance, the mudflow travelled more than 120 km into the Cowlitz and Columbia valleys. This and other flows ultimately deposited more than 50×10^6 m^3 of sediment along the lower Cowlitz and Columbia rivers and raised the bed of the former by about 5 m (Lombard et al., 1981).

Heavy rains on newly fallen pyroclastic deposits

Newly erupted pyroclastic deposits lack a protective cloak of vegetation and are especially susceptible to saturation and erosion during periods of heavy rainfall. Sliding and flowage of masses commonly cause debris and mud flows. These "rain lahars" were almost a daily occurrence during the monsoon season following an eruption of the Indonesian volcano Kelut in 1919 (Kemmerling, 1921). A few weeks after the eruption of La Soufrière in 1902, heavy rains mobilized still-hot pyroclastic deposits, and a series of boiling-hot flows swept down valleys on the flanks of the volcano (Anderson and Flett, 1903). Finally during the 1963-1964 eruptions of Irazú in Costa Rica, flash floods scoured poorly consolidated pyroclastic deposits, thus transforming them into debris flows. One of these flows caused great damage and some loss of life in the city of Cartago at the south base of the volcano (Murata et al., 1966; Waldron, 1967).

Mass movements unrelated to volcanic eruptions

Mass movements on volcanoes occur not only during eruptions, but also during periods of quiescence and after the volcano has become extinct. The presence of steep slopes, abundant moisture, and a source of debris ensure that mass movements will occur, although perhaps not with the same frequency as during eruptions. These conditions also are found in many nonvolcanic mountain areas, and in this sense, noneruptive volcanic mass movements resemble those of other areas. However, because many volcanoes consist almost entirely of extremely weak, poorly lithified materials that are especially prone to failure, the likelihood of mass movements in such areas is greater than elsewhere.

Most noneruptive mass movements on Quaternary volcanoes are classified as debris flows because they exhibit a dominantly flowing motion and are derived from heterogeneous, poorly consolidated volcanic materials (commonly pyroclastic debris). However, some large flows are initiated by falls or slides, and involve little water; these are more appropriately labelled rock or debris avalanches.

As mentioned previously, noneruptive volcanic mass movements may be subdivided into those resulting from the failure along well defined rupture surfaces and those resulting from erosion by running water. Some of the triggering mechanisms identified in the section dealing with eruptive mass movements also apply here, including heavy rainfall, earthquakes, and melting of snow and ice. In addition, normal fluvial and glacial erosion operating over a long period of time may gradually reduce the stability of volcanic slopes and eventually lead to their failure. As in nonvolcanic areas, human activities may play a critical role in destabilizing volcanic slopes, for example through road construction, erection of buildings, and deforestation.

The following examples illustrate the character and variety of noneruptive volcanic mass movements and highlight the hazards they pose.

Mass movements resulting from failure along well defined rupture surfaces

Coherent masses of pyroclastic deposits, lava flows, and related rocks may detach from underlying materials along well defined surfaces of rupture. On steep slopes the failed material may become airborne, resulting in a rockfall or rock avalanche. On lesser slopes, the failed mass slides off the rupture surface and rapidly disintegrates into a chaotic mélange that flows at high speeds downslope. These mass movements range from individual blocks to large rock avalanches and debris flows of up to 10^7–10^8 m^3 volume.

In the Canadian Cordillera more than half of the large ($>10^6$ m^3) historical landslides have occurred on Quaternary volcanoes. These failures include the 1855-1856 Rubble Creek slide near Mount Garibaldi (Moore and Mathews, 1978), the 1963 Dusty Creek landslide on Mount Cayley (Clague and Souther, 1982), and the 1975 Devastation Glacier slide near Meager Mountain (Mokievsky-Zubok, 1977), all of which are in British Columbia. Mount Garibaldi, Mount Cayley, and Meager Mountain are dissected stratovolcanoes consisting of pyroclastic materials, lavas, and hypabyssal intrusive rocks. These volcanoes were constructed largely during the Pleistocene, although the oldest activity probably is Pliocene (Souther, 1980). Mount Garibaldi was last active at the end of the Pleistocene, and the Meager Mountain complex last erupted less than 2500 years ago (Nasmith et al., 1967); Mount Cayley apparently has not been active during the Holocene.

The Rubble Creek slide resulted from a failure of "The Barrier", a 500 m high face marking the front of a lava flow that cooled in contact with a late Pleistocene glacier (Fig. 25). During the winter of 1855-1856, a major section of The Barrier failed along a near-vertical composite fracture zone causing a stream of rock rubble, 25×10^6 m^3 in volume, to flow down the valley of Rubble Creek and across the debris fan at the mouth of the valley. Traces of rubble at the edge of the debris track and superelevation at its bends suggest that the front of the mobile mass flowed down Rubble Creek valley at high velocity, locally in excess of 20 m/s, and lapped up to 80 m on the valley walls. The main debris stream, carrying slabs of volcanic rock several metres in diameter, spread over the northern half of the Rubble Creek fan and blocked Cheakamus River. Subsequent bouldery flows, composed of reworked slide material and rockfall talus from the toe of The Barrier, covered the southern sector of the fan. Debris floods, launched when the debris dam was overtopped by the impounded Cheakamus River, buried tracts of forest on the floor of Cheakamus Valley below Rubble Creek. The Barrier probably failed in a similar fashion several times prior to 1855. This is indicated by the unusually large size of the Rubble Creek debris fan and by the presence of pre-1400 A.D. landslide deposits beneath the 1855 debris lobe. The trigger mechanisms for both the 1855 landslide and earlier failures are unknown.

Figure 25: View of "The Barrier", source of the 1855-1856 Rubble Creek slide in British Columbia. This near-vertical cliff is the ice-contact front of a late Pleistocene lava flow. (GSC 204165-I)

The Dusty Creek landslide occurred in 1963 when a large (5×10^6 m^3) block of poorly consolidated tuff-breccia and columnar-jointed dacite became detached from the subvolcanic basement on the west flank of Mount Cayley and slid into the valley of Dusty Creek, a small tributary of Turbid Creek (Fig. 26). As the detached block accelerated, it quickly fragmented into an aggregate consisting of angular clasts up to several metres across, partially supported by a matrix of fine comminuted rock material. This debris moved down Dusty Creek as a wedge-shaped mass up to 70 m thick, banking up on turns, and attaining a maximum velocity of 15-20 m/s. The debris mass thinned as it spread across the wider and flatter valley of Turbid Creek, and came to rest as an irregular blanket with a maximum thickness of 65 m along a 1 km length of this valley. As a result of the landslide, Turbid and Dusty creeks were blocked, and lakes formed behind the debris. These debris dams were soon overtopped and breached, causing floods and debris flows to sweep down Turbid Creek valley far beyond the terminus of the landslide. Because this failure occurred in a remote area of the southern Coast Mountains of British Columbia, it was not observed directly. However, tree-ring analysis of slide-damaged trees indicates that the event probably occurred in July 1963. The cause of the landslide is uncertain, but it appears that the stability of the slope at the head of Dusty Creek gradually deteriorated over a long period of time until a relatively minor event, such as a small earthquake or storm, triggered the failure.

On 22 July 1975, an estimated 2.5×10^6 m^3 of ice and 26×10^6 m^3 of weakly consolidated pyroclastic rocks broke away from the northwest slope of Pylon Peak near Meager Mountain, slid across the stagnant toe of Devastation Glacier, and flowed at high velocity over 6 km down Devastation Creek to its mouth (Fig. 27). As the debris descended Devastation Creek, it banked up the valley walls at each bend

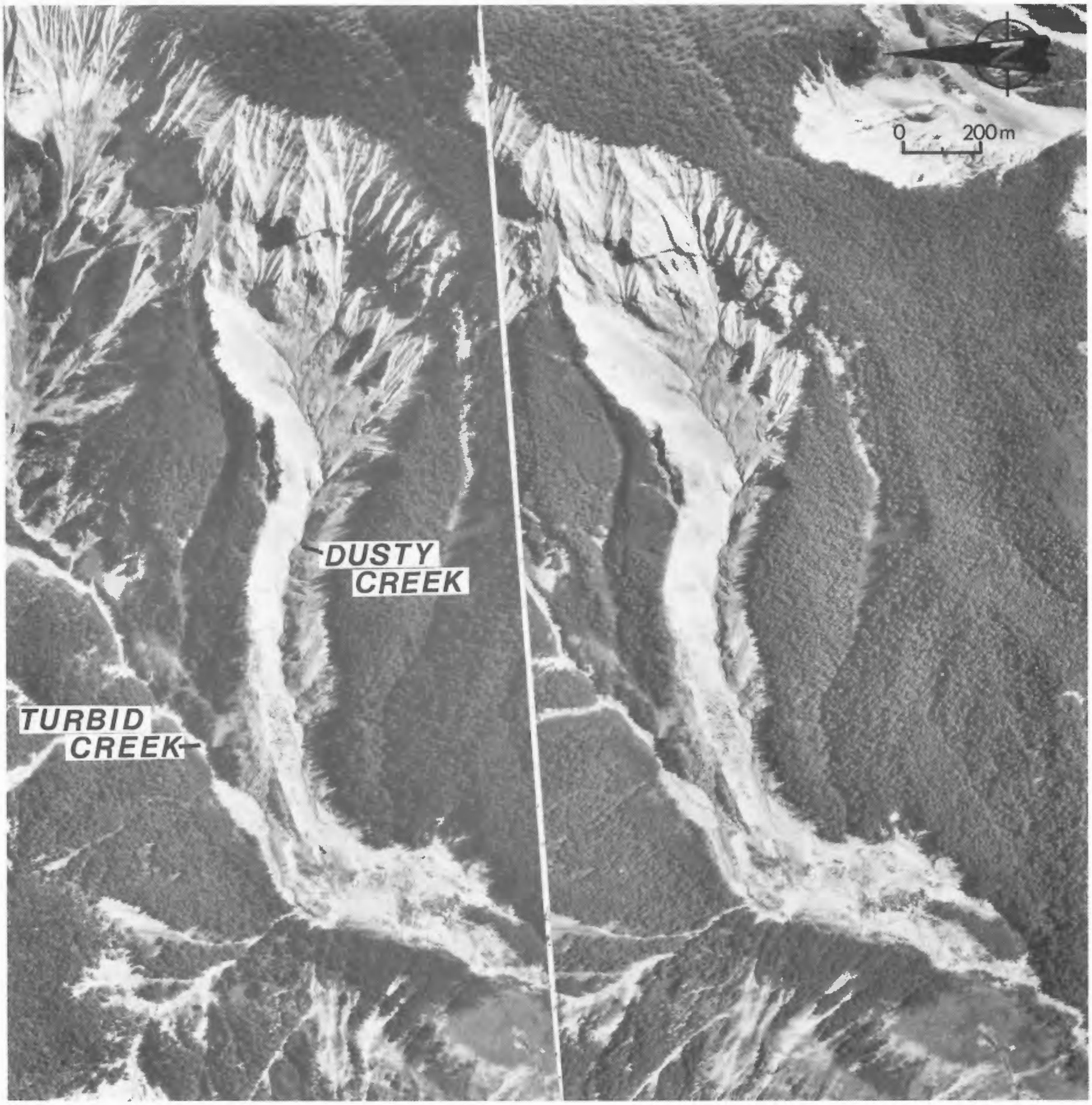


Figure 26: The 1963 Dusty Creek slide, British Columbia. This slide resulted from the failure of Pleistocene lava flows and pyroclastic rocks of the Mount Cayley volcanic complex. Photos taken in August 1973. Stereoscopic pair, Province of British Columbia photos BC7520-258, BC7521-025.

and triggered minor slides that joined the main flow mass. The debris came to rest in the valley of Meager Creek where it impounded a small lake. Here, the flow claimed the lives of four people who were awaiting the arrival of a helicopter. The cause of the failure probably was the discharge of large amounts of meltwater from the base of Devastation Glacier into underlying weak volcanic materials (Mokievsky-Zubok, 1977).

Occasionally, large masses of volcanic rock slide into lakes and trigger secondary debris flows. On 15 January 1967, Lake Steinsholtslon in south-central Iceland was emptied when struck by a rockslide from the flank of an adjacent ice-capped Pleistocene stratovolcano (Kjartansson, 1967). The waters of the lake became mixed with fragmented glacier ice and slide debris to produce an ice-debris flood with a total volume of $1.5–2.5 \times 10^6$ m^3. This flood swept 35 km to the sea and transported blocks that were many metres in diameter for distances of over 5 km.

Rock avalanches from Little Tahoma Peak, Washington, in 1963 illustrate another type of noneruptive volcanic mass movement — one involving a free fall of volcanic rock. On 14 December 1963, and probably on several subsequent occasions during the same month, very large masses of lava rock and interbedded breccia fell as much as 520 m from the face of Little Tahoma Peak on the east side of Mount Rainier onto Emmons Glacier (Fig. 28; Crandell and Fahnestock, 1965). As the rock masses struck the glacier, they shattered and formed rock avalanches that rushed as much as 7 km downvalley. In this distance the rock debris descended up to 1890 m in altitude. Minor lithological differences of the rubbly debris lobes and their crosscutting relationships indicate that there were at least seven distinct avalanches separated by intervals of minutes, hours, or possibly days. The total volume of transported material was about 11×10^6 m^3. During movement, some of the avalanches caromed from

Figure 27: The valley of Devastation Creek, British Columbia, shortly after the Devastation Glacier slide of 22 July 1975. Note the sharp trimlines which delineate the upper limit of the flowing mass of volcanic debris in the valley. (GSC 204165-J)

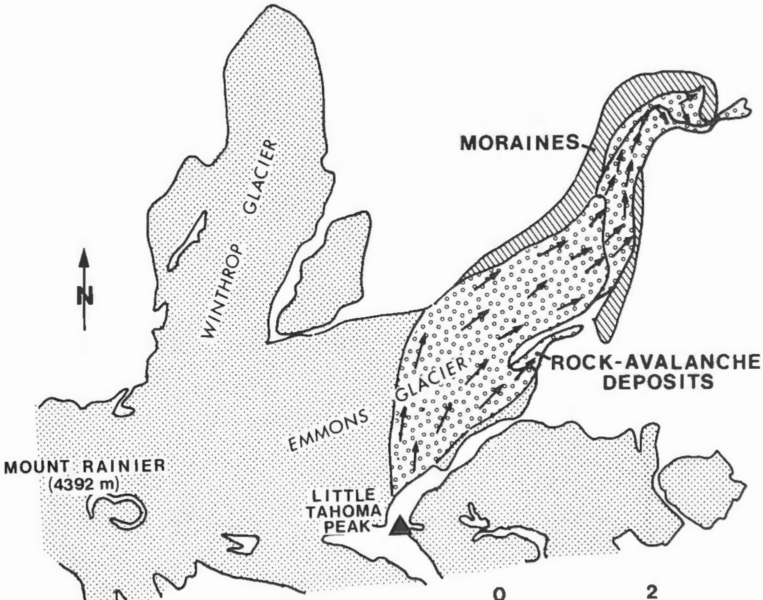


Figure 28: Map of the northeast side of Mount Rainier, Washington, showing the path of the rock avalanches from Little Tahoma Peak in December 1963; arrows indicate the inferred direction of movement. (Adapted from Crandell and Fahnestock, 1965, Fig. 1).

one side of the valley to the other. Calculations based on the height to which the avalanches rose on the valley walls suggest that velocities in excess of 35 m/s were attained by the streaming debris. The unusually long distance that some of the avalanches travelled has been attributed to a cushion of trapped and compressed air at their base, which buoyed them up and reduced friction (Crandell and Fahnestock, 1965). The initial rockfall may have been triggered by a small steam explosion near the base of Little Tahoma Peak.

Earthquakes trigger many mass movements in Quaternary volcanic rocks, just as they cause landslides in other earth materials. The role played by earthquakes is illustrated by an example from Guatemala. The Guatemala earthquake of 4 February 1976 (M = 7.5), generated more than 10 000 landslides throughout an area of approximately 16 000 km^2 (Fig. 29; Harp et al., 1981). Over 90% of these landslides occurred on steep erosional slopes developed in Pleistocene pumice deposits which crop out over about 20% of the area affected. These landslides caused hundreds of fatalities as well as extensive property damage, and severely disrupted road and rail traffic in Guatemala. The main types of earthquake-triggered landslides were rockfalls and debris slides of less than 15 000 m^3 volume; in addition to these relatively small landslides, a few large (>100 000 m^3) block slides, slumps, and "rock avalanches" occurred in the pumice deposits.

One of the most dramatic earthquake-triggered mass movements in volcanic rocks occurred in 1840 on Mount Ararat (5165 m) in eastern Turkey. The composite cone of Mount Ararat is located in a region of extremely high seismicity. It also is capped by glaciers whose termini generally are located above elevations of 3500 m. On the northeast side of the volcano, however, the debris-covered Cehennem-Dere Glacier flows down a deep gorge to an elevation of 2400 m. On 20 June 1840, a strong earthquake released a major rock-ice avalanche from the high cliffs above this glacier. As a result, the gorge was blocked and a lake began to form upstream. About 72 hours later, the impounded

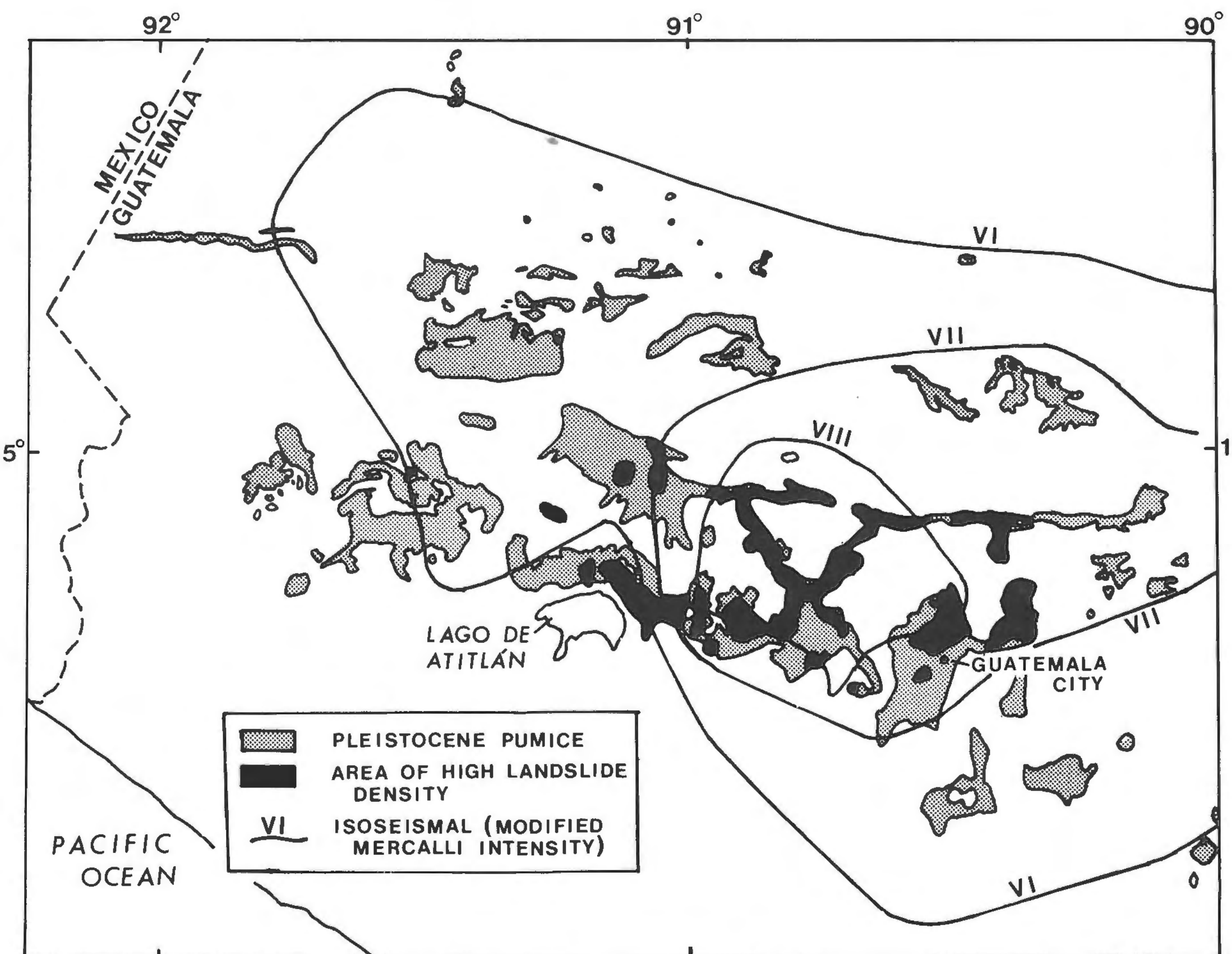


Figure 29: Map of southwestern Guatemala showing areas of high density of landslides triggered by the earthquake of 4 February 1976 (Harp et al., 1981); also shown are isoseismals (Espinosa et al., 1976) and the distribution of Pleistocene pumice deposits (Bonis et al., 1970; Koch and McLean, 1975).

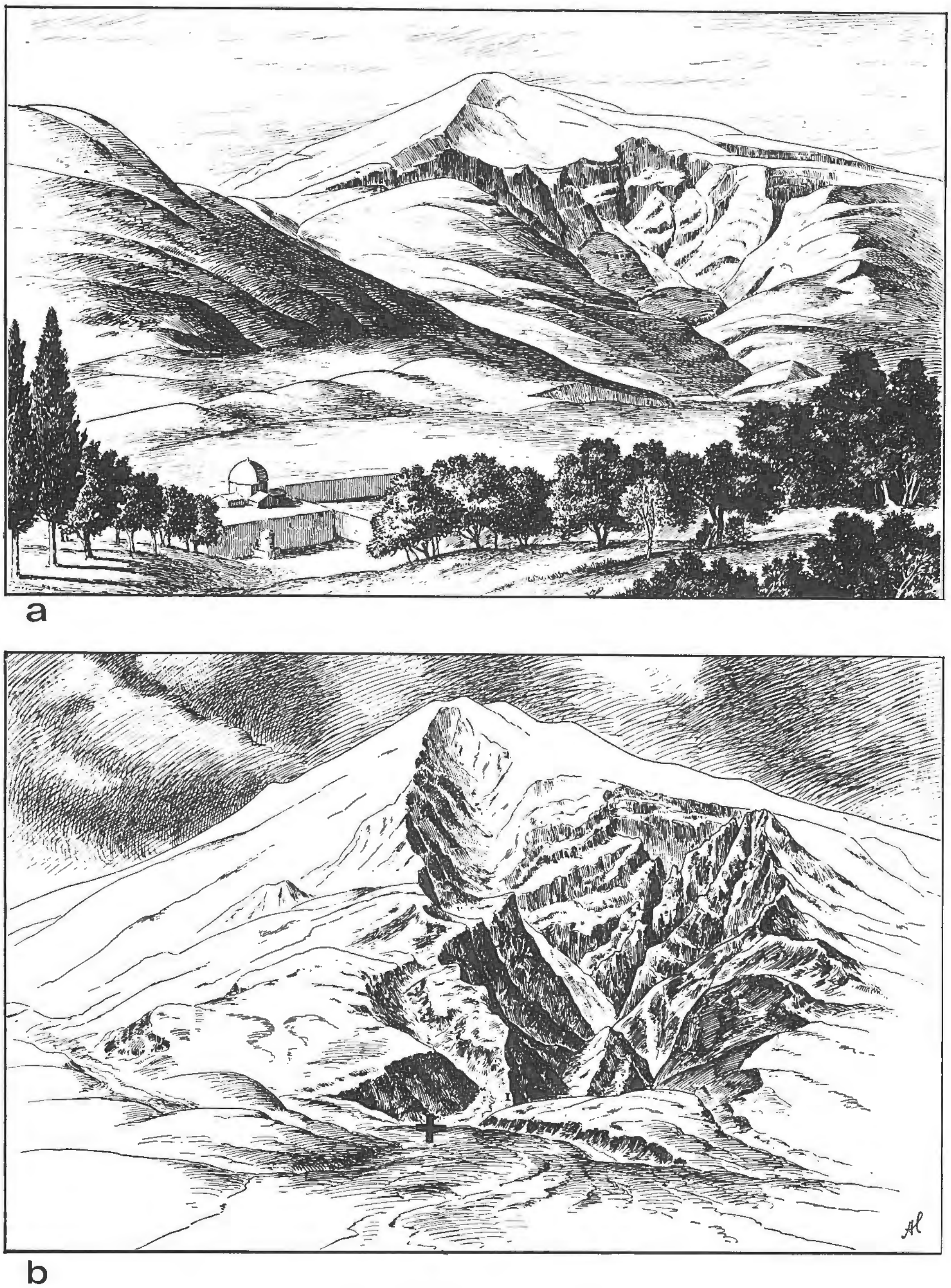


Figure 30: Sketches of the northeast face of Mount Ararat, Turkey, a) before and b) after the catastrophic debris flow of 1840 which erased the village of Achury (location of village indicated by a cross; from Ebeling, 1899.

waters burst through the debris dam, and an ice-debris flow moved rapidly downvalley, erasing a village and monastery near the base of the mountain (Abich, 1847). Sketches of the northeast face of Mount Ararat before and after the catastrophe of 1840 are shown in Figure 30 (Ebeling, 1899).

Perhaps the most common triggering mechanism of noneruptive mass movements in volcanic rocks is heavy rainfall and snowmelt. Intense rainstorms and heavy runoff from snowpacks or glaciers often cause excessive infiltration of water into slopes underlain by weak volcanic materials. Water-saturated pyroclastic deposits and fractured lava flows are especially susceptible to failure by slumping and debris avalanching. The failed masses, often relatively small in size (10^5–10^6 m^3), may transform directly into mobile debris flows that sweep down valleys and ravines on the flanks of a volcano. Alternatively, slumps and debris avalanches may block a gorge and temporarily stem the flow of water in it. When these dams fail due to overtopping by water or to the pressure of the impounded water itself, debris flows often are produced. These types of failures are extremely common on most stratovolcanoes in humid areas and are capable of great destruction. In Japan, for example, widespread property damage and much loss of life have resulted from debris flows triggered by severe rainstorms, and such debris flows have been particularly common on slopes underlain by Quaternary volcanic materials (Ikeya, 1976).

Mass movements caused by erosion

Frequently, debris flows are generated by the rapid erosion of volcanic materials along pre-existing gullies and ravines (Okuda et al., 1980). This commonly occurs during periods of extremely heavy rainfall or at times when bodies of water within or at the margins of glaciers drain catastrophically. At such times torrential runoff may deeply scour the floors or walls of a channel, entraining large quantities of loose volcanic sediment. This mobilized sediment may then be transported down the gully as a debris flood or debris flow. Mass movements of this sort commonly occur at the same time as rain-induced slumps, debris avalanches, and debris flows that result from failure along well defined rupture surfaces.

Protection against volcanic mass movements

Many mass movements caused by volcanic eruptions are so large that protective structures are of little value. In such instances complete protection is provided only by avoiding potentially hazardous areas. In most cases, however, rigid zoning is not practical for economic or social reasons. Protection of people and property in such areas is dependent largely on successful forecasts of eruptions and related mass movements. If the time and place of an eruption can be predicted, people can be evacuated from the threatened area, and some property may be moved to safety. A tremendous amount has been written on this subject, and little can be added here (for further information, the reader should consult general texts on volcanology, for example Macdonald, 1972; Williams and McBirney, 1979; and Decker and Decker, 1981). Suffice it to say that two main techniques have been successfully used in predicting eruptions: (1) monitoring of earthquake activity related to the movement of magma beneath or in a volcano with seismographs; and (2) detection of slight changes in the surface shape of a volcano using tiltmeters, geodimeters, and conventional surveying techniques. In addition, changes in the Earth's gravitational, magnetic, and electric fields near volcanoes, minor changes in heat flux, and changes in the temperature or chemistry of fumaroles may signal impending eruptions, although these techniques have not been used with much success to date.

Studies of hazards associated with specific volcanoes complement research in eruption forecasting. A researcher who knows the past history of a volcano may be able to identify the types of hazards and zones of potential danger in the event of a future eruption. This information is best depicted on maps that delineate hazard areas such as potential pyroclastic- or debris-flow zones. Such maps have been prepared for several volcanoes in the Cascades of western North America (Crandell, 1973; Crandell and Mullineaux, 1978; Hyde and Crandell, 1978) the Hawaiian Islands (Mullineaux and Peterson, 1974; Crandell, 1975), and for several volcanoes outside the United States, including Kelut (Neumann van Padang, 1960), Merapi (Pardyanto et al., 1978), Mount Vesuvius (Pinna and Scandone, *in* Barberi and Gasparini, 1976), Tenerife (Booth, 1977), and Cotopaxi (Miller et al., 1978). Volcanic-risk maps like these, along with nontechnical descriptions of the hazards, point out areas

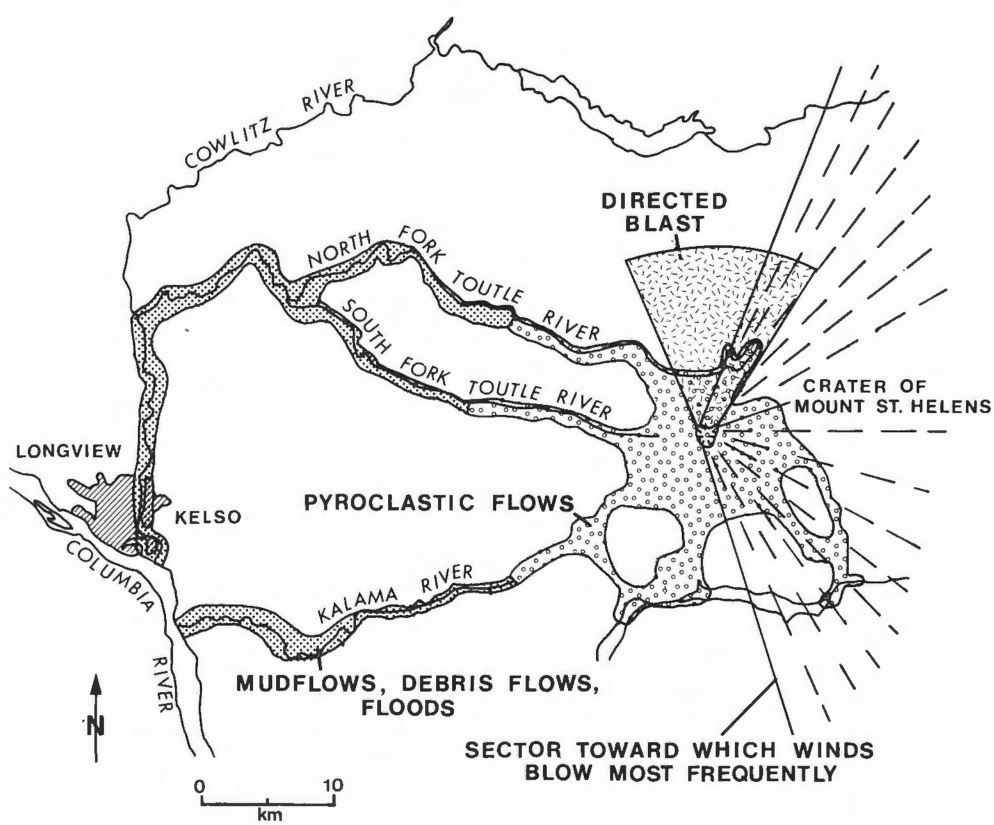


Figure 31: Map of the region near Mount St. Helens, Washington, showing zones of potential hazard from pyroclastic flows, debris flows, mudflows, floods, and directed blasts; also shown is the area into which winds are most likely to blow tephra during an eruption. The map was constructed by staff of the United States Geological Survey shortly after the cataclysmic eruption of Mount St. Helens on 18 May 1980 (Miller et al., 1981, Fig. 455).

of potential danger during future eruptions, and also can be used for long-term, land-use planning for regions around volcanoes. They show, for example, the likely extent of debris flows in valleys surrounding a volcano (Fig. 31). They may also show areas of potential nuée ardente and blast damage, although uncertainties as to the direction and strength of future explosive eruptions limit the value of such predictions.

Even if a volcanic eruption can be forecast, the timing of mass movements resulting from that eruption may be very difficult to predict. Mass movements may occur at the beginning of an eruption, at any time during the eruption, or for some time after the eruption has ended. Once set off, they may move so rapidly that it is not possible to provide adequate warning to people in their paths.

A classic example of the recognition of a hazard and the application of remedial measures is the drainage of the crater lake of Kelut in Indonesia. During several previous eruptions of Kelut, the crater lake was ejected, causing destructive debris flows in valleys surrounding the volcano. After the disastrous eruption of 1919 when over 5000 people were killed by debris flows, Dutch engineers constructed a series of tunnels through the flank of the volcano to drain most of the lake so that future eruptions would do less damage (Fig. 32; Zen and Hadikusumo, 1965; Macdonald, 1972). The volume of the lake was reduced to about 2×10^6 m^3, and when another violent eruption took place in 1951, no large debris flows were generated. However, the 1951 eruption destroyed the intakes of the tunnel system, and the crater was deepened about 70 m; only the main shaft and the lowest tunnel could be rebuilt (Zen and Hadikusumo, 1965). Thus when the lake filled to the level of the lowest tunnel, it contained about 23×10^6 m^3 of water. It was apparent that if this volume of water remained, devastating debris flows would likely occur during the next eruption. Consequently, another tunnel was driven 20 m lower, but was stopped short of the lake. It was hoped that seepage through permeable materials of the crater would drain the lake to the level of this tunnel, but this did not happen, and when Kelut erupted again in 1966, debris flows killed almost 300 people. A new, low-level tunnel was completed in 1966, and the lake was again drained to a low level.

Most eruptions in inhabited regions must be carefully monitored. The growth of a dome or the outpouring of lava in the crater of a snow-covered volcano commonly produces abundant meltwater which may initiate debris flows. The heads of valleys on volcanoes should be kept under careful observation during eruptions for signs of impending debris flows. On Merapi volcano in Indonesia, thermoelectric sensors were installed at one time in the valley heads to warn of the passage of hot debris flows and pyroclastic flows; however, these devices were not successful. The slopes of volcanoes and nearby hillsides should be watched during and after an eruption for the accumulation of a cover of loose tephra that might become mobilized by heavy rains.

Some protection may be provided against small debris and mud flows by the construction of diversion dams, retention basins, and elevated artificial structures that serve as refuges. For example, mounds have been built near some Indonesian villages; people may flee onto them in the event of a debris flow. Of course, to be effective, protective structures must be large enough to deflect or impound expected flows. Large reservoirs may contain debris flows of moderate volume very successfully. Crandell (1971) has pointed out, for example, that an existing reservoir near the base of Mount Rainier could contain most or all of a debris flow of the size of the Electron flow (150×10^6 m^3) which took place about 500 years ago (Fig. 22).

Finally, control works and protective measures that have been so successfully applied in nonvolcanic areas may be used on extinct volcanoes to protect against rain-induced debris flows. However, foundation sills and abutments of erosion check dams must be emplaced deeply into the substratum in volcanic terrain to be effective (Ikeya, 1976). Also, protective dams must be large and should include selective discharge sections so as to not fill up between debris flows.

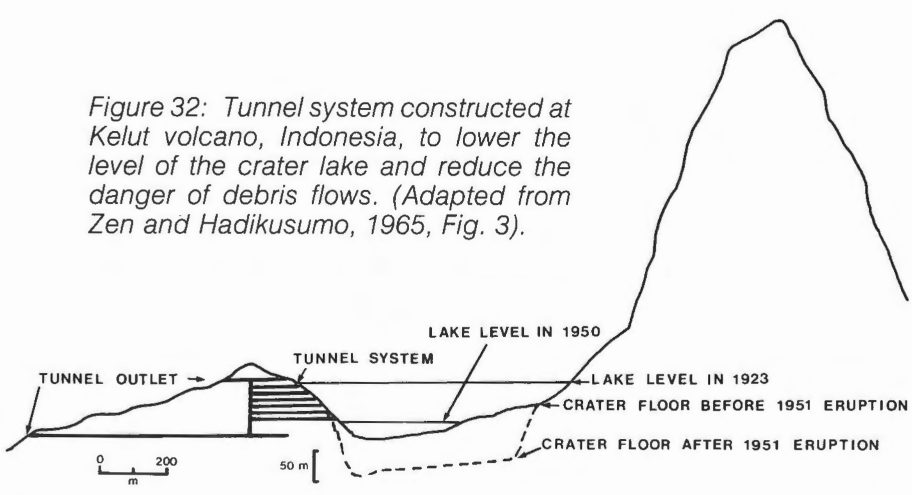


Figure 32: Tunnel system constructed at Kelut volcano, Indonesia, to lower the level of the crater lake and reduce the danger of debris flows. (Adapted from Zen and Hadikusumo, 1965, Fig. 3).

GLACIER - RELATED MASS MOVEMENTS

Many mass movements in high mountain areas are the direct or indirect result of processes related to the unstable margins of glaciers and to the mobilization of loose, poorly vegetated debris lodged immediately below the glacier termini. The mechanisms of these mass movements are strongly influenced by the shape of the ice masses, the geometry of the glacier bed, and the volume of morainal debris deposited at the front.

For purposes of the discussion which follows glaciers in mountainous regions can be broadly grouped into cirque glaciers, valley glaciers, and outlet glaciers of mountain ice fields. A cirque glacier is wholly or largely confined to a cirque. Glaciers that flow down a valley from a cirque or from an ice sheet are termed valley and outlet glaciers, respectively. Mountain ice sheets overlie or originate in two or more mountain masses, burying most interfluves, but leaving high crests and peaks exposed; outflow of ice takes place in two or more directions.

A glacier is an active and very sensitive body. We can view it as an open system, with input of snow at the surface in the zone of accumulation, gradual conversion of this snow into ice, downslope flow, and output of water, water vapour, and ice in the zone of depletion. In the case of a small glacier, the local topography, orientation, exposure, relation to prevailing winds, size, shape, and location of the accumulation region all play a part in determining the glacier's response to climatic variations (Sharp, 1960). Because these factors may vary markedly over short distances, it is possible for adjacent glaciers to exhibit different behaviour in response to climatic change; one glacier may recede while its neighbour advances. The size of a glacier also is an important determinant of its response to a change in climate. In general, a small glacier reacts quickly to temperature and precipitation changes; its terminus generally advances or retreats within a year or two of such a change. In contrast, the termini of large glaciers may not respond to climatic change for many years or even decades. Some valley and outlet glaciers are also subject to sudden advances, or surges, that seemingly are unrelated to climate. Glacier surges usually follow periods of downwasting and retreat extending over many decades. During such an inactive phase, the downstream part of a glacier thins and may even become stagnant, reflecting a negative mass balance. Then suddenly a surge begins. The terminal zone of the glacier thickens and becomes bulbous, while upstream portions thin. The snout of the glacier begins to move forward at rates many times faster than normal (in some instances, more than 150 m/day). Some large glaciers surge more than 10 km before a new phase of stagnation and retreat begins (Flint, 1971).

Glaciers in mountain areas have fluctuated markedly during historical times due to global and regional climatic changes. Today, most cirque and valley glaciers are smaller than they have been during the last few centuries, reflecting the somewhat warmer climate of the present. Many glaciers attained their maximum Holocene extensions between about A.D. 1550 and 1850, during a cool wet period referred to as the 'Little Ice Age' (Matthes, 1939; Porter and Denton, 1967; LeRoy Ladurie, 1971). The Little Ice Age is part of the longer cool period that began as early as 5000 years ago and is called 'Neoglaciation'. Neoglaciation followed an interval of maximum Holocene temperatures ('Hypsithermal') when most alpine glaciers retreated to highest postglacial positions, and when some cirque glaciers even disappeared. Because of the advances during the Little Ice Age most mountain glaciers are bordered by a zone that only recently was covered by ice and is composed of a great variety of unconsolidated bouldery sediments, chiefly till, outwash, and ice-contact debris.

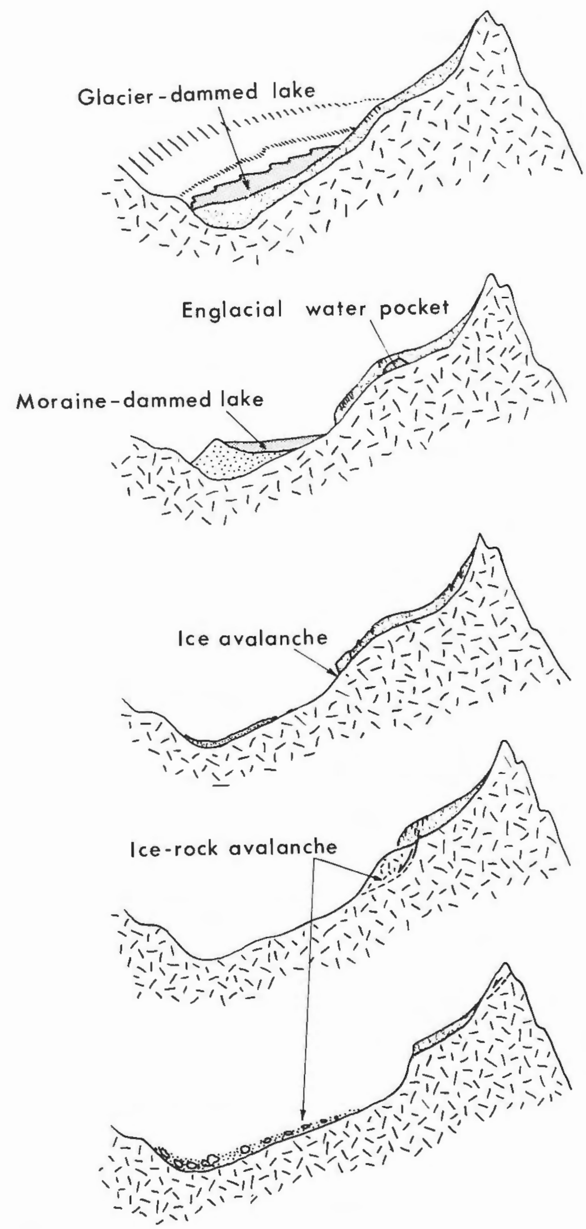


Figure 33: The principal situations in which glacier-related mass movements occur (see text for explanation).

Glacier-related mass movements fall into five major categories (Fig. 33): (1) debris flows caused by bursts of ice-dammed and ice-marginal lakes; (2) debris flows resulting from bursts of moraine-dammed lakes; (3) debris flows resulting from bursts of subglacial and englacial water pockets; (4) failures of glacier ice (ice avalanches); and (5) failures of bedrock and superincumbent glacier ice (ice-rock avalanches).

Debris flows from ice-dammed and ice-marginal lakes

Ice dammed and ice-marginal lakes are common in many glaciated regions. Rapid draining of these lakes generally causes floods, known as 'jökulhlaups' in Iceland where they are common (Thorarinsson, 1939). These floods occur when the pressure exerted by the lake waters exceeds the strength of the ice dam (Bjornsson, 1975). Drainage may occur either at the base of the glacier or through one or more englacial channels. Floods of this type have been reported from almost all temperate glacierized regions of the world (e.g. Fig. 34). They are known from the Himalayas (Mason, 1929, 1935), the Andes (King, 1934; Helbing, 1940; Liss, 1970), and western North America (Kerr, 1934; Stone, 1963; Mathews, 1965; Post and Mayo, 1976; Young, 1980; Clague and Rampton, 1982). Such floods may have considerable impact on the valleys below the ice dam. However, the destructive force of sudden bursts of ice-dammed lakes is particularly

Figure 34: Lake Tulsequah in the northern Coast Mountains of British Columbia shortly after it emptied during a jökulhlaup. The lake is dammed by a lobe of Tulsequah Glacier (foreground). Photo by Austin Post.

severe if the valley below the bursting dam is narrow, the torrent channel steep, and the bed boulder-strewn. In such a case the spurt of water changes quickly into a blocky debris flow. Historical debris flows originating by the burst of ice dams and discharge of large quantities of water into steep mountain torrents have caused considerable damage in Norway (Liestol, 1956), central Asia (Nijasov, 1975), the Caucasus (Hoinkes, 1972), and in the Alps (*see* case histories in the Appendix under Val de Bagnes, Ötz Valley Ice Floods, Martell Ice Lake, and Mattmark).

The case histories from the Alps illustrate dramatically that remedial measures can be directed at the lakes themselves only as long as the accumulated water volume is still relatively small: trenching across the ice and tunneling through rock or ice may allow the slow controlled drainage of the rising lake (Richter, 1889b; Röthlisberger, 1979, 1981).

Figure 35: Nostetuko Lake and the breached late Neoglacial end moraine of Cumberland Glacier in the southern Coast Mountains of British Columbia. The lake drained catastrophically in July 1983 when a wave overtopped the moraine. Escaping waters rapidly cut down through the loose morainal debris, producing an alluvión. Photo courtesy of British Columbia Hydro and Power Authority.

Under circumstances where draining of an ice-dammed lake is deemed impractical flood basins and debris retention dams may be constructed below the ice dam. Where technical work is not feasible monitoring of both the lake and the ice dam provide for adequate warnings of an impending burst and thus permit evacuation of endangered communities in the valley below.

Debris flows from moraine-dammed lakes

Many cirque and valley glaciers are fringed by bulky terminal moraines. During retreat of the glaciers the moraines may define the frontal portion of a depression that fills up with water forming a lake. The size of the lake and volume of water accumulated depends on the gradient and width of the depression and on the height of the morainal dam. Most Neoglacial moraines consist of a heterogeneous noncohesive mixture of blocks, boulders, sand and silt that is highly susceptible to erosion. During periods of enhanced overflow, such as extreme snow-ice melt events, unusually heavy rainfall, or times of wave activity caused by icefalls, the channel at the outlet of the lake may be cut deeply into the morainal debris. This, in general, triggers a self-accelerating process of overflow and erosional downcutting that mobilizes large quantities of debris and leads to potentially destructive flows (Fig. 35). Debris flows from breached morainal dams have caused enormous damage in the mountains of north-central Peru (Cordillera Blanca and Huayuash), where these events are termed 'alluviónes'. Many of the moraines in this region are composed of granitic blocks derived from the precipitous cliffs flanking the glaciers. During the late 1920s the glaciers began to retreat from the morainal fringe and since then the lakes that accumulated on the upstream side have threatened settlements in the nearby valleys (Kinzl, 1949; Lliboutry et al., 1977). For example, on 13 March 1941, Laguna Cohup in the Cordillera Blanca drained rapidly to produce a debris flow which erased a major section of the town of Huaraz, killing several thousand people.

The most effective countermeasures against debris flows from moraine-dammed lakes are: (1) preparation of a deep and stable cut across the moraine *prior* to the formation of a major lake, and (2) construction of masonry-concrete revetments along the outlet channel to prevent entrenchment and collapse of embankments during periods of enhanced overflow. Both measures have been applied successfully in the past.

Debris flows from bursts of subglacial or englacial water pockets

Some glaciers, particularly small cirque glaciers perched against steep and irregular bedrock terrain, develop basal or internal cavities filled with water during times of enhanced snow-ice melt. Channels through or beneath the glacier may drain such water-filled cavities rapidly (Jackson, 1979). Alternatively, drainage may occur when the frontal part of the glacier retaining the water pocket collapses. The sudden release of water into steep debris-choked upland torrents generally leads to debris flows. Historical debris flows of destructive impact in the Alps originating in this manner include those of St. Gervais-Tête Rousse and Ötz Valley (*see* Appendix).

Direct countermeasures against bursting subglacial and englacial water pockets are generally expensive or impractical. If the location of a cavity is known, however, it can be drained by a tunnel. Also, in areas of recurrent flows, stabilization of torrent embankments and protective works against debris flows may be appropriate.

Ice avalanches

Ice avalanches occur where a glacier terminates on a smooth bedrock slope inclined more than 30°. Failure of glacier snouts under such conditions is promoted by high ambient air temperatures leading to the development of a thin film of water along the base of the body of ice (Heim, 1895). This water triggers an increase in the rate of slippage of the glacier over its bed, which in turn leads to the widening of transverse crevasses above a zone of extension. Accelerated slippage may lead to failure along concave transverse crevasses reaching to bedrock; however, accelerated slippage does not always lead to complete failure of the extending zone of ice and may result only in increased icefall activity at the toe of the glacier (Röthlisberger and Aellen, 1970; Röthlisberger, 1974, 1977). Monitoring by photogrammetric techniques and by cable tools attached to the moving ice have been used to chart the development of enhanced slippage of the glacier bed and thus to provide an indication of potential failure (Röthlisberger, 1974, 1981).

Alpine case histories of destructive ice avalanches are those of Altels, Simplon, Randa, Mattmark and Glacier du Tour.

Ice-rock avalanches

Where hanging glaciers or ice caps rest on fractured bedrock infiltration of meltwater from the base of the glacier into open fractures may lead to failure involving both rock and ice. This type of failure is particularly common in glacier covered slopes underlain by volcanic materials (*see* chapter on Mass Movements on Volcanoes): however, ice-rock avalanches occur in other materials as well. Indeed, the most tragic failures of this type, which led to the rock-ice avalanches from the spectacular summit wall of Huascaran, Peru, in 1962 and 1970, involved a pedestal of massive granitic rocks which failed along a composite steep fracture zone beneath the summit glacier (Morales Arnao, 1966; Plafker and Erikson, 1978).

Large rock-ice avalanches are released during strong earthquakes (Post, 1967) and may develop into devastating debris flows as the constituent ice melts (*see* Alpine case histories of Tête Noire, Diablerets, Val Ferret — Val Veni, Simplon, and Becca de Luseney).

ROCKFALLS AND ROCK AVALANCHES

Huge lobes composed of blocky debris stretching across the floor of drift-filled mountain valleys have always attracted the attention of travellers and pioneer settlers, simply because such features are obstacles to transportation and agricultural activites. In the European Alps many composite place names contain the local expressions for blocky debris, such as 'liappey', 'clap', 'rovina', 'runia', 'Schutt', 'masere'. However, early settlers made only a vague distinction between the deposits of debris flows, rockfalls, and rock avalanches because often these phenomena resulted in similar chaotic block fields. Only the direct observations of major catastrophic rockfalls and rock avalanches, first made during the last century, have led to an improved understanding of the destructive impact and complexity of failure and subsequent motion of large rockslides. Eyewitness reports of survivors became the most important source of information used by early scientific investigators (Zay, 1807; Buss and Heim, 1881) and are still significant today.

The most baffling and therefore most frequently discussed aspect of rockfalls and avalanches is that, given a certain volume, grain size, and geometry of runout zone, the failed rock mass flows or streams (Heim, 1882a, 1932; Hsü, 1975). A growing literature has confirmed that the 'mobility' ('reach') of dry rock avalanches increases markedly with increasing volume (Heim, 1932; Shreve, 1968a; Scheidegger, 1973; Abele, 1974; Hsü, 1975; Eisbacher, 1979; Lucchitta, 1979; Davies, 1982). Although for the purpose of dealing with the hazard of potential rockfalls or avalanches the general quantitative relationship between volume and reach is far from satisfying, it is clear that there is a certain threshold volume at which the free tumbling of individual blocks (rockfall) changes into a coherent streaming of disintegrating rock masses (rock avalanche). Depending on the geology of the detachment zone and the geometry of the rock avalanche path this transition takes place at volumes ranging from about 0.1×10^6 m^3 to 1×10^6 m^3.

Rockfall deposits accumulate characteristically as talus cones and aprons whose inclination varies between 20 and 40° depending on particle size, lithology, and the degree of reworking by running water. Only the largest blocks tumble beyond the toe of the talus slope (Broili, 1974). In contrast, rock avalanches have a far greater reach; some extend many kilometres beyond the toes of existing talus slopes, and their fronts may climb ridges up to several hundred metres high.

For a brief discussion, the rock avalanche process is best viewed in two stages: the mechanics of cliff failure (stability) and the dynamics of motion (mobility, reach) of the failed rock masses.

Mechanics of failure

Rockfalls and rock avalanches develop most commonly on steep bedrock slopes underlain by fractured but otherwise competent rock formations such as thick-bedded sedimentary rocks (e.g. carbonates, sandstones, and conglomerates) or massive igneous or metamorphic rocks (granite and gneiss).

They occur rarely on slopes underlain by thin-bedded sedimentary rocks and low grade metamorphics. Most rockfalls and rock avalanches originate on bare bedrock cliffs rising high above the valley floors.

One of the requirements for the generation of a rock avalanche is the rapid detachment of a large, relatively intact rock mass along a well defined rupture surface. Slow detachment creates only rockfalls from the toe of the incipient failure. Whether rockfall activity from the toe is the only thing that will occur along a rock face or whether this activity represents the preparatory stage for a major rock avalanche is one of the most difficult questions regarding failure mechanics. Indeed, rockfall activity does precede the detachment of major rock slides, particularly those related to the buildup of high pore water pressures during rainfall or snowmelt, and rockfall activity simply indicates that the stressed toe zone of the slope is expanding by cracking and toppling in its outermost rock skin. However, rockfall activity need not precede failure of cliffs triggered by shaking during earthquakes. Seismic shaking in the epicentral region of major earthquakes (magnitude 6 and greater) is the most common trigger mechanism for large rock avalanches. The distortion of massive rock faces by seismic shaking invariably causes widening of existing fractures and thus sets in motion a complex process involving loss of cohesion, slippage, internal toppling, crushing, and eventual failure along a composite detachment surface. The geometry of this surface is generally determined by prominent internal discontinuities such as bedding planes, fractures and faults (Fig. 36). In many seismically active mountain chains the largest rock avalanches are located close to surfaces of seismic rupture (e.g. Hadley, 1964; Solonenko, 1977; Keefer and Tannaci, 1981).

Commonly, the failure of a rock mass is also 'prepared' by a slow deterioration of its cohesive properties during repeated intervals of extreme saturation with rising groundwater tables or recurrent seismic activity (Voight, 1978). The exact geometry of the eventual rupture surface for major rockslides therefore develops by a progressive amalgamation or propagation of single discontinuities and only a rough idea of the potential detachment zone may be gained from the geometry of incipient crown cracks. In addition, it is often difficult to decide whether a set of surface cracks corresponds to prehistorical fault scarps or to incipient crown cracks of gravitationally unstable slope segments (e.g. Radbruch-Hall, 1978; Clague, 1982; Bovis, 1982). With respect to composition and bedrock structure of natural mountainsides three main categories of slopes producing rock avalanches can be differentiated: scarp slopes in sedimentary rocks, dip slopes in sedimentary rocks, and fracture-controlled slopes in schist, gneiss, and granite.

Scarp slopes in sedimentary rocks

These develop preferentially in competent formations such as limestone-dolomite successions and conglomerates; occasionally, deformed sandstone-slate assemblages form impressive scarp slopes. Scarp slopes may be controlled by regional fracture sets trending parallel to major fold structures ('tectonic joints') or by fractures related to erosion and unloading ('sheeting'). In practice these two types of fracture sets are difficult to separate from each other. Surface-parallel sheeting tends to be wider spaced at increasing depth behind the rock face and may parallel irregularities of the rock face surface itself. Most scarp faces are sculptured by bedrock ravines that follow fractures or faults. Crushed and shattered bedrock flanking the ravines supplies a steady source of rockfall debris to talus cones at the base of the cliff (Schumm and Chorley, 1964; Gardner, 1980).

Incipient failure of scarp slopes is commonly preceded by the development of arcuate crown cracks in competent bedrock formations resting on more incompetent formations such as shale or evaporites. Subsidence and outward flow of the incompetent formations induce a propagation of steep cracks and their coalescence into potential detachment surfaces across the competent rock units above. Water infiltrating the cracks accelerates subsidence, thus bringing closer the time of large-scale failure. Depending on the location and geometry of the detachment surface, failure may proceed by forward toppling, backward rotation and sliding, or by complex internal collapse. Alpine case histories presented in the Appendix include Reisskofel, Dobratsch, Ziano, Radmer an der Hasel, Corbeyrier-Yvorne, Vorder Glärnisch, Salzburg, La Valle Agordina, Diablerets, Massif de Platé, Lake Lucerne, Brannenburg, Bilten, Elm, Bocca di Brenta, Altdorf-Spiringen, Pra-Lagunàz, Vandans, Sandling, Linthal, Zambana, Illgraben, and Lecco.

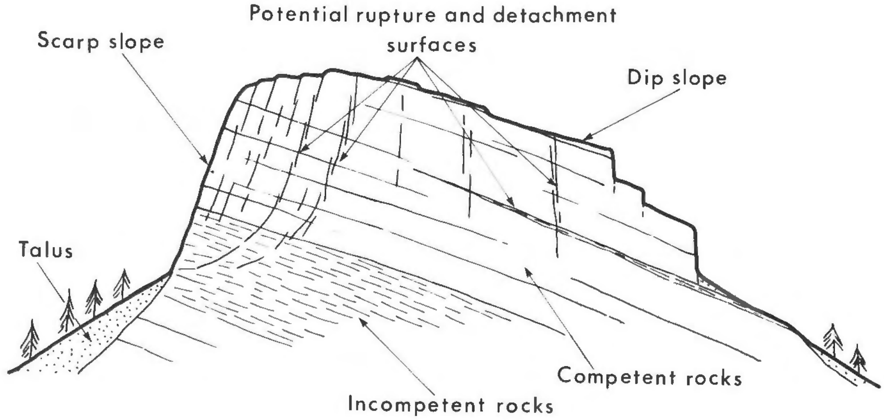


Figure 36: Schematic illustration of the geological discontinuities controlling dip slope and scarp slope failures with a potential for rockfall and rock avalanche development.

Dip slopes in sedimentary rocks
These are controlled by bedding planes of competent lithological units such as thick-bedded carbonates, conglomerates, and sandstones. Steep regional fracture sets cutting across bedding commonly create a stepped profile of otherwise uniformly rising bedding planes and also control the orientation of high back walls at the crest of many dip slopes. Dip slopes of massive sedimentary rocks are susceptible to the development of large rock avalanches if they are characterized by smooth bedding planes or by intercalated recessive shale horizons that outcrop above the bottom of the valley (Fig. 36).

Dip slope failures have resulted in some of the most devastating historical rock avalanches (Heim, 1932; Hsü, 1975). The largest rock avalanches (i.e. those in excess of 300×10^6 m^3) tend to break away on detachment surfaces that dip less than 20°. Most commonly, however, dip slopes fail along bedding planes inclined about 30°.

The mechanism of the often explosive failure of dip slopes in sedimentary rocks is still poorly understood. The angle of friction for dry rock surfaces, which ranges between about 28 and 32°, can explain failure of dry dip slopes along through-going bedding planes or shear surfaces that dip about 30° (Cruden and Krahn, 1973; Cruden, 1976; Mathews and McTaggart, 1978). Dip slopes with bedding inclinations of less than 30° commonly fail along argillaceous horizons above which elevated pore water pressures develop during sustained periods of rainfall or snowmelt. However, many rock avalanches become detached on dip slopes with gently inclined bedding planes (10 to 20°) without obvious impermeable interbeds. These failures are difficult to explain. Most of them are probably triggered by earthquakes which drastically reduced the cohesional bonds between large blocks of the rock mass. A complex mechanism of progressive failure involving rotation, translation, crushing, and collapse of constituent blocks may thus be initiated by the earthquake (Eisbacher, 1979; McLellan, 1983). Dilation of the rock mass above gently dipping bedding planes also seems to be required to account for the great reach attained by some of the resulting rock avalanches. Failure of dip slopes with bedding inclinations of more than 50° generally involves buckling and toppling of beds, and propagation of shear planes across surfaces of stratification at the toe of the incipient failure. Alpine case histories described in the Appendix include Lavini di Marco, Masière de la Vedana, Mont Granier, Clavans, Leytron, Meiringen, Lago di Alleghe, Lake Lucerne, Goldau, Antelao, Felsberg, Bec Rouge, Blisadona, Fidaz-Flims, Vaiont, and Tagliamento Valley.

Fracture-controlled slopes in schist, gneiss, and granite
These form some of the most spectacular mountains on Earth, ranging from vertical towers of granite to smooth or intricately sculptured scarp faces of metamorphic rocks in the core zones of many young orogenic belts. Although schistosity and foliation commonly impart a distinct asymmetry to mountains, these pervasive internal structures rarely control the detailed slope morphology or the trend of major detachment surfaces of rock avalanches. In gneissic or granitic terrain rock avalanches break away most commonly along composite fracture zones that transect the internal planar fabrics of the rock mass at high angles, and outcrop at the bottom of the cliffs. Failure often involves only the outermost wedge-shaped segments of large mountainsides undergoing slow and deep-seated deformation. Rapid failure of such bedrock spurs is often preceded by long preparatory periods during which concave crown cracks and lateral fractures expand and coalesce into an anastomosing pattern of potential rupture surfaces. Metamorphic bedrock slopes covered by thick surficial debris may fail as complex rock-debris avalanches. Alpine case histories include Lienz, Mottec, Plaine d'Oisans, Ganderberg, Zarera, Biasca, Piuro, Salvan, Antronapiana, Disentis, Val Ferret-Val Veni, Grächen, Monbiel, Spriana, Airolo, and Simplon.

Dynamics of motion

One of the most interesting aspects of rock avalanches is the exceptional travel distance of their frontal parts. Most of the destructive impact of rock avalanches results from this seemingly incomprehensible reach of the dry blocky material. That rock avalanches 'stream' was first documented by Heim (1882a, b) for the catastrophic rock avalanche at Elm, Switzerland. Since then, many other far-travelled prehistorical and historical rock avalanches have been described including gigantic examples from extraterrestrial settings (Heim, 1932; Shreve, 1968a; Scheidegger, 1973; Abele, 1974; Hsü, 1975; Pariseau and Voight, 1979; Eisbacher, 1979; Lucchitta, 1979).

The paradox of rock avalanche mobility is that an essentially dry blocky mass spreads and streams far beyond the toe zone of the failed slope, often climbing over major obstacles before coming to rest. This unusual behaviour is best expressed in terms of an 'excessive travel distance' (Hsü, 1975) or the 'deposit length' (Davies, 1982) of rock avalanches. Excessive travel distance L_e is the distance between the frontal edge of a slide mass with an expected angle of friction of 32°, and the *actual* frontal edge of a rock avalanche. It is expressed by the formula

$$L_e = L - H/\tan 32°$$

where L and H are horizontal and vertical distances as shown in Figure 37. The simpler 'deposit length' is the horizontal distance between the most proximal and most distal parts of a rock avalanche deposit.

If rock avalanches travel over reasonably gentle valley floors their excessive travel distances or deposit lengths are related to the volume of the failed rock masses. Davies (1982) has plotted deposit length against volume on a doubly logarithmic chart and found that many historical and prehistorical rock avalanches cluster along a relatively straight band. This relationship is expressed by

$$L_d = 9.98\ V^{0.33}$$

where L_d is deposit length in m, and V is volume in m^3. Such a broad relationship between the reach of rock avalanches and their volume has also led to the suggestion that their 'coefficient of friction' (Scheidegger, 1973) or 'effective coefficient of friction' (Shreve, 1968a), defined by the ratio H/L of Figure 37, varies with their volume.

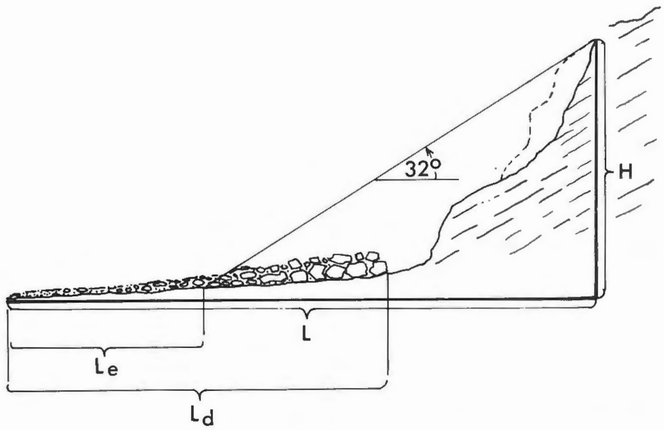


Figure 37: Descriptive elements of rock avalanches; broken line indicates the position of the rock mass before failure (see text for explanation).

Unfortunately, the predictive value of such calculations has only limited applicability, as it requires a knowledge of the exact volume of the potential rock avalanche *prior* to failure. This value is generally poorly known, because even if a rock mass shows incipient failure it may become detached as a single piece or break away in a series of minor rock avalanches or rockfalls. Furthermore, the reach of rock avalanches also depends on the lithology of the failed mass, its exact path below the detachment zone, and the kind of material picked up along the path (e.g. water, snow, or ice). Abele (1974) has demonstrated statistically that rock avalanches composed of massive blocky sedimentary rocks have a greater reach than those composed of high grade metamorphic rocks; the latter tend to be coherent after initial failure. A winding, narrow, or rising path also introduces baffling complexities in rock avalanche behaviour (Eisbacher, 1979). Incorporation of significant amounts of wet soil, ice, snow, or water along the front of a moving rock avalanche tends to change its character into that of a debris flow; a complete range of transitional phenomena from 'stiff' dry rock avalanches to water-saturated flows have been described (Zankl, 1958; Plafker and Erikson, 1978; Kelly, 1980). Notwithstanding these complexities, dry rock avalanches with volumes of less than $10 \times 10^6 m^3$ rarely travel more than about 2 km, and those with volumes in excess of $100 \times 10^6 m^3$ in general travel more than 3 km.

It is generally recognized that during their initial movement rock avalanches gain velocity by gravity-induced acceleration of the entire mass. However, the detailed dynamics of streaming by which excessive travel distances are attained are still far from resolved. Much has been learned from eyewitness accounts of major catastrophes, such as those collected by Buss and Heim (1881). However, such observations are commonly hampered by the development of enormous clouds of rock dust rising from and enveloping the moving stream of fragmented rock. Dust is also blown ahead of the rock avalanche by powerful air blasts. Thus details of the movement dynamics must be derived from the deposits of rock avalanches that have come to rest completely.

In most rock avalanche deposits the bulk of the shattered mass comes to rest at the foot of the failed cliff. In both the proximal parts and the thinning distal lobes the sequential order of rock units as it existed prior to failure on the cliff is often preserved. Although local upward mobility of fine debris is indicated by clastic dykes and other features, the preservation of internal vertical stratigraphic ordering rules out large scale turbulence during the rapid motion (Heim, 1882a; McSaveney, 1978, Eisbacher, 1979; Erismann, 1979). Diverging (and locally converging) 'stream lines' on fresh surfaces of rock avalanches suggest a dilating sheet-like motion of the frontal part (McSaveney, 1978). Sharp-crested lateral and transverse ridges point to sliding during the final stages of movement and also confirm observations of the sudden arrest of moving rock avalanche fronts. Only near sharp bends and at steps are individual blocks propelled outside the main stream (Plafker and Erikson, 1978; Eisbacher, 1979).

The kinematics 'frozen' into deposits of rock avalanches have to be satisfied by any proposed explanation of their mobility. Such explanations have invoked transfer of momentum between small fragments during streaming ('living force' of Heim, 1882a), transfer of momentum between major blocks shortly after failure (Eisbacher, 1979), the formation of trapped cushions of air (Shreve, 1968b), fluidization of coarse material in a fine grained matrix (Kent, 1965; Hsü, 1975; Davies, 1982), and generation of steam along the basal zone of rupture and shear (Goguel, 1978; Erismann, 1979). All of these mechanisms may interact and enhance mobility along certain reaches of the rock avalanche path. Where rockfalls or avalanches plunge into lakes or fiords the wave trains set up by the impact may have considerable destructive potential (Jorstad, 1968). A large slide volume and shallow water favour the development of powerful waves (Slingerland and Voight, 1979; Huber, 1980).

In summary, the mechanics of slope failure and the dynamics of rock avalanche mobility depend greatly on the geology of the mountainside, the volume of the detached rock mass and the character of the path. Management of rockfall and rock avalanche hazard has to contend with these variables, for which data are generally difficult to obtain.

Management of rockfall and rock avalanche hazard

The management of rockfall and rock avalanche hazard, in analogy with management of debris flow hazard, follows three steps: hazard appraisal, application of passive measures, and application of active measures. In contrast to the problems encountered with debris flows, the large size and low recurrence of rock avalanches in any particular area may require that risk to property be separated from risk to life — destruction of property may be an acceptable risk, loss of life may not be acceptable.

Hazard appraisal

Recurrent rockfall activity from a steep bedrock slope generally results in fresh aprons or cones of talus at the foot of the slope. Blocks that have tumbled down a talus slope may scar trees (Fig. 38). Careful analysis of trees may give a hint as to the frequency of the falls, if such information cannot be obtained from local newspapers or other accounts. Accelerating rockfall activity from steep cliffs also tends to precede major slope failures; rockfall activity thus can guide a closer inspection of the area in which the rockfalls originated. The possibility of a major failure that might result in a rock avalanche with a reach beyond the foot of the talus slope is real when crown cracks and open fractures delimit an incipient slide mass in excess of 0.5 to $1 \times 10^6 m^3$. Under such conditions detailed monitoring by surveys and/or evacuation of settlements may be required.

Deposits of large prehistorical rock avalanches, particularly those more than 1000 years old, are hazard indicators only in a very general sense. A prehistorical failure may have improved the stability of the remaining cliff to such a degree that the present hazard is negligible; in other cases, however, a failure along one section of a bedrock slope may reduce the stability and set the stage for a slope failure on an even larger scale along an adjacent section. Each case has to be judged individually; there are no general rules. Whether a rock slope will continue to deteriorate in the form of small and tolerable rockfalls or whether rockfalls will change into a catastrophic cliff failure has to be determined mainly from the observation of survey points installed on the slope: if the cliff on the whole is at rest and only its lower part produces rockfalls, no major hazard exists; if, however, an ever increasing volume of rock shows a tendency towards accelerated motion, a major rockslide may be in the making. Theoretical calculations of reach and velocities attained by rock avalanches have been attempted by Banks and Strohm (1974), Korner (1976), and McLellan (1983). Such analyses may prove useful once the volume of an incipient failure is known.

The release of large rock avalanches in most cases tends to be preceded by periods of accelerated creep, thus providing a means for hazard appraisal (Heim, 1932; Müller, 1968; Pariseau and Voight, 1979). However, large earthquakes commonly break cohesional bonds of the rock mass within seconds or minutes, and in such cases timely warnings are impossible. In mountainous regions with known patterns of historical seismicity, such as New Zealand, the dating of past rock avalanches provides a rough estimate of the regional rock avalanche hazard posed by the recurrent earthquakes (Adams, 1980; Whitehouse and Griffiths, 1983).

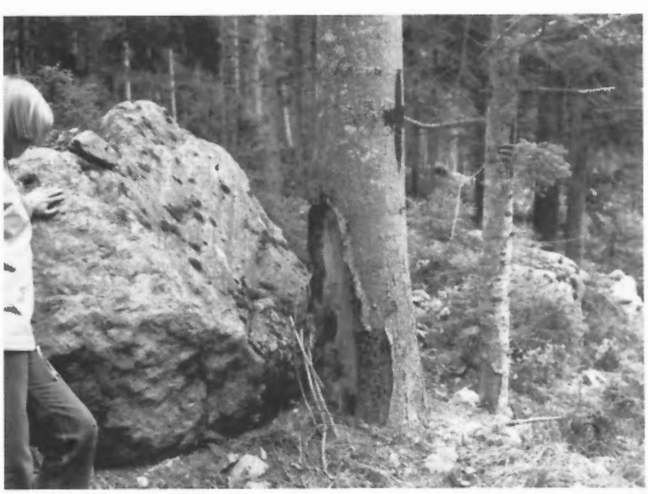


Figure 38: Tree scarred by impact of a rockfall. New growth of tree rings around the scar permits dating of the impact. (GSC 204165-K)

Passive measures

Areas exposed to recurrent rockfall activity are obviously unsuitable for permanent habitation. However, even cliffs fringed by aprons of fresh rockfall talus may not fail as catastrophic rock avalanches that reach beyond the foot of the talus. In principle such areas are difficult to place into a restricted land-use category. Indeed, many case histories and the ubiquitous settlement on prehistorical rock avalanche lobes in the Alps suggest that in densely settled mountain regions the hazard from rare, single, and large slope failures is generally accepted if they are due to natural causes. Nevertheless, regular inspection of the slopes for incipient instabilities or a monitoring program by surveys can reduce this accepted risk by making timely evacuation possible. The risk of massive slope failures created by human activity such as failure of slopes along reservoirs is generally no longer accepted or acceptable.

Active measures

Active measures against small rockfalls are mainly preventive engineering structures placed on and along steep rock faces. A wide variety of techniques are available for coping with these problems and have been reviewed extensively in the literature on rock slope engineering (e.g. Müller, 1963; Piteau and Peckover, 1978). Preventive engineering works such as retaining walls, anchored beams, shotcrete, bulkheads, and toe buttresses are costly. They may have to be supplemented by protective catchnets, timber fences or ditches, particularly along highways. Where high costs can be justified, galleries or tunnels may be constructed to completely prevent rockfall damage to roadworks and avoid interference with highway traffic.

Where incipient rock slope failure is suspected, drainage or artificial piece-by-piece removal of the threatening rock mass may be attempted. Large masses of rock (i.e. volumes in excess of $1 \times 10^6 m^3$), once in rapid motion, are almost impossible to control and engineering works tend to be futile. Timely evacuation of settlements threatened by rock avalanches is therefore the only means available to save lives.

ACCEPTABLE AND UNACCEPTABLE RISK OF DESTRUCTIVE MASS MOVEMENTS

Whether or not potential mass movements are perceived as acceptable or unacceptable hazards depends to a large extent on the social and economic context in which they occur. In the relatively wealthy mountain districts of central Europe, western North America, Japan, and New Zealand a knowledge of past mass movements based on historical data or hazard indicators is generally taken seriously during the planning of new communities, the establishment of transportation routes, and the siting of reservoirs. In developing mountain regions where such data are sparse and where scarcity of land, food, and health services are more pressing, potential slope hazards are often ignored; the eventual mass movements may exact a considerable toll in terms of human lives and economic hardship.

Although the protective function of a healthy forest cover on steep slopes is now widely appreciated, the trade-off between immediate gain by harvesting trees in large clear-cuts with poorly managed roads and the eventual loss of soil resources is still commonly decided in favour of immediate economic gain — even in highly developed regions such as parts of western North America. The risk of deteriorating slope stability after timber harvesting is thus still widely accepted. In the Alps and other parts of central Europe serious debris flow disasters in the 19th century led to legislation which substantially changed forest-related attitudes from the short-term gain towards long-range plans of reforestation and forest management. Slope hazards induced by reckless forest practices are no longer acceptable there. Restrictive regional legislation evolved from local self-imposed rules that regulated the use of land, forests, and water in steep uplands for the benefit of the whole community. Regional legislation provided additional funds for reforestation, torrent control works, and improved access roads. However, the responsibility for the safety of individual mountain settlements remained with the community leaders (mayors and councils). Because the whole community had to help members afflicted by debris disasters, there remained an incentive to keep permanent dwellings away from sites known to have been affected by recurrent mass movements.

Rapid expansion of tourism and hydroelectric developments throughout many mountain regions of the world are proceeding within a social context different from that of pioneer communities: self reliance of pioneer communities necessarily included acceptance of high levels of natural or self-inflicted risks. In contrast the modern consumer commonly expects to be protected from hazards, at least from those created by human activities. Whether such protection should be extended to naturally recurring mass movements is problematic. On the one hand strict hazard zoning *prior* to development is highly unpopular with developers and individuals who feel that such measures interfere with their freedom of choice. On the other hand the same citizens often tend to blame inadequate government involvement after destructive mass movements have occurred. A balance between these two attitudes has to be achieved.

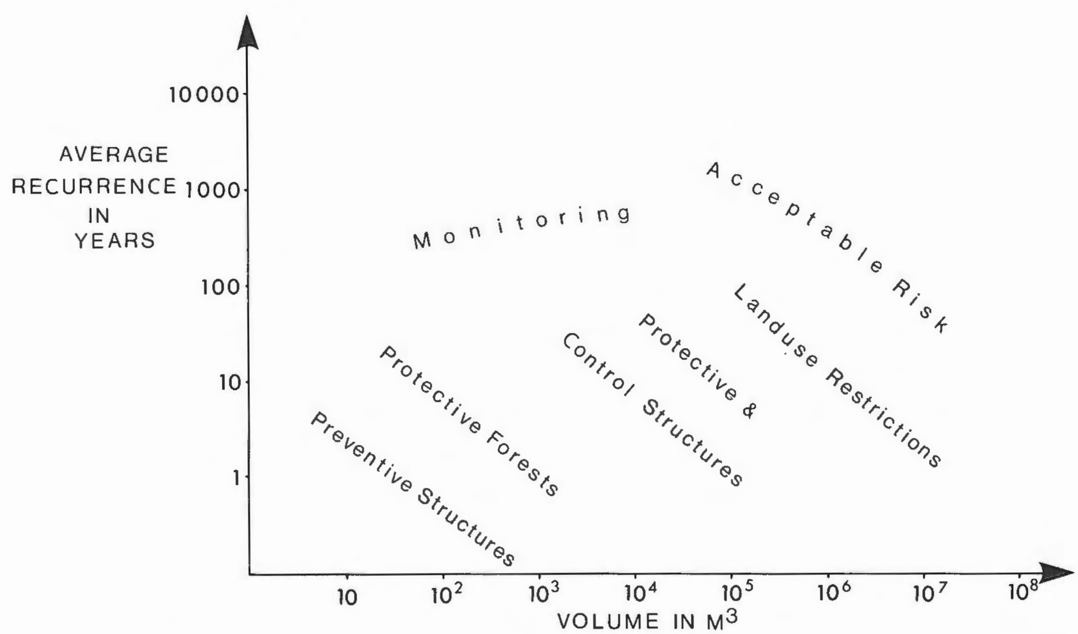


Figure 39: Summary diagram derived from numerous case histories illustrating the most effective use of passive and active measures in reducing the impact of destructive mass movements in developed mountain regions depending on their volume and projected recurrence. Most of the measures shown can be applied in combination with each other and their position in the diagram depends greatly on the social-economic context of a specific situation.

A clear requirement for future development of mountain valleys will be a very high standard of forest management in the vicinity of permanently inhabited areas. Beyond that, hazard appraisals with respect to potential mass movements, construction of control or protective works, the local zoning may be desirable prior to and during development. In the densely populated tourist districts of central Europe zoning and protective engineering are now mainly directed towards mass movements with potential recurrence intervals of less than 100 to 200 years (Eisbacher, 1982); the cost of protective works is borne by individuals, communities, or higher levels of regional governments. In general, it makes little economic sense to provide protective works if costs exceed a significant proportion (say 30 to 50%) of the value of properties to be protected. Potential for loss of life, however, may render the economic argument irrelevant.

The incidence of single large events beyond the time frame of a few hundred years may fall under acceptable risk. Slopes which show signs of incipient failure, located directly above developed valley bottoms or in the source area of high-gradient torrents, should be monitored by inexpensive and regular surveys. Timely evacuation may save lives.

Figure 39 is a diagrammatic summary of the approaches that are used to combat mass movement hazards in mountain regions. These measures may be taken singly or in combination. The diagram, based on the case histories described in the Appendix, underlines the fact that no hard-and-fast rules exist as to what is an acceptable or unacceptable risk. To be useful regional legislation should provide planners not only with enforceable tools ('zoning laws'), but also ensure that potentially afflicted individuals have an incentive to bear the risk of damage or the cost of remedial structures ('buyer beware'). The resulting land use pattern therefore always requires a social consensus as to what risk is acceptable.

REFERENCES

(Includes references cited in the Appendix)

Abele, G.
1974: Bergstürze in den Alpen; Wissenschaftliche Alpenvereinshefte, no. 25, 230 p.

Abich, H.
1847: Geognostische Reise zum Ararat und Verschüttung des Thales von Arguri im Jahre 1840; Monatsberichte uber die Verhandlungen der Gesellschaft fur Erdkunde zu Berlin, Neue Folge, v. 4, p. 28-62.

Adams, J.
1980: Earthquake-dammed lakes in New Zealand; Geology, v. 9, p. 215-219.

Agassiz, L.
1840: Etudes sur les Glaciers; Neuchatel, 346 p.

Allix, A.
1929: L'Oisans au Moyen Age; Paris, 266 p.

Ampferer, O.
1939: Über einige Formen der Bergzerreissung; Sitzungsberichte der Akademie der Wissenschaften Wien, Math.-Naturw. Klasse, Abt. 1, v. 148, Heft 1/2, p. 1-14.

Anderson, M.G. and Burt, T.P.
1978: The role of topography in controlling throughflow generation; Earth Surface Processes, v. 3, p. 331-344.

Anderson, T.
1908: Report on the eruptions of the Soufrière in St. Vincent in 1902, and on a visit to Montagne Pelée in Martinique, Part II. The changes in the districts and the subsequent history of the volcanoes; Royal Society of London, Philosophical Transactions, ser. A, v. 208, p. 275-303.

Anderson, T. and Flett, J.S.
1903: Report on the eruptions of the Soufrière, in St. Vincent, in 1902, and on a visit to Montagne Pelée, in Martinique, Part I; Royal Society of London, Philosophical Transactions, ser. A, v. 200, p. 353-553.

Andreatta, C.
1956: I crolli della parete Paganella ed il convogliamento di detriti su Zambana (Trento); Acta geologica Alpina (Bologna), no. 6, p. 1-30.

Angelini, G.
1976: Civetta, per le Vie del Passato; Bologna, 361 p.

Antoine, P.
1978: Glissements de terrains et aménagement de la montagne; Bulletin des Laboratoires Géologiques, Université de Lausanne, v. 74, fasc. 1, p. 1-14.

Antoine, P., Dayre, M., Fabre, D., Giraud, A., and Letoureur, J.
1979: Notice explicative de la carte des zones exposées à des risques liés aux mouvement du sol et du sous-sol (Région de Moutiers, Savoie); Bureau de Recherches Géologiques et Minières, Orléans, 31 p.

Aramaki, S.
1956: The 1783 activity of Asama volcano, part 1; Japanese Journal of Geology and Geography, v. 27, no. 2-4, pt. 1, p. 189-229.

Aretin, G.
1808: Über Bergfälle und die Mittel, Denselben Vorzubeugen; Innsbruck, 86 p.

Aulitzky, H.
1968: Analyse der Schadenursachen von Unwetterkatastrophen zum Zweck der Vorbeugung; Österreichische Wasserwirtschaft, v. 20, p. 90-97 and p. 144-154.
1970: Der Enterbach (Inzing in Tirol) am 26.Juli 1969; Wildbach- und Lawinenverbau, v. 34, p. 31-66.
1972: Möglichkeiten und Grenzen der Vorbeugung vor Unwetterkata-strophen; Österreichische Wasserwirtschaft, v. 24, p. 8-17.
1974: Endangered Alpine regions and disaster prevention measures; Council of Europe, Nature and Environment Series, no. 6, 103 p.
1980: Preliminary two-fold classification of torrents; International Symposium Interpraevent, 1980, v. 4, p. 285-309.

Azimi, C., Desvarreux, P., and Plotto, P.
1980: Les coulées de boue du Freney, près de Modane, en 1977 et 1978; Revue Francaise de Géotechnique, no. 13, p. 5-13.

Baak, J.A.
1949: A comparative study on recent ashes of the Java volcanoes Smeru, Kelut, and Merapi; Buitenenzorg, Java, Mededeelingen, Algemeene, Proefstation voor Landbouw, no. 83, 60 p.

Bachmann, R.C.
1978: Gletscher der Alpen; Hallwag Verlag, Bern, 304 p.

Bagnold, R.A.
1968: Deposition in the process of hydraulic transport; Sedimentology, v. 10, p. 45-56.

Baltzer, A.
1873: Der Glärnisch — ein Problem alpinen Gebirgsbaues; Zürich, 100 p.
1875: Über einen neuerlichen Felssturz am Rossberg, nebst einigen allgemeinen Bemerkungen über derartige Erscheinungen in den Alpen; Neues Jahrbuch für Geologie und Paläontologie, 1875, p. 15-26.

Banks, D.C. and Strohm, W.E.
1974: Calculations of rock-slide velocities; *in* Advances in Rock Mechanics; National Academy of Sciences, Washington, D.C., p. 839-847.

Baratta, M.
1901: I Terremoti d'Italia; Rome, 950 p.

Barberi, F. and Gasparini, P.
1976: Volcanic hazards; International Association of Engineering Geology, Bulletin, no. 14, p. 217-232.

Bauer, W.
1971: Die Brienzer Wildbäche; International Symposium Interpraevent 1971, v. 2, p. 229-237.

Baumgartner, P.
1980: Erd- und Schuttströme im Gschliefgraben bei Gmunden am Traunsee — Zu ihrer Entstehung, Entwicklung und Sanierung; International Symposium Interpraevent 1980, v. 4, p. 85-102.

Bazing, V.
1872: Waldvernachlässigung und Waldverwüstung in Tirol; Zeitschrift des Deutschen und Österreichischen Alpenvereins, v. 3, p. 319-350.

Beattie, W.
1836: Switzerland; London, 2 volumes.

Becker, F.
1883: Der Bergsturz der Diablerets; Jahrbuch des Schweizer Alpenclub, v. 18, p. 310-316.

Bell, D.H.
1976: High intensity rainstorms and geological hazards: cyclone Alison, March 1975, Kaikoura, New Zealand; International Association of Engineering Geology, Bulletin, v. 14, p. 189-200.

Bénévent, E. and Maury, E.
1927: Les grandes pluies de l'automne 1926 et la catastrophe de Roquebillière (Alpes Maritimes); Revue de Géographie Alpine, v. 15, p. 151-157.

Bertrand, L.
1927: Sur les causes géologiques des recents éboulement de la vallée moyenne de la Vésubie, dans les Alpes Maritimes; Société Géologique de France, Bulletin, ser. 4, v. 27, p. 35-42.

Bibolini, A.
1909: La catastrophe di Pra e Lagunàz nell' Agordino; Bolletino del Regio Comitato Geologico d'Italia, Anno 1909, p. 62-67.

Bjornsson, H.
1975: Explanation of jökulhlaups from Grimsvotn, Vatnajokull, Iceland; Jökull, v. 24, p. 1-26.

Blumenthal, M.
1925: Der Bergsturz von Monbiel bei Klosters aus dem Jahre 1770 und die geologische Beschaffenheit seines Gehänges; Bündner Monatsblatt, p. 97-112.

Boffo, B.
1977: Nidwalden; Hochwasserschutz in der Schweiz, Bern, p. 79-82.

Böhm, A.
1886: Die Hochseen der Ostalpen; Mitteilungen der Geographischen Gesellschaft Wien, v. 29, p. 625-647.

Bonis, S., Bohnenberger, O.H., and Dengo, G.
1970: Mapa geologico de la Republica de Guatemala (1st ed.); Instituto Geografico Nacional, Guatemala City.

Booth, B.
1977: Mapping volcanic risk; New Scientist, v. 75, p. 743-745.

Bovis, M.J.
1982: Uphill-facing (antislope) scarps in the Coast Mountains, southwest British Columbia; Geological Society of America, Bulletin, v. 93, p. 804-812.

Bradley, W.C. and Mears, A.I.
1980: Calculations of flows needed to transport coarse fraction of Boulder Creek alluvium at Boulder, Colorado; Geological Society of America, Bulletin, v. 91, pt. 1, p. 135-138, and pt. 2, p. 1057-1090.

Broili, L.
1967: New knowledge on the geomorphology of the Vaiont slide slip surface; Rock Mechanics and Engineering Geology, v. 5, p. 38-88.
1974: Ein Felssturz im Grossversuch; Rock Mechanics, Supplement 3, 1974, p. 69-78.

Brückl, E. and Scheidegger, A.E.
1972: The rheology of spatially continuous mass creep in rock; Rock Mechanics, v. 4, p. 237-250.

Brunsden, D., Jones, D.K.C., Martin, R.P., and Doornkamp, J.C.
1981: The geomorphological character of part of the Lower Himalaya of Eastern Nepal; Zeitschrift für Geomorphologie, Supplement, v. 37, p. 25-72.

Bullard, F.M.
1976: Volcanoes of the Earth (revised ed.); University of Texas Press, Austin, 579 p.

Bunza, G., Karl, J., and Mangelsdorf, J.
1976: Geologisch-morphologische Grundlagen der Wildbachkunde; Schriftenreihe der Bayerischen Landesstelle für Gewässerkunde München, Heft 11, 128 p.

Burri, M. and Gruner, E.
1976: Phénomènes d'instabilité dans les vallées des Dranse (Valais); Eclogae Geologicae Helvetiae, v. 69, p. 75-83.

Burroughs, E.R., Chalfant, G.R. and Townsend, M.A.
1976: Slope Stability in Road Construction; U.S. Bureau of Land Management, Oregon State Office, 102 p.

Buss, E. and Heim, A.
1881: Der Bergsturz von Elm, den 11 September 1881; Zürich, 163 p.

Buxtorf, A. and Wilhelm, O.
1920: Über Sackungserscheinungen im Safiental, Graubünden; Eclogae Geologicae Helvetiae, v. 16, p. 520-527.

Cadisch, J.
1970: Bergstürze in der Sicht des Geologen, mit besonderer Berücksichtigung der Katastrophe von Vaiont; Bulletin der Vereinigung Schweizerischer Petroleum-Geologen und-Ingenieure, v. 36, p. 53-66.

Caine, N.
1980: The rainfall intensity — duration control of shallow landslides and debris flows; Geografiska Annaler, v. 62A, p. 23-27.

Caine, N. and Mool, P.K.
1982: Landslides in the Kolpu Khola drainage, Middle Mountains, Nepal; Mountain Research and Development, v. 2, p. 157-173.

Campbell, R.H.
1975: Soil slips, debris flows, and rainstorms in the Santa Monica Mountains and vicinity, southern California; U.S. Geological Survey, Professional Paper 851, 51 p.

Casal, E.
1898: La cuduta del Monte Spiz, nel 1771; Rivista Mensuale del Club Alpino Italiano, v. 17, no. 6, p. 212-215.

Castiglioni, G.B., Pellegrini, G.B., Nardin, M., Ferrari, L., and Mattana, U.
1974: Note di commento alla carta dell'alluvione del novembre 1966 nel Veneto e nel Trentino-Alto Adige:effeti morphologici e allagamenti (con una carta allegata); *in* Le Calamitá Naturali nelle Alpi, ed. G.B. Castiglioni; Istituto di Geographia dell'Universita di Padova, Novara, p. 269-290.

Catullo, A.
1814: Sopra le Rovine ch'ebbero Luogo nel Commune di Borca nel Cadorino; Belluno, 15 p.

Cavallin, A. and Martinis, B.
1974: Studio geologico della grande frana di Borta (Ampezzo); Cronaca della Societa Alpina Friuliana, v. 58, p. 297-319.

Chabert, L.
1978: Aimer la Maurienne; St.-Alban-Leysse, 190 p.

Chavoutier, L.
1979: Aime et ses environs; L'histoire en Savoie, v. 14, no. 54, 48 p.

Clague, J.J.
1979: The Denali Fault System in southwest Yukon Territory — a geologic hazard?; *in* Current Research, Part A; Geological Survey of Canada, Paper 79-1A, p. 169-178.
1982: The role of geomorphology in the identification and evaluation of natural hazards; *in* Applied Geomorphology, ed. R.G. Craig and J.L. Craft; George Allen & Unwin, London, p. 17-43.

Clague, J.J. and Rampton, V.N.
1982: Neoglacial Lake Alsek; Canadian Journal of Earth Sciences, v. 19, p. 94-117.

Clague, J.J. and Souther, J.G.
1982: The Dusty Creek landslide on Mount Cayley, British Columbia; Canadian Journal of Earth Sciences, v. 19, p. 524-539.

Clar, E.
1959: Geologische Beobachtungen von der steirischen Unwetter-katastrophe vom August 1958; Geologie und Bauwesen, v. 24, p. 131-140.

Clar, E. and Weiss, E.
1965: Erfahrungen in Talzuschub des Magnesitbergbaues auf der Millstätter Alpe; Berg- und Hüttenmännische Monatshefte, v. 110, p. 447-460.

Clavuot, O.
1973: Heimatbuch Thusis-Viamala; Thusis, 296 p.

Cotecchia, V.
1978: Systematic reconnaissance mapping and registration of slope movements; International Association of Engineering Geology, Bulletin, v. 17, p. 5-37.

Crandell, D.R.
1971: Postglacial lahars from Mount Rainier volcano, Washington; U.S. Geological Survey, Professional Paper 677, 75 p.
1973: Map showing potential hazards from future eruptions of Mount Rainier, Washington; U.S. Geological Survey, Miscellaneous Geologic Investigations, Map I-836.
1975: Assessment of volcanic risk on the Island of Oahu, Hawaii; U. S. Geological Survey, Open-File Report 75-287.

Crandell, D.R. and Fahnestock, R.K.
1965: Rockfalls and avalanches from Little Tahoma Peak on Mount Rainier, Washington; U.S. Geological Survey, Bulletin 1221-A, 30 p.

Crandell, D.R. and Mullineaux, D.R.
1967: Volcanic hazards at Mount Rainier, Washington; U.S. Geological Survey, Bulletin 1238, 26 p.
1978: Potential hazards from future eruptions of Mount St. Helens volcano, Washington; U.S. Geological Survey, Bulletin 1383-C, 26 p.

Crandell, D.R. and Waldron, H.H.
1956: A recent volcanic mudflow of exceptional dimensions from Mt. Rainier, Washington; American Journal of Science, v. 254, p. 349-362.

Crandell, D.R., Mullineaux, D.R., and Miller, C.D.
1979: Volcanic-hazards studies in the Cascade Range of the western United States; *in* Volcanic Activity and Human Ecology, ed. P.D. Sheets and D.K. Grayson; Academic Press, New York, p. 195-219.

Croce, D., Nodari, P., Pellegrini, G.B., and Tessari, F.
1974: Effeti dell'alluvione del novembre 1966 sulle sedi abitate delle Tre Venezie; *in* Le Calamitá Naturali nelle Alpi, ed. G.B. Castiglioni; Istituto di Geographia dell'Universita di Padova, Novara, p. 291-302.

Crozier, M.J., Eyles, R.J., Marx, S.L., McConchie, J.A., and Owen, R.C.
1980: Distribution of landslips in the Wairarapa hill country; New Zealand Journal of Geology and Geophysics, v. 23, p. 575-586.

Cruden, D.M.
1976: Major rock slides in the Rockies; Canadian Geotechnical Journal, v. 13, p. 8-20.

Cruden, D.M. and Krahn, J.
1973: A reexamination of the geology of the Frank Slide; Canadian Geotechnical Journal, v. 10, p. 581-591.

Daimer, J.
1879: Die Katastrophe in den Zillertaler Alpen am 16.und 17. August 1878 und ihre Ursachen; Zeitschrift des Deutschen und Österreichischen Alpenvereins, v. 10, p. 1-16.

Dalla Torre, K.W.
1913: Tirol, Vorarlberg, und Liechtenstein; Berlin, 382 p.

Davies, T.R.H.
1982: Spreading of rock avalanche debris by mechanical fluidization; Rock Mechanics, v. 15, p. 9-24.

Day, A.L. and Allen, E.T.
1925: The volcanic activity and hot springs of Lassen Peak (California); Carnegie Institution of Washington, Publication 360, 190 p.

Decker, R. and Decker, B.
1981: Volcanoes; W.H. Freeman and Company, San Francisco, 244 p.

de Saussure, H.B.
1781: Voyages dans les Alpes; 4 volumes.

Dow, V., Kienholz, H., Plam, M., and Ives, J.D.
1981: Mountain hazards mapping: the development of a prototype combined hazards map, Monarch Lake Quadrangle, Colorado, U.S.A.; Mountain Research and Development, v. 1, p. 55-64.

Dragogna, G.
1975: Trasporto di massa e briglie selettive; International Symposium Interpraevent 1975, v. 1, p. 75-89.

Duile, J.
1834: Über Verbauung der Wildbäche in Gebirgsländern; Innsbruck, 179 p.

Ebeling, M.
1899: Der Ararat; Zeitschrift des Deutschen und Österreichischen Alpenvereins, v. 30, p. 144-163.

Eicher, W.
1977: Obwalden; *in* Hochwasserschutz in der Schweiz; Bern, p. 71-77.

Eisbacher, G.H.
1979: Cliff collapse and rock avalanches (sturzstroms) in the Mackenzie Mountains, northwestern Canada; Canadian Geotechnical Journal, v. 16, p. 309-334.
1980: Debris torrents across the Alaska Highway near Muncho Lake, northern British Columbia; *in* Current Research, Part C; Geological Survey of Canada, Paper 80-1C, p. 29-36.
1982: Slope stability and land use in mountain valleys; Geoscience Canada, v. 9, p. 14-27.

Eisbacher, G.H. and Clague, J.J.
1981: Urban landslides in the vicinity of Vancouver, British Columbia, with special reference to the December 1979 rainstorm; Canadian Geotechnical Journal, v. 18, p. 205-216.

End, G.
1922: Biasca und Val Pontirone; Jahrbuch des Schweizer Alpenclub, v. 75, p. 58-187.

Erismann, T.
1979: Mechanisms of large landslides; Rock Mechanics, v. 12, p. 15-46.

Erismann, T., Heuberger, H., and Preuss, E.
1977: Der Bimsstein von Köfels (Tirol), ein Bergsturz-Friktionit; Tschermaks Mineralogische Petrographische Mitteilungen, v. 24, p. 67-119.

Eschner, A.R. and Patric, J.H.
1982: Debris avalanches in eastern upland forests; Journal of Forestry, v. 80, p. 343-347.

Espinosa, A.F., Husid, R., and Quesada, A.
1976: Intensity distribution and source parameters from field observations; *in* The Guatemalan Earthquake of February 4, 1976, a Preliminary Report, ed. A.F. Espinosa; U.S. Geological Survey, Professional Paper 1002, p. 52-66.

Fiebiger, G.
1974: Besichtigung des Schesatobels; Wildbach und Lawinenverbauung, Sonderheft April 1974, p. 41-47.

Finch, R.H.
1930: Mud flow of Lassen volcano; Volcano Letter 266, p. 1-3.

Finsterwalder, S.
1897: Der Vernagtferner; Wissenschaftliche Ergänzungshefte zur Zeitschrift des Deutschen und Österreichischen Alpenvereins, v. 1, p. 6-13.

Fischer, K.
1967: Erdströme in den Alpen; Mitteilungen der Geographischen Gesellschaft München, v. 52, p. 231-246.

Flint, R.F.
1971: Glacial and Quaternary Geology; John Wiley and Sons, New York, 892 p.

Fliri, F.
1977: Die physiographischen Regionen des Alpenraumes; *in* Natur und Mensch im Alpenraum, ed. F. Wolkinger; Graz, p. 13-25.

Forbes, J.D.
1845: Travels Through the Alps of Savoy and Other Parts of the Pennine Chain, With Observations on the Phenomena of Glaciers; Edinburgh, 456 p.

Forcher, M.
1980: Matrei in Osttirol; Innsbruck, 408 p.

Fresacher, W.
1965: Eine Abrutschung von der Koralm im Jahre 1660; Carinthia I, v. 155, p. 209-214.

Fuganti, A.
1969: Studio geologico di sei grandi frane di roccia nella regione Trentino-Alto Adige; Museo Tridentino di Scienze Naturali, Memorie, Anno 31-32, v. 17, fasc. 3, 70 p.

Fumagalli, E. and Camponuovo, G.F.
1975: Verfica su modello di opere protettive contra la caduta di masse lapidee; International Symposium Interpraevent 1975, v. 1, p. 33-48.

Fuxjäger, B.
1975: Analyse der Wildbachkatastrophen in Osttirol; Österreichische Wasserwirtschaft, v. 27, p. 175-185.

Gardner, J.S.
1980: Frequency, magnitude, and spatial distribution of mountain rock falls and rockslides in the Highwood Pass area, Alberta, Canada; *in* Thresholds in Geomorphology, ed. D.R. Coates and J.D. Vitek; London, p. 269-295.

Gedney, D.S. and Weber, W.G.
1978: Design and construction of soil slopes; *in* Landslides, Analysis and Control, ed. R.L. Schuster and R.J. Krizek; National Academy of Sciences, Transportation Research Board, Special Report 176, p. 172-191.

Getaz, H.
1977: Amenagement de la Mortivue à Semsales; Hochwasserschutz in der Schweiz, Bern, p. 96-97.

Gex, F.
1931: L'éboulement du Châtelard; Revue de Géographie Alpine, v. 19, p. 499-521.

Goguel, J.
1968: Les enseignement du glissement de Pontamafrey; Bureau de Recherches Géologiques et Minières, Bulletin, série 2, sec. 2, no. 3, p. 55-62.
1978: Scale-dependent rockslide mechanisms, with emphasis on the role of pore fluid vaporization; *in* Rockslides and Avalanches, v. 1, ed. B. Voight, Elsevier Scientific Publishing Co., Amsterdam, p. 693-705.

Goguel, J. and Pachoud, A.
1972: Géologie et dynamique de l'ecroulement du Mont Granier dans le Massif de Chartreuse, en novembre 1248; Bureau de Recherches Géologiques et Minières, Bulletin, série 2, sec. 3, no. 1, p. 29-38.
1978: Les mouvements de terrain du versant sud du Massif de Platé (Haute-Savoie, France); Bureau de Recherches Géologiques et Minières, Bulletin, série 2, sec. 4, no. 3, p. 201-214.

Gorfer, A.
1975: Le valli del Trentino — Trentino occidentale; Calliano, 895 p.
1977: Le valli del Trentino — Trentino orientale; Calliano, 1118 p.

Gorshkov, G.S.
1959: Gigantic eruption of volcano Bezymianny; Bulletin Volcanologique, v. 20, p. 77-109.

Gortani, M.
1928: Il terremoto del 27 Marzo 1928 nelle Prealpi dell'Arzino (Friuli); L'Universo, v. 9, no. 12, p. 1-56.

Govi, M. and Sorzana, P.F.
1977: Effeti geologici del terremoto: frane; *in* Studio Geologico dell'Area Maggiormente Colpita dal Terremoto Friuliano del 1976, ed. B. Martinis; Consiglio Nazionale delle Ricerche, Milano, p. 329-368.

Gray, D.H. and Leiser, A.T.
1982: Biotechnical slope protection and erosion control; Van Nostrand Reinhold Co., New York, 271 p.

Grunder, M.
1980: Beispiel einer anwendungsorientierten Gefahrenkartierung 1:25 000 für forstliche Sanierungsprojekte im Berner Oberland (Schweiz); International Symposium Interpraevent 1980, v. 4, p. 353-360.

Guichonnet, P.
1950: La catastrophe du glacier du Tour; Revue de Géographie Alpine, v. 38, p. 198-201.

Guidicini, G. and Iwasa, O.Y.
1977: Tentative correlation between rainfall and landslides in a humid tropical environment; International Association of Engineering Geology, Bulletin, v. 16, p. 13-20.

Guillomin, A.
1937: Les Abîmes de Mayans (Savoie); Revue de Géographie Alpine, v. 25, p. 583-617.

Gümbel, W.V.
1861: Geognostische Beschreibung des Bayrischen Alpengebietes; Gotha, 730 p.

Guntschl, E.
1966: Die Hochwasserkatastrophe im Jahre 1965; Österreichische Wasserwirtschaft, v. 18, Heft 3/4, p. 62-71.

Gwinner, M.P.
1971: Geologie der Alpen; Stuttgart, 477 p.

Hadley, J.B.
1964: Landslides and related phenomena accompanying the Hebgen Lake earthquake of August 17, 1959; U.S. Geological Survey, Professional Paper 435, p. 107-138.

Haefeli, R.
1967: Zum progressiven Bruch in Schnee, Bøden Fels und Eis; Felsmechanik und Ingenieurgeologie, v. 5, p. 17-37.

Hampel, R.
1970: Die Grundlagen der Wildbachverbauung; Wildbach- und Lawinen-verbau, v. 34, no. 2, p. 69-139.

Hampton, M.A.
1979: Buoyancy in debris flows; Journal of Sedimentary Petrology, v. 49, p. 753-758.

Hanausek, E.
1975: Wildbach und Lawinenschutz in Tirol; *in* Hochwasser und Lawinenschutz in Tirol, ed. H. Aulitzky; Innsbruck, p. 108-128.

Harden, C.P.
1976: Landslides near Aspen, Colorado; University of Colorado, Institute of Arctic and Alpine Research, Occasional Paper 20, 60 p.

Harp, E.L., Wilson, R.C., and Wieczorek, G.F.
1981: Landslides from the February 4, 1976, Guatemala earthquake; U.S. Geological Survey, Professional Paper 1204-A, 35 p.

Harris, S.L.
1976: Fire and Ice, the Cascade Volcanoes; The Mountaineers, Seattle, 320 p.

Hartwagner, H.
1954: Die Schmittenbach-Aufforstung bei Zell am See; Allgemeine Forstzeitung, v. 65, no. 23/24, p. 21-23.

Hauswirth, E.K., Lahodnysky, R., Roch, K.H., and Scheidegger, A.E.
1982: Geophysikalische Untersuchungen an der Grosshangbewegung Wörschachwald (Ennstal, Steiermark); Mitteilungen Naturwissen-schaftlicher Verein Steiermark, v. 112, p. 75-90.

Hauswirth, E.K., Pirkl, H., Roch, K.H., and Scheidegger, A.E.
1979: Untersuchungen eines Talzuschubes bei Lesach (Kals, Osttirol); Verhandlungen der Geologischen Bundesanstalt, v. 1979, p. 51-76.

Heede, B.H.
1980: Gully erosion — a soil failure; possibilities and limits of control; International Symposium Interpraevent 1980, v. 1, p. 317-330.

Heilprin, A.
1903: Mont Pelée and the tragedy of Martinique; J.B. Lippincott, Philadelphia.

Heim, A.
1882a: Der Bergsturz von Elm; Deutsche Geologische Gesellschaft, Zeitschrift, v. 34, p. 74-115.
1882b: Über Bergstürze; Neujahrsblatt der Naturforschenden Gesellschaft Zürich, v. 84, 31 p.
1895: Die Gletscherkatastrophe an der Altels am 11 September 1895; Neujahrsblatt der Naturforschenden Gesellschaft in Zürich, v. 98, 63 p.
1921: Geologie der Schweiz; Leipzig, 2 volumes.
1932: Bergsturz und Menschenleben; Beiblatt zur Vierteljahrsschrift der Naturforschenden Gesellschaft in Zürich, v. 77, p. 1-217.

Helbing, R.
1940: Ausbruch eines Gletschersees in den argentinischen Anden und aussergewöhnliche Gletscherschwankungen im Allgemeinen; Schweizerische Bauzeitung, v. 115, no. 11, p. 1-8.

Hellwald, F.
1869: Die Elementarereignisse in den Alpen im Herbste 1968; Jahrbuch des Österreichischen Alpenvereins, v. 5, p. 250-270.

Heuberger, H.
1975: Das Ötztal; Innsbrucker Geographische Schriften, v. 2, p. 213-249.

Hirsbrunner, G.
1960: Beiträge zur Morphologie und Hydrologie der Rovanataler; Beiträge zur Geologie der Schweiz, Hydrologie 11, 79 p.

Hoffmann, F.
1972: Die Hochwasserkatastrophen im Oberpinzgau in den Jahren 1970 und 1971; Wildbach und Lawinenverbau, v. 36, Sonderheft Marz, p. 4-24.

Hoinkes, H.
1969: Surges of the Vernagtferner in the Ötztal Alps since 1599; Canadian Journal of Earth Sciences, v. 6, p. 853-862.
1972: Die Ausbrüche (surges) des Kola-Gletschers in Nord-Ossetien, zentraler Kaukasus; Zeitschrift für Gletscherkunde und Glazialgeologie, v. 8, p. 253-270.

Hölzl, F.
1975: 1200 Jahre Zell am See; Salzburg, 423 p.

Hopf, J. and Wanner, J.
1975: Probleme eines murstossfähigen Baches im Inntal: Enterbach Gemeinde Inzing (Oberlauf- und Unterlaufregulierung); *in* Hochwasser- und Lawinenschutz in Tirol, ed. H. Aulitzky; Innsbruck, p. 108-128.

Horninger, G. and Weiss, E.H.
1980: Engineering geology in mountainous regions; International Geological Congress, 26th, Paris, Excursion 132C; Abhandlungen der Geologischen Bundesanstalt, v. 34, p. 257-286.

Hovey, E.O.
1902: Martinique and St. Vincent: a preliminary report upon the eruption of 1902; American Museum of Natural History, Bulletin, v. 16, p. 333-373.

Hsü, K.
1975: Catastrophic debris streams (sturzstroms) generated by rockfalls; Geological Society of America, Bulletin, v. 86, p. 129-140.
1978: Albert Heim: Observations on landslides and relevance to modern interpretations; *in* Rockslides and Avalanches, v. 1, ed. B. Voight; Elsevier Scientific Publishing Co., Amsterdam, p. 71-96.

Huber, A.
1980: Felssturze in Seen und hierdurch erzeugte Schwallwellen — Erkenntnisse über die Entstehung und Ausbreitung dieser Wellen; International Symposium Interpraevent 1980, v. 1, p. 57-68.

Huder, J.
1976: Creep in Bündner Schist; Laurits Bjerrum Memorial Volume, Norges Geotechnisk Institut, Oslo, p. 125-153.

Hueber, A.
1906: Die gefährlichsten Gletscher von Tirol; Program der K.K. Oberrealschule Innsbruck, 40 p.

Humbert, M.
1977: La cartographie ZERMOS, modalites d'etablissement des cartes exposees a des risques lies aux mouvements du sol et du sous-sol; Bureau de Recherches Géologiques et Minières, Bulletin, série 2, sec. 3, no. 1/2, p. 5-8.

Hyde, J.H. and Crandell, D.R.
1978: Postglacial volcanic deposits at Mount Baker, Washington, and potential hazards from future eruptions; U.S. Geological Survey, Professional Paper 1022-C, 17 p.

Ikeya, H.
1976: Introduction to Sabo works; The Japan Sabo Association, Tokyo, 168 p.

Imrie, A.S. and Patton, F.D.
1977: Dams and associated landslides; Association of Engineering Geologists, 1977 National Meeting, Guidebook to Field Trips, Seattle, 133 p.

Iwamotu, M. and Hirano, M.
1982: Mechanical characteristics of debris flow; Mitteilungen der Forstlichen Bundesanstalt Wien, v. 144, p. 151-161.

Jäckli, H.
1957: Gegenwartsgeologie des bündnerischen Rheingebiets; Beiträge zur Geologie der Schweiz, Geotechnische Serie, v. 36, 136 p.

Jackson, L.E.
1977: Dating and recurrence frequency of prehistoric mudflows near Big Sur, Monterey County, California; U.S. Geological Survey, Journal of Research, v. 5, p. 17-32.
1979: A catastrophic glacial outburst flood (jökulhlaup) mechanism for debris flow generation at the Spiral Tunnels, Kicking Horse River basin, British Columbia; Canadian Geotechnical Journal, v. 16, p. 806-813.

Jaeger, C.
1979: Rock Mechanics and Engineering; Cambridge University Press, Cambridge, 523 p.

Jail, M.
1975: Trois crues torrentielles en Haute-Maurienne (Savoie-France); International Symposium Interpraevent 1975, v. 1, p. 187-190.

Jail, M. and Marnezy, A.
1980: Phénomènes naturels récent et problèmes de circulation en Maurienne (Savoie); International Symposium Interpraevent 1980, v. 3, p. 249-258.

Jail, M. and Vivian, R.
1971: Les glissements de terrain et les éboulements dans les Alpes francaises du Nord en 1970; Revue de Géographie Alpine, v. 59, p. 473-502.

Janda, R.J., Scott, K.M., Nolan, K.M., and Martinson, H.A.
1981: Lahar movement, effects, and deposits; *in* The 1980 Eruptions of Mount St. Helens, Washington, ed. P.W. Lipman and D.R. Mullineaux; U.S. Geological Survey, Professional Paper 1250, p. 461-478.

Japanese Society of Landslides
1972: Landslides in Japan; Tokyo, 42 p.

Jeannet, A.
1918: Monographie géologique des Tours d'Ai; Matériaux pour la Carte Géologique de la Suisse, Nouvelle Série, no. 34, p. 690-695.

Jeglitsch, F., Jelem, H., Kilian, W., Kronfellner-Kraus, G., Neuwinger, I., Noistering, H., and Stern, R.
1975: Über die Einschätzung von Wildbächen — Der Trattenbach; Mitteilungen der Forstlichen Bundes-Versuchsanstalt Wien, v. 112, 162 p.

Jian, L. and Defu, L.
1980: The formation and characteristics of mudflow and flood in the mountain area of the Dachao Torrent and its prevention; International Symposium Interpraevent 1980, v. 2, p. 3-13.

Johnson, A.M.
1970: Physical Processes in Geology; Freeman, Cooper, and Co., San Francisco, 577 p.

Johnson, A.M. and Rahn, P.H.
1970: Mobilization of debris flow; Zeitschrift für Geomorphologie, Supplement v. 10, p. 168-186.

Johnson, K., Olson, E.A., and Manandhar, S.
1982: Environmental knowledge and response to natural hazards in mountainous Nepal; Mountain Research and Development, v. 2, p. 175-188.

Jones, F.O.
1973: Landslides of Rio de Janeiro and the Serra das Araras Escarpment, Brasil; U.S. Geological Survey, Professional Paper 697, 42 p.

Jorstad, F.
1968: Waves generated by landslides in Norwegian fjords and lakes; Norwegian Geotechnical Institut, Publication no. 79, 11 p.

Keefer, D.K. and Tannaci, N.E.
1981: Bibliography on landslides, soil liquefaction, and related ground failures in selected historic earthquakes; U.S. Geological Survey, Open-File Report 81-572, 38 p.

Kelly, M.
1980: A prehistoric catastrophic rock avalanche at Holsteinsborg, West Greenland; Geological Society of Denmark, Bulletin, v. 28, p. 73-79.

Kemmerling, G.L.L.
1921: De uitbarsting van den G. Keloet in den nacht van den 19den op den 20sten Mei 1919; Dienst Mijnwezen, Vulkanoloisch Mededeelingen, no. 2, 120 p.

Kent, P.E.
1965: The transport mechanism in catastrophic rockfalls; Journal of Geology, v. 74, p. 79-83.

Kerr, F.A.
1934: The ice dam and floods of the Talsekwe, British Columbia; Geographical Review, v. 24, p. 643-645.

Kienholz, H.
1977: Kombinierte geomorphologische Gefahrenkarte 1:10,000 von Grindelwald; Geographica Bernensia, G4, 104 p.

Kienholz, H., Hafner, H., Schneider, G., and Tamrakar, R.
1983: Mountain hazards mapping in Nepal's Middle Mountains — Maps of Land use and geomorphic damages (Kathmandu — Kakani area); Mountain Research and Development, v. 3, p. 195-220.

King, W.D.V.O.
1934: The Mendoza River flood of 10-11 January 1934, Argentina; Geographical Journal, v. 84, p. 319-326.

Kinzl, H.
1932: Die grössten nacheiszeitlichen Gletschervorstösse in den Schweizer Alpen und in der Mont Blanc Gruppe; Zeitschrift für Gletscherkunde, v. 20, p. 269-397.
1949: Die Vergletscherung in der Südhälfte der Cordillera Blanca (Peru); Zeitschrift für Gletscherkunde und Glazialgeologie, v. 1, p. 1-28.

Kjartansson, G.
1967: The Steinsholtshlaup, central-south Iceland on January 15th, 1967; Jökull, v. 17, p. 249-262.

Knoblauch, P. and Reinhard, M.
1939: Erläuterungen zum Geologischen Atlas der Schweiz, 1:25,000; Atlas Blatt 516, Jorio, 87 p.

Knoblauch, P. and Zurbuchen, M.
1927: Dei Bewegungen am Motto d'Arbino bei Bellinzona; Die Alpen, v. 3, p. 361-373.

Koch, A.J. and McLean, H.
1975: Pleistocene tephra and ash flow deposits in the volcanic highlands of Guatemala; Geological Society of America, Bulletin, v. 86, p. 529-541.

Koch, G.A.
1875: Über Mürbrüche in Tirol; Jahrbuch der Geologischen Reichsanstalt, v. 25, p. 97-128.
1883: Die Ursachen der Hochwasser Katastrophe in den Südalpen; Zeitschrift des Deutschen und Österreichischen Alpenvereins, v. 14, p. 136-169.
1892: Ein kalbender Gletscher in den Ostalpen; Mitteilungen der Kaiserlich-Königlichen Geographischen Gesellschaft Wien, v. 35, no. 5, p. 176-193.

Kopp, J.
1936: Die Bergsturze des Rossberges; Eclogae Geologicae Helvetiae, v. 29, p. 490-493.

Korner, H.J.
1976: Reichweite und Geschwindigkeit von Bergstürzen und Fliesslawinen; Rock Mechanics, v. 8, p. 225-256.

Kravogel, A. and Wurzer, E.
1967: Die Hochwasserschaden 1966 in den Katastrophengebieten; Österreichische Wasserwirtschaft, v. 19, p. 41-45.

Kronfellner-Kraus, G.
1974: Die Wildbacherosion im allgemeinen und der Talzuschub im besonderen; *in* 100 Jahre Forstliche Bundesanstalt, p. 309-342.
1975: Zur Geschiebebilanzierung; International Symposium Interpraevent 1975, v. 1, p. 49-61.
1980: Neue Untersuchungen in Wildbächen — Der Talzuschub in Abhängigkeit von Niederschlägen; International Symposium Interpraevent 1980, v. 1, p. 179-192.
1982: Über den Geschiebe- und Feststofftransport in Wildbächen; Österreichische Wasserwirtschaft, v. 34, p. 12-21.

Lacroix, A.
1904: La Montagne Pelée et Ses Eruptions; Masson et Cie, Paris.

Largaiolli, T. and Siro, L.
1977: Studio geologico e geotecnico dell'area franosa a nord di Tesino (Trentino); Studi Trentini di Scienze Naturali, v. 54, p. 3-16.

Largaiolli, T. and Vaia, F.
1975: Indagini sulla situazione strutturale del versante roccioso sovrastante Mezzocorona (Trentino); Studi Trentini di Scienze Naturali, v. 52, p. 115-138.

Lauscher, F.
1973: Unwetter im Lande Salzburg im Zeitraum 1946-1970; *in* Beiträge zur Klimatologie, Meteorologie und Klimamorphologie; Geographisches Institut, Universität Salzburg, p. 143-167.

Lehmann, F.W.P.
1879: Die Wildbäche der Alpen; Breslau, 107 p.

Lehmann, O.
1926: Die Verheerungen in der Sandlinggruppe; Denkschrift der Akademie der Wissenschaften in Wien, mathematisch-naturwissenschaftliche Klasse, v. 100, p. 259-299.
1942: Über Böschungswinkel und Böschungshöhen im Hinblick auf den Bergstruz von Goldau; Eclogae Geologicae Helvetiae, v. 35, p. 55-65.

Lemoine, M.
1978: Geological Atlas of Alpine Europe and Adjoining Alpine Areas; Elsevier Scientific Publishing Co., Amsterdam, 584 p.

LeRoy Ladurie, M.
1971: Times of Feast, Times of Famine. A History of Climate since the Year 1000; Doubleday, London.

Leys, E.
1977: Murenchronik des Ötztales; Wildbach und Lawinenverbauung, Sonderheft November 1977, p. 63-79.
1980: Die Grenzen der Machbarkeit in Gebirgslandschaften; International Symposium Interpraevent 1980, v. 2, p. 265-277.

Lichtenhahn, C.
1971a: Zwei Betonmauern: Die Geschieberückhaltesperre am Illgraben (Wallis); International Symposium Interpraevent 1971, v. 3, p. 451-456.
1971b: Zwei Stollenbauten; International Symposium Interpraevent 1971, v. 3, p. 465-475.
1979: Die Verbauung des Fallbaches in der Gemeinde Saas-Balen (Wallis); Mitteilungen der Versuchsanstalt fur Wasserbau, Hydrologie und Glaziologie an der ETH Zurich, no. 41, p. 169-176.

Liestol, O.
1956: Glacier-dammed lakes in Norway; Norsk Geografisk Tidsskrift, v. 15, p. 122-149.

Lipman, P.W. and Mullineaux, D.R. (editors)
1981: The 1980 eruptions of Mount St. Helens, Washington; U.S. Geological Survey, Professional Paper 1250, 844 p.

Liss, C.C.
1970: Der Morenogletscher in der Patagonischen Kordillere; Zeitschrift für Gletscherkunde und Glazialgeologie, v. 6, p. 161-180.

Lliboutry, L., Morales Arnao, B., Pautre, A., and Schneider, B.
1977: Glaciological problems set by the control of dangerous lakes in Cordillera Blanca, Peru. I. Historical failures of morainic dams, their causes and prevention; Journal of Glaciology, v. 18, p. 239-254.

Lombard, R.E., Miles, M.G., Nelson, L.M., Kresch, D.L., and Carpenter, P.J.
1981: Channel conditions in lower Toutle and Cowlitz rivers resulting from the mudflows of May 18, 1980; U.S. Geological Survey, Circular 850-C, 16 p.

Lotze, F.
1960: Notizen zur Aktuogeologie; Neues Jahrbuch für Geologie und Paläontologie, Monatshefte, v. 1960, p. 182-190.

Lucchitta, B.K.
1979: Landslides in Valles Marineris, Mars; Journal of Geophysical Research, v. 84, p. 8097-8113.

Macdonald, G.A.
1972: Volcanoes; Prentice-Hall, Englewood Cliffs, 510 p.

Mahr, T.
1977: Deep-reaching gravitational deformations of high mountain slopes; International Association of Engineering Geology, Bulletin, no. 16, p. 121-127.

Malatrait, M.
1975: Analyse et classement des mouvements gravitaire, Feuille St. Jean-de-Maurienne; Thése troisième cycle, Université de Grenoble, 219 p.

Mariétan, I.
1927: Éboulement de la Cime de l'Est et coulées du St. Barthélemy; Société Vaudoise des Sciences Naturelles, Bulletin, v. 56, p. 249-262.

Marinelli, G.
1878: L'Antelao; Boletino del Club Alpino Italiano, 1878, p. 26-42.

Martin, J.J.E. and Messines du Sourbier, J.
1936: L'éboulement de Serrières-en-Chautagne; Revue de Géographie Alpine, v. 24, p. 239-245.

Martinis, B. and Salvini, G.
1972: Dissesti idrogeologici nell'Alta Valle del Tagliamento (Friuli); Milan Universita, Istituto di Geologia e Paleontologia, Pubblicazione, no. 101, 54 p.

Mason, K.
1929: Indus flood and Shyok glaciers; Himalayan Journal, v. 1, p. 11-29.
1935: The study of threatening glaciers; Geographical Journal, v. 85, p. 24-41.

Mathews, W.H.
1965: Two self-dumping ice dammed lakes in British Columbia; Geographical Review, v. 55, p. 46-52.

Mathews, W.H. and McTaggart, K.C.
1978: Hope rockslides, British Columbia; *in* Rockslides and Avalanches, v. 1, ed. B. Voight; Elsevier Scientific Publishing Co., Amsterdam, p. 259-275.

Matthes, F.E.
1939: Report of Committee on Glaciers, April 1939; American Geophysical Union, Transactions, v. 20, pt. 4, p. 518-523.

Mayr, F.
1895: Die Gletscherausbrüche im Martellthale und der Klausenbau am Zufallboden; Österreichische Monatsschrift für den öffentlichen Baudienst, Heft 1, 1895, 30 p.

McLellan, P.J.A.
1983: Investigation of some rock avalanches in the Mackenzie Mountains; unpublished M.Sc. thesis, University of Alberta, 281 p.

McSaveney, M.J.
1978: Sherman Glacier rock avalanche, Alaska, U.S.A.; *in* Rockslides and Avalanches, v. 1, ed. B. Voight; Elsevier Scientific Publishing Co., Amsterdam, p. 197-258.

Mears, A.I.
1977: Debris-flow hazard analysis and mitigation; Colorado Geological Survey, Information Series, no. 8, 45 p.

Megahan, W.F.
1977: Reducing erosional impacts of roads; FAO Conservation Guide, Guidelines for Watershed Management, Rome, p. 237-261.

Meneroud, J.P. and Calvino, A.
1976: Carte ZERMOS, Région de la Moyenne Vésubie (Alpes Maritimes); Bureau de Recherches Géologiques et Minières, 11 p.

Messerli, B., Zumbühl, H.J., Ammann, K., Kienholz, H., Oeschger, H., Pfister, C., and Zurbuchen, M.
1975: Die Schwankungen des unteren Grindelwaldgletschers seit dem Mittelalter; Zeitschrift für Gletscherkunde und Glazialgeologie, v. 11, p. 3-110.

Miller, C.D., Mullineaux, D.R., and Crandell, D.R.
1981: Hazards assessment at Mount St. Helens; *in* The 1980 Eruptions of Mount St. Helens, Washington, ed. P.W. Lipman and D.R. Mullineaux; U.S. Geological Survey, Professional Paper 1250, p. 789-802.

Miller, C.D., Mullineaux, D.R., and Hall, M.L.
1978: Reconnaissance map of potential volcanic hazards from Cotopaxi volcano, Ecuador; U.S. Geological Survey, Map I-1072.

Mokievsky-Zubok, O.
1977: Glacier-caused slide near Pylon Peak, British Columbia; Canadian Journal of Earth Sciences, v. 14, p. 2657-2662.

Moll, L.
1976: Zur Geologie des Gridlontobels bei Pettneu am Arlberg und seines Einzugsgebietes, unter besonderer Berücksichtigung der Massenbewegungen; Ph.D. thesis, University of Innsbruck, 94 p.

Montandon, F.
1933: Chronologie des grands éboulements alpins, du début de l'ére chrétienne a nos jours; Société de Géographie Genève, Matériaux pour l'Étude des Calamités, no. 32, p. 271-340.

Moore, D.P. and Mathews, W.H.
1978: The Rubble Creek landslide, southwestern British Columbia; Canadian Journal of Earth Sciences, v. 15, p. 1039-1052.

Moore, J.G. and Melson, W.C.
1969: Nuées ardentes of the 1968 eruption of Mayon volcano, Philippines; Bulletin Volcanologique, v. 33, p. 600-620.

Morales Arnao, B.
1966: The Huascaran avalanche in the Santa Valley, Peru; International Association of Scientific Hydrology, Publication no. 69, p. 304-315.

Moret, L.
1943: Les conditions géologiques du glissement de terrain de Béllevaux du Haute Savoie (Mars 1943); Revue de Géographie Alpine, v. 31, p. 423-428.

Moser, M.
1973: Analyse der Ausbruchsbildung bei den Hochwasserkatastrophen der Jahre 1965 und 1966 im mittleren Lesachtal (Karnten); Carinthia II, v. 83, p. 179-234.

Moser, M. (cont.)
1980: Zur Analyse von Hangbewegungen in schwachbindigen bis rolligen Lockergesteinen im alpinen Raum anlässlich von Starkniederschlägen; International Symposium Interpraevent 1980, v. 1, p. 121-148.

Moser, M. and Glumac, S.
1983: Geotechnische Untersuchungen zum Massenkriechen in Fels am Beispiel des Talzuschubes Gradenbach (Karnten); Verhandlungen der Geologischen Bundesanstalt, v. 1982, p. 203-241.

Mougin, P.
1914: Les Torrents de la Savoie; Grenoble, 1250 p.
1931: La Restauration des Alpes; Paris, 582 p.

Müller, L.
1963: Der Felsbau; Stuttgart, 624 p.
1968: New considerations of the Vaiont Slide; Rock Mechanics and Engineering Geology, v. 6, p. 1-91.

Mullineaux, D.R. and Peterson, D.W.
1974: Volcanic hazards on the Island of Hawaii; U.S. Geological Survey, Open-File Report 74-239.

Murai, I.
1963: A brief note on the eruption of the Tokachi-dake volcano of June 29 and 30, 1962; Tokyo University Earthquake Research Bulletin, v. 41, p. 185-208.

Murata, K.J., Dondoli, C., and Saenz, R.
1966: The 1963-65 eruption of Irazú volcano, Costa Rica (the period of March 1963 to October 1964); Bulletin Volcanologique, v. 29, p. 765-796.

Mussio, G.
1974: Calamitá naturali ricorrenti nel Triangolo Lariano in una prospettiva geographica; *in* Le Calamitá Naturali nelle Alpi, ed. G.B. Castiglioni; Istituto di Geographia dell'Universitá di Padova, Novara, p. 73-84.

Nasmith, H., Mathews, W.H., and Rouse, G.E.
1967: Bridge River ash and some other recent ash beds in British Columbia; Canadian Journal of Earth Sciences, v. 4, p. 163-170.

Neiger, H.
1978: Bergstürze im Haslital; Meiringen, 41 p.

Nemčok, A.
1972: Gravitational slope deformation in high mountains; International Geological Congress, 24th, Montreal, Proceedings, sec. 13, p. 132-141.

Neuhauser, E. and Schober, W.
1970: Das Kriechen der Talgehänge und elastische Hebungen beim Speicher Gepatsch; 2nd Congress of Rock Mechanics, Belgrade, Proceedings v. 3, p. 447-458.

Neumann van Padang, M.
1933: De uitbarsting van den Merapi (Midden Java) in den jaren 1930-1931; Dienst van den Mijnbouw in Nederlandsch-Indie, Vulkanologische en Seismologische Mededeelingen, no. 12, 135 p.
1960: Measures taken by the authorities of the Volcanological Survey to safeguard the population from the consequences of volcanic outbursts; Bulletin Volcanologique, v. 23, p. 181-192.

Niederer, J.
1941: Der Bergsturz am Flimserstein, Fidaz, am 10. April 1939; Jahrbuch der Naturforschenden Gesellschaft Graubündens, v. 77, p. 3-27.

Nijasov, B.
1975: Zu Problemen der Murenbekämpfung in Kasachstan; International Symposium Interpraevent 1975, v. 2, p. 225-228.

Nilsen, T.H.
1982: Alluvial fan deposits; *in* Sandstone Depositional Environments, ed. P.A. Scholle and D. Spearing; American Association of Petroleum Geologists, Memoir 31, p. 49-86.

Nilsen, T.H., Wright, R.H., Vlasic, T.C., and Spangle, W.E.
1979: Relative slope stability and land-use planning in the San Francisco Bay Region, California; U.S. Geological Survey, Professional Paper 944, 96 p.

Nussbaum, F.
1957: Über rezente Erdrutsche und Felsstürze in der Schweiz; Geographica Helvetica, v. 12, p. 216-221.

Oberhauser, R.
1980: Der geologische Aufbau Österreichs; Springer-Verlag, Vienna, 700 p.

Oberholzer, J.
1900: Monographie einiger prähistorischer Bergstürze in den Glarneralpen; Beiträge zur geologischen Karte der Schweiz, Neue Folge 9, 209 p.
1933: Geologie der Glarneralpen; Beiträge zur geologischen Karte der Schweiz, Neue Folge 28, 626 p.

Okuda, S., Suwa, H., Okunishi, K., Yokoyama, K., and Nakano, M.
1980: Observations on the motion of a debris flow and its geomorphological effects; Zeitschrift für Geomorphologie, Supplement, v. 35, p. 142-163.

O'Shea, B.E.
1954: Ruapehu and the Tangiwai disaster; New Zealand Journal of Science and Technology, ser. B, v. 36, p. 174-189.

Pachoud, A.
1976: Notice explicative de la carte des zones exposées a des risques liés aux mouvements du sol et du sous-sol a 1/20,000, Platé, region Passy-Servoz; Bureau de Recherches Géologiques et Minières, 21 p.
1979: Notice explicative de la carte des zones exposées a des risques liés aux mouvements du sol et du sous-sols, région Bourg-St.-Maurice (Savoie); Bureau de Recherches Géologiques et Minières, Orléans, 16 p.

Palluel-Guillard, A.
1978: Estampes de Tarentaise; Chambery, 94 p.

Pardyanto, L., Reksowirogo, L.D., Mitrohartono, F.X.S., and Hardjowarsito, S.H.
1978: Volcanic hazard map, Merapi volcano, central Java; Geological Survey of Indonesia, Bandung.

Pariseau, W.G. and Voight, B.
1979: Rockslides and avalanches: basic principles and perspectives in the realm of civil and mining operations; *in* Rockslides and Avalanches, v. 2, ed. B. Voight; Elsevier Scientific Publishing Co., Amsterdam, p. 1-92.

Patzelt, G.
1973: Die neuzeitlichen Gletscherschwankungen in der Venedigergruppe (Hohe Tauern, Ostalpen); Zeitschrift für Gletscherkunde, v. 9, p. 5-57.

Pavoni, N.
1977: Erdbeben im Gebiet der Schweiz; Eclogae Geologicae Helvetiae, v. 70, p. 351-370.

Pictet, P.
1818: Account of the formation of the Lake of Mauvoisin by the descent of glacier and of the inundations of the Val de Bagnes in 1595 and 1818; Edinburgh Philosophical Journal, v. 1, p. 187-191.

Pierson, T.C.
1983: Soil pipes and slope stability; Quarterly Journal of Engineering Geology, London, v. 16, p. 1-11.

Pippan, T.
1981: Studies on present day geomorphological processes of high morphogenetic intensity in the High Calcareous Alps of Salzburg; Zeitschrift für Geomorphologie, Supplement, v. 37, p. 73-90.

Piteau, D.R. and Peckover, F.L.
1978: Engineering of rock slopes; *in* Landslides — Analysis and Control, ed. R.L. Schuster and R.J. Krizek; National Academy of Sciences, Transportation Research Board, Special Report 176, p. 192-228.

Piteau, D.R., Mylrea, F.H., and Blown, I.G.
1978: Downie Slide, Columbia River, British Columbia, Canada; *in* Rockslides and Avalanches, v. 1, ed. B. Voight; Elsevier Scientific Publishing Co., Amsterdam, p. 365-392.

Pitra, F.
1921: Zur Katastrophe vom 9.August 1921; Der Schlern, v. 2, p. 399-401.

Plafker, G. and Erikson, G.E.
1978: Nevados Huascarán avalanches, Peru; *in* Rockslides and Avalanches, v. 1, ed. B. Voight; Elsevier Scientific Publishing Co., Amsterdam, p. 277-314.

Pollack, V.
1892: Der Bergsturz im 'Grossen Tobel' nächst Langen am Arlberg am 9.Juli 1892; Jahrbuch der Geologischen Reichsanstalt, v. 1892, p. 661-671.

Pomeroy, J.S.
1980: Storm-induced debris avalanching and related phenomena in the Johnstown area, Pennsylvania, with references to other studies in the Appalachians; U.S. Geological Survey, Professional Paper 1191, 24 p.

Porter, S.C. and Denton, G.H.
1967: Chronology of Neoglaciation in the North American Cordillera; American Journal of Science, v. 265, p. 177-210.

Porter, S.C. and Orombelli, G.
1981: Alpine rockfall hazards; American Scientist, v. 69, p. 67-75.

Post, A.
1967: Effects of the March 1964 Alaska earthquake on glaciers; U.S. Geological Survey, Professional Paper 544-D, p. D1-D42.

Post, A. and Mayo, L.R.
1976: Glacier dammed lakes and outburst floods in Alaska; *in* Focus on Environmental Geology, ed. R.W. Tank; Oxford University Press, London, p. 205-224.

Preiswerk, H.
1918: Geologische Beschreibung der Lepontinischen Alpen; Beiträge zur Geologischen Karte der Schweiz, v. 26, pt. 2, 80 p.

Querini, R.
1980: Nouve norme per la correzione dei torrenti ed il consolidamento dei versanti franosi nelle zone di sismicita nelle Alpi Sud-Orientali; International Symposium Interpraevent 1980, v. 1, p. 351-359.

Rabot, C.
1905: Glacial reservoirs and their outbursts; Geographical Journal, v. 25, p. 534-548.
1910: Le glacier de Lepenáz de 1818 a 1847; Zeitschrift für Gletscherkunde, v. 5, p. 133-142.

Radbruch-Hall, D.H.
1978: Gravitational creep of rock masses on slopes; *in* Rockslides and Avalanches, v. 1, ed. B. Voight; Elsevier Scientific Publishing Co., Amsterdam, p. 607-657.

Radbruch-Hall, D.H., Varnes, D.J., and Savage, W.Z.
1976: Gravitational spreading of steep-sided ridges ('sackung') in western United States; International Association of Engineering Geology, Bulletin no. 14, p. 23-35.

Rapp, A. and Stromquist, L.
1976: Slope erosion due to extreme rainfall in the Skandinavian mountains; Geografiska Annaler, v. 58A, p. 195-200.

Rice, R.M.
1977: Forest management to minimize landslide risk; FAO Conservation Guide, Guidelines for Watershed Management, Rome, p. 271-287.

Richter, E.
1885: Der Bergsturz an der Bocca di Brenta; Mitteilungen des Deutschen und Österreichischen Alpen Vereins 1885, p. 72-73.
1889a: Der Gletscherausbruch im Martelltal und seine Wiederkehr; Mitteilungen des Deutschen und Österreichischen Alpen Vereins 1889, p. 231-233.
1889b: Die Hilfsmittel gegen Ausbrüche von Eisseen; Mitteilungen des Deutschen und Österreichischen Alpen Vereins 1889, p. 291-295.
1891: Geschichte der Schwankungen der Alpengletscher; Zeitschrift des Deutschen und Österreichischen Alpenvereins, v. 22, p. 1-74.
1892: Urkunden über die Ausbrüche des Vernagt- und Gurglergletschers im 17. und 18.Jahrhundert; Forschungen zur deutschen Landes- und Volkskunde, v. 6, p. 345-440.

Richter, H.
1959: Die Wildbachkatastrophen des Jahres 1958 in Steiermark und Kärnten; Österreichische Wasserwirtschaft, v. 10, p. 241-247.

Ringgenberg, F.
1975: Chronik der Oberhasli und Meiringer Lokalgeschichte; Meiringen, 58 p.

Robinson, C.S. and Lee, F.T.
1972: Geological, geophysical, and engineering investigations of the Loveland Basin Landslide, Clear Creek County, Colorado, 1963-1965; U.S. Geological Survey, Professional Paper 637, 43 p.

Rogers, N.W. and Selby, M.J.
1980: Mechanisms of shallow translational landsliding during summer rainstorms: North Island, New Zealand; Geografiska Annaler, v. 62, p. 11-21.

Röthlisberger, H.
1974: Möglichkeiten und Grenzen der Gletscherüberwachung; Neue Züricher Zeitung, 29.April 1974, Nr. 196, 15 p.
1977: Ice avalanches; Journal of Glaciology, v. 19, p. 669-671.
1979: Glaziologische Arbeiten im Zusammenhang mit den Seeausbrüchen am Grubengletscher, Gemeinde Saas-Balen (Wallis); Mitteilungen der Versuchsanstalt für Wasserbau, Hydrologie und Glaziologie an der ETH Zürich, no. 41, p. 233-256.
1981: Destructive power of glaciers; *in* Swiss National Tourist Office: Switzerland and Her Glaciers; Kummerley and Frey, p. 128-165.

Röthlisberger, H. and Aellen, M.
1970: Bewegungsregistrierung an der Zunge des Gietrogletschers; Schweizerische Bauzeitung, v. 88, no. 43, p. 33-36.

Sacco, F.
1918: I ghiacciai italiani del gruppo del Monte Bianco; Bolletino Comitato Glaciologico Italiano, v. 1, no. 3, p. 21-102.

Salis, A.
1892 (and 1914): Die Wildbachverbauung in der Schweiz; Bern, 3 volumes.

Salis, F.V.
1868: Hochwasser im Jahre 1868; Jahrbuch des Schweizer Alpenclub, v. 5.

Sapper, K. and Termer, F.
1930: Der Ausbruch des Vulkans Santa Maria in Guatemala vom 2-4 November 1929; Zeitschrift für Vulkanologie, v. 13, p. 73-101.

Schaubach, A.
1865 to 1873: Die deutschen Alpen; Jena, 5 volumes.

Schauer, T.
1975: Die Blaikenbildung in den Alpen; Schriftenreihe Bayrisches Landesamt für Wasserwirtschaft, Heft 1, München, 80 p.

Scheidegger, A.E.
1973: On the prediction of the reach and velocity of catastrophic landslides; Rock Mechanics, v. 5, p. 231-236.

Schiechtl, H.M.
1954: Die Folgen der Entwaldung am Beispiel des Finsingtales in Nordtirol; Zentralblatt für das gesamte Forstwesen, 73, Jahrgang, Heft 1/2, p. 13-28.
1980: Bioengineering for land reclamation and conservation; University of Alberta Press, Edmonton, 404 p.

Schiermböck, O. and Geyer, J.
1974: Mustrigilbach; Wildbach und Lawinenverbau, v. 37, Sonderheft April, p. 7-16.

Schmidt, C.
1896: Der Murgang des Lammbaches bei Brienz; Himmel und Erde, v. 9, p. 49-74.

Schneebeli, W. and Rothlisberger, F.
1976: 8000 Jahre Walliser Gletschergeschichte; Verlag des Schweizer Alpen Clubs, 152 p.

Schorn, J.
1902: Die Erdbeben von Tirol und Vorarlberg; Innsbruck, 186 p.

Schreiber, H. and Zettl, H.
1967: Hydrographische Chrakteristik der Hochwasserkatastrophen im August und November 1966 in Österreich; Österreichische Wasserwirtschaft, v. 19, p. 46-55.

Schumm, S.A. and Chorley, R.J.
1964: The fall of threatening rock; American Journal of Science, v. 262, p. 1041-1054.

Schuster, R.L. and Krizek, R.J. (ed.)
1978: Landslides — Analysis and Control; National Academy of Sciences, Transportation Research Board, Special Report 176, 234 p.

Schwinner, R.
1912: Der Monte Spinale bei Campiglio und andere Bergstürze in den Sudalpen; Mitteilungen der Geologischen Gesellschaft Wien, v. 5, p. 128-197.

Scott, K.M. and Williams, R.P.
1978: Erosion and sediment yields in the Transverse Ranges, southern California; U.S. Geological Survey, Professional Paper 1030, 38 p.

Seckendorff, A.V.
1884: Verbauung der Wildbäche, Aufforstung und Berasung der Gebirgstgründe; Wien, 319 p.

Sekiya, S. and Kikuchi, Y.
1889: The eruption of Bandai-san; Imperial University of Japan, College of Science, Journal, v. 3, pt. 2, p. 91-172.

Selby, M.J.
1982: Hillslope Materials and Processes; Oxford University Press, London, 264 p.

Sharp, R.P.
1960: Glaciers; University of Oregon, Eugene, 75 p.

Sharp, R.P. and Nobles, L.H.
1953: Mudflow of 1941 at Wrightwood, southern California; Geological Society of America, Bulletin, v. 64, p. 547-560.

Shreve, R.L.
1968a: The Blackhawk landslide; Geological Society of America, Special Paper 108, 47 p.
1968b: Leakage and fluidization in air-layer lubricated avalanches; Geological Society of America, Bulletin, v. 79, p. 653-658.

Sidle, R.C. and Swanston, D.N.
1982: Analysis of a small debris slide in coastal Alaska; Canadian Geotechnical Journal, v. 19, p. 167-174.

Simkin, T., Siebert, L., McClelland, L., Bridge, D., Newhall, C., and Latter, J.H.
1981: Volcanoes of the World; Smithsonian Institution, Hutchinson Ross Publishing Co., Stroudsburg, 232 p.

Slingerland, R.L. and Voight, B.
1979: Occurrences, properties, and predictive models of landslide-generated water waves; *in* Rockslides and Avalanches, v. 2, ed. B. Voight; Elsevier Scientific Publishing Co., Amsterdam, p. 317-400.

Smith, T.C. and Hart, E.W.
1982: Landslides and related storm damage, January 1982, San Francisco Bay region; California Geology, p. 139-157.

Solonenko, V.P.
1977: Landslides and collapses in seismic zones and their prediction; International Association of Engineering Geology, Bulletin, v. 15, p. 4-8.

Sommer, N.
1980: Untersuchungen über Geschiebe- und Schwebstoffführung und den Transport von gelösten Stoffen in Gebirgsbächen; International Symposium Interpraevent 1980, v. 2, p. 69-94.

Sonklar, K.
1883: Von den Überschwemmungen; Vienna, 151 p.

Souther, J.G.
1980: Geothermal reconnaissance in the central Garibaldi belt, British Columbia; *in* Current Research, Part A; Geological Survey of Canada, Paper 80-1A, p. 1-11.

Stacul, P.
1979: Wildbachverbauung in Südtirol, gestern und heute; Autonome Provinz Bozen-Südtirol, 116 p.

Staffler, J.J.
1846: Tirol und Vorarlberg, statistisch und topographisch; Innsbruck, 4 volumes.

Starkel, L.
1972: The role of catastrophic rainfall in the shaping of the relief of the lower Himalaya (Darjeeling Hills); Geographica Polonica, v. 21, p. 103-147.
1976: The role of extreme (catastrophic) meteorological events in contemporary evolution of slopes; *in* Geomorphology and Climate, ed. E. Derbyshire; London, p. 203-246.

Stini, J.
1910: Die Muren; Innsbruck, 139 p.
1931: Die geologischen Grundlagen der Verbauung der Geschiebeherde in Gewässern; Wien, 121 p.
1938: Über die Regelmässigkeit der Wiederkehr von Rutschungen, Bergstürzen, und Hochwasserschäden in Österreich; Geologie und Bauwesen, v. 10, p. 9-31 and 33-48.
1941: Unsere Taler wachsen zu; Geologie und Bauwesen, v. 13, p. 71-79.
1942: Talzuschub und Bauwesen; Die Bautechnik, v. 1942, p. 80-81.

Stone, K.H.
1963: The annual emptying of Lake George, Alaska; Arctic, v. 16, p. 26-40.

Stotter, M.
1846: Die Gletscher des Vernagttales in Tirol und ihre Geschichte; Innsbruck, 75 p.

Stragiotti, L. and Peretti, L.
1953: Osservazione di coni di ghiacciao sulla frane della Becca di Luseney (Vall d'Aosta); Bolletino del Comitato Glaciologico Italiano, ser. 2, no. 4, p. 213-219.

Strele, G.
1899: Die Wildbäche und ihre Verbauung; Zeitschrift des Deutschen und Österreichischen Alpenvereins, v. 30, p. 110-132.
1932: Die Quellen der Geschiebeführung; Geologie und Bauwesen, v. 4, p. 119-145.

Strele, G. (cont.)
1936: Chronologie des grands éboulements alpins, Premier Supplement, Société de Géographie Genève, Matériaux pour l'Étude des Calamités, no. 38, p. 121-137.
1950: Grundriss der Wildbach und Lawinenverbauung; Wien, 279 p.

Sturm, M. and Matter, A.
1978: Turbidites and varves in Lake Brienz (Switzerland): deposition of clastic detritus by density currents; *in* Modern and Ancient Lake Sediments, ed. A. Matter and M.E. Tucker; International Association of Sedimentologists, Special Publication 2, p. 147-168.

Surell, A.
1841: Étude sur les Torrents des Hautes-Alpes; Paris, 161 p.

Swanson, F.J. and Dyrness, C.T.
1975: Impact of clearcutting and road construction on soil erosion by landslides in the western Cascade Range, Oregon; Geology, v. 3, p. 393-396.

Swanston, D.N.
1969: Mass wasting in coastal Alaska; U.S. Department of Agriculture, Forest Service Research Paper PNW-83, 15 p.
1970: Mechanics of debris avalanching in shallow till soils of southeast Alaska; U.S. Department of Agriculture, Forest Service Research Paper PNW-103, 17 p.

Swanston, D.N. and Swanson, F.J.
1976: Timber harvesting, mass erosion, and steepland forest geomorphology in the Pacific Northwest; *in* Geomorphology and Engineering, ed. D.R. Coates; Dowden, Hutchinson, and Ross, Stroudsburg, p. 199-221.

Takahashi, T.
1981: Debris flow; Annual Review of Fluid Mechanics, v. 13, p. 57-77.

Taylor, G.A.
1958: The 1951 eruption of Mount Lamington, Papua; Bureau of Mineral Resources, Geology and Geophysics (Australia), Bulletin, no. 38, 117 p.

Temple, P.H. and Rapp, A.
1972: Landslides in the Mgeta area, western Uluguru Mountains, Tanzania, Geografiska Annaler, v. 54A, p. 157-193.

Terzaghi, K.
1950: Mechanics of landslides; *in* Application of Geology to Engineering Practice; Geological Society of America, New York, p. 83-123.

Thorarinsson, S.
1939: The ice dammed lakes of Iceland with particular reference to their values as indicators of glacier oscillations; Geografiska Annaler, v. 21, p. 216-242.

Tiefenthaler, H.
1973: Natur und Verkehr auf der Arlberg Westseite; Innsbrucker Geographische Studien, v. 1, 250 p.

Till, A.
1907: Das grosse Naturereignis von 1348 und die Bergstürze des Dobratsch; Mitteilungen der Geographischen Gesellschaft Wien, v. 50, p. 535-645.

Tosolini, G.
1974: Le calamitá naturali nelle Alpi e Prealpi Carniche e Giulie (1700-1928); *in* Le Calamitá Naturali nelle Alpi, ed. G.B. Castiglioni; Istituto di Geographia dell' Universita di Padova, Novara, p. 59-72.

Toula, F.
1893: Der Bergsturz vom Arlberge und die Katastrophe von St. Gervais; Vorträge des Vereins zur Verbeitung naturwissenschaftlicher Kenntnisse in Wien, v. 33, Heft 14, p. 1-26.

Trener, G.B.
1957: Osservazioni geologiche sulla portata solida dell'Avisio; Studi Trentini di Scienze Naturali, v. 34, p. 246-305.

Trümpy, R. ed.
1980: Geology of Switzerland, a Guide-book; Wepf & Co., Bazel, 334 p., 2 parts.

Valbusa, U.
1921: La catastrophe del Monte Bianco e del ghiacciaio della Brenva del 14 e 19 Novembre 1920; Bolletino della Società Geographica Italiana, v. 10, p. 95-114 and 151-162.

van Bemmelen, R.W.
1970: The Geology of Indonesia, Volume 1A, General Geology of Indonesia and Adjacent Archipelagoes; Martinus Nijhoff, The Hague, 732 p.

Van Husen, D.
1981: Geologisch sedimentologische Aspekte im Quartär von Österreich; Mitteilungen der Österreichischen Geologischen Gesellschaft, v. 74/75, p. 197-230.

Varnes, D.J.
1978: Slope movement types and processes; *in* Landslides — Analysis and Control, ed. R.L. Schuster and R.J. Krizek; National Academy of Sciences, Transportation Research Board, Special Report 176, p. 11-33.

Venzo, G.A. and Largaiolli, T.
1968: Il bacino del Cinaga (Trentino); Museo Tridentino Scienze Naturali, Memorie v. 18/1, 32 p.
1969: Il bacino del Chieppena (Trentino); Museo Tridentino Scienze Naturali, Memorie v. 17/2, 103 p.

Venzo, G.A. and Vaia, F.
1972: Distribuzione e frequenza dei fenomeni franosi in alcuni bacini idrografici del Trentino e dell'Alto Adige durante le alluvioni del novembre 1966; Studi Trentini Scienze Naturali, Sezione A, v. 49, p. 3-13.

Vivian, R.
1966: La catastrophe du Glacier Allalin; Revue de Géographie Alpine, v. 54, p. 97-112.

Vogt, J.
1979: Les tremblement de terre en France; Bureau de Recherches Géologiques et Minières, Mémoire, no. 96, 220 p.

Voight, B.
1978: Lower Gros Ventre slide, Wyoming, U.S.A.; *in* Rockslides and Avalanches, v. 1, ed. B. Voight; Elsevier Scientific Publishing Co., Amsterdam, p. 113-166.

Voight, B., Glicken, H., Janda, R.J., and Douglass, P.M.
1981: Catastrophic rockslide avalanche of May 18; *in* The 1980 Eruptions of Mount St. Helens, Washington, ed. P.W. Lipman and D.R. Mullineaux; U.S. Geological Survey, Professional Paper 1250, p. 347-377.

Voight, B., Janda, R.J., Glicken, H., and Douglass, P.M.
1983: Nature and mechanics of the Mount St. Helens rockslide-avalanche of 18 May 1980; Geotechnique, v. 33, p. 243-273.

Von Moos, A.
1948: Setzungsmessungen 1887-1944 und Baugrund im Rutschgebiet von Zug; Eclogae Geologicae Helvetiae, v. 41, p. 113-123.

Wadell, H.
1935: Ice floods and volcanic eruptions in Vatnajökull; Geographical Review, v. 25, p. 131-136.

Walcher, J.
1773: Nachrichten von den Eisbergen in Tryol; Wien, 96 p.

Waldron, H.H.
1967: Debris flow and erosion control problems caused by the ash eruptions of Irazú volcano, Costa Rica; U.S. Geological Survey, Bulletin 1241-I, 37 p.

Weber, F.H. and Treiman, J.A.
1979: Slope stability and debris flows, Los Angeles area; California Geology, v. 1979, p. 3-12.

Weiss, E.H.
1964: Vaiont- Geologische Betrachtungen zur Felsgleitung in den Stausee; Steirische Beiträge zur Hydrogeologie.
1969: Geologische Übersicht (Voraussage und Erfahrung beim Bau) Kraftwerksgruppe Fragant; Österreichische Zeitschrift für Elektrizitatswirtschaft, v. 22, p. 464-472.

Whitehouse, I.E. and Griffiths, G.A.
1983: Frequency and hazard of large rock avalanches in the central southern Alps, New Zealand; Geology, v. 11, p. 331-334.

Williams, G.P. and Guy, H.P.
1973: Erosional and depositional aspects of Hurricane Camille in Virginia, 1969; U.S. Geological Survey, Professional Paper 804, 80 p.

Williams, H. and McBirney, A.R.
1979: Volcanology; Freeman, Cooper & Co., San Francisco, 397 p.

Wittmann, H.
1952: Eine Naturkatastrophe im Lavanttal; Carinthia I, v. 142, p. 430-443.

Wolf, H.
1875: Ber Bergsturz am Unterstein an der Salzburg-Tiroler Bahn; Verhandlungen der Kaiserlich-Königlichen Geologischen Reichsanstalt, 1875, p. 175-181.

Wolf, T.
1878: Geognostische Mittheilungen aus Ecuador. 5. Der Cotopaxi und seine letzte Eruption am 26. Juni 1877; Neues Jahrbuch für Mineralogie, Geologie und Palaeontologie (1878), p. 113-167.

Wu, T.H., McKinnell III, W.P., and Swanston, D.N.
1979: Strength of tree routes and landslides on Prince of Wales Island, Alaska; Canadian Geotechnical Journal, v. 16, p. 19-33.

Wühl, A.
1980: Die Wildbäche und Lawinen Oberösterreichs; International Symposium Interpraevent, Fieldtrip Guidebook 1980, p. 157-168.

Xizhi, D.
1981: Collected papers on debris flows (No. 1) in China; Peiping, 152 p.

Young, G.J.
1980: Monitoring glacier outburst floods; Nordic Hydrology, v. 11, p. 285-300.

Zankl, H.
1958: Der Bergsturz am 6/7 Februar 1959 im Wimbachtal (Berchtegadener Land), ein Beispiel für Bewegungsablauf und Erscheiningsform glazialer Bergsturze; Zeitschrift für Gletscherkunde und Glazialgeologie, v. 4, p. 207-214.

Záruba, Q. and Mencl, V.
1982: Landslides and Their Control; Elsevier Scientific Publishing, Amsterdam, 324 p.

Zay, K.
1807: Goldau und seine Gegend, wie sie war und was sie geworden; Zurich, 390 p.

Zehnder, J.N.
1974: Der Goldauer Bergsturz; Goldau, 207 p.

Zeller, G.
1964: Morphologische Untersuchungen in den östlichen Seitentalern des Val Blenio; Beiträge zur Geologische Karte der Schweiz, Hydrologie, no. 13, 108 p.

Zeller, J.
1972a: Beitrag zur Wildbachgeographie der Schweiz; Schweizer Zeitschrift für Forstwesen, v. 1972, p. 767-774.
1972b: Die Schwierigkeit einer technisch korrekten Festlegung der Wildbachgefahrenzonen; *in* 100 Jahre Hochschule für Bodenkultur in Wien, v. 4/2, p. 169-186.
1982: Einige Hinweise zur Geschiebeführung von Wildbächen; Mitteilungen der Forstlichen Bundesanstalt Wien, v. 144, p. 125-138.

Zen, M.T. and Hadikusumo, D.
1965: The future danger of Mt. Kelut (eastern Java — Indonesia); Bulletin Volcanologique, v. 28, p. 275-282.

Zischinsky, U.
1969: Über Sackungen; Rock Mechanics, v. 1, p. 30-52.

Zollinger, F.
1976: Die Ausscheidung von Gefahrenzonen; Dokumente und Informationen zur Schweizerischen Orts-, Regional-, und Landesplanung, no. 42, p. 28-38.

APPENDIX

ALPINE CASE HISTORIES

Setting

The high ranges of the Alps have been inhabited for thousands of years. Human experience with natural and man-induced mass movements has been profound and is well documented. This experience has contributed significantly to advances in the control of debris flows and unstable slopes. Yet no single publication, particularly in English, has attempted to bring together this Alpine experience which should be of great interest for the development of other high mountain regions, such as those of western Canada and the United States.

The Appendix consists of the most relevant case histories of Alpine mass movements, arranged chronologically, using the date of the most significant event where two or more recurrent events have occurred in the same region.

Initially this compilation was guided by summaries of mass movements that are difficult to obtain in standard libraries of North America (i.e. Mougin, 1914, 1931; Heim, 1932; Montandon, 1933; Strele, 1936; Stini, 1938; Stacul, 1979). In the later stages, literature on slope management and debris flow problems was examined by one of us (G.H.E.) in libraries in Austria, Switzerland, and France. To appraise ongoing technical or cultural adjustments in the vicinity of known historical mass movements all sites discussed in the text were visited in the field. Since locally published accounts of torrent and slope stability problems are numerous, many relevant case histories probably have been missed; the Appendix thus is not exhaustive. We feel, however, that it provides a fair representation of the most interesting, dramatic, costly and tragic mass movements witnessed in 2000 years of Alpine settlement.

The brief accounts may guide the reader to older and often unjustly forgotten accounts in German, French, and Italian. The sketch maps, photographs, and line drawings accompanying the text cannot substitute for the more detailed documentation that some of the examples have received and others still deserve. Rather, the main purpose of the Appendix is to describe Alpine mass movements within a coherent topographic, geological, climatic, seismic, and historical framework that transcends national boundaries. It must be kept in mind that in an area like the Alps a long history of border changes, migrations, and distant administrative centres have led to a variety of names and spellings for specific localities. In order to avoid ambiguity we have not transformed local nomenclature into English equivalents and have used the one generally preferred by the inhabitants of an area or that shown on widely distributed maps. Figure 40 is an index map showing the location of the 137 sites described in the Appendix. Case histories can be located by use of the Index of Case Histories at the end of the Appendix.

Bedrock geology and seismicity of the Alps

The Alps have been the focus of intense geological research for 200 years. Variety, complexity, and accessibility have made this mountain chain one of the classical areas of structural geology. Publications dealing with Alpine geology are therefore countless. Well illustrated recent summaries include those by Gwinner (1971), Lemoine (1978, p. 17-66), Trümpy (1980), and Oberhauser (1980).

The Alps originated by compression and foreshortening of continental and oceanic crust located originally between Africa and Europe. This process of crustal shortening, which began in late Mesozoic time and continues to the present, has displaced the crystalline basement and its sedimentary cover complexes by hundreds of kilometres northward; several distinct facies zones have been juxtaposed along north-directed thrust faults. During the main stages of deformation in latest Mesozoic and early Tertiary time some of the rocks overridden by thrust nappes from the south suffered intense regional metamorphism. Locally, granitic intrusives rose along fractures and faults into the deformed pile of crustal rocks.

The bedrock geology of the Alps can be conveniently summarized in terms of five realms or zones, characterized by pre-Mesozoic crystalline basement and a distinct Mesozoic-Tertiary sedimentary cover. From southeast to northwest these zones are: South Alpine Zone, Austroalpine Zone, Pennine Zone, Helvetic Zone, and Molasse Zone (Fig. 41, 42).

The South Alpine Zone consists of a Paleozoic metamorphic-volcanic basement overlain by thick Mesozoic carbonates and shales. Mountainsides underlain by metamorphic basement rocks tend to display deep-seated creep (sagging) and bedrock slumps. Massive carbonates of the cover complex (Sciliar Formation, Dolomia Principale etc.) often collapse along scarp slopes underlain by recessive evaporite-marl units (Werfeniano, Raibliano etc.). Dip slope failures are observed in thick-bedded carbonate units. Large earth and debris flows develop in shale terrain at the foot of soaring carbonate massifs.

The geology of the Austroalpine Zone is very similar to that of the South Alpine Zone. The two realms are separated by an east-trending strike slip (?) fault, the Periadriatic (Insubric, Gail) Line. The Austroalpine Zone is characterized by a Paleozoic basement of phyllite, gneiss, and minor carbonate rocks, overlain by a thick Mesozoic carbonate-shale cover. Again mountainsides underlain by metamorphic rocks tend to fail by deep-seated creep and slumping. The presence of evaporitic formations (Werfen, Raibl etc.) favours scarp slope failure in overlying resistant carbonates.

The Pennine Zone (Briançonnais in France) consists of a metamorphic basement and a thick, predominantly slaty-sandy Mesozoic cover. Over wide areas the rocks of the

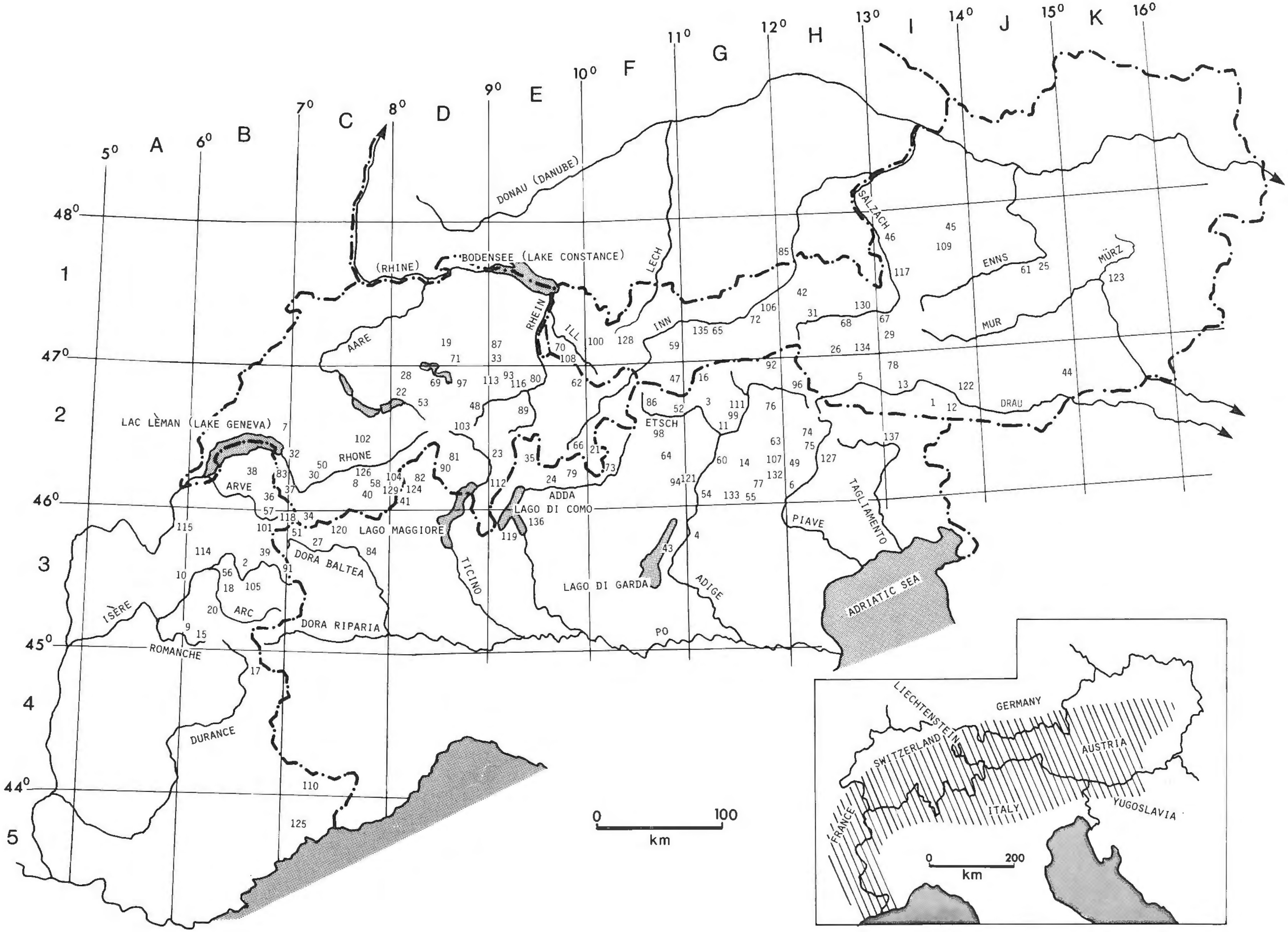


Figure 40: Index map of the Alps showing the location of case histories presented in the Appendix.

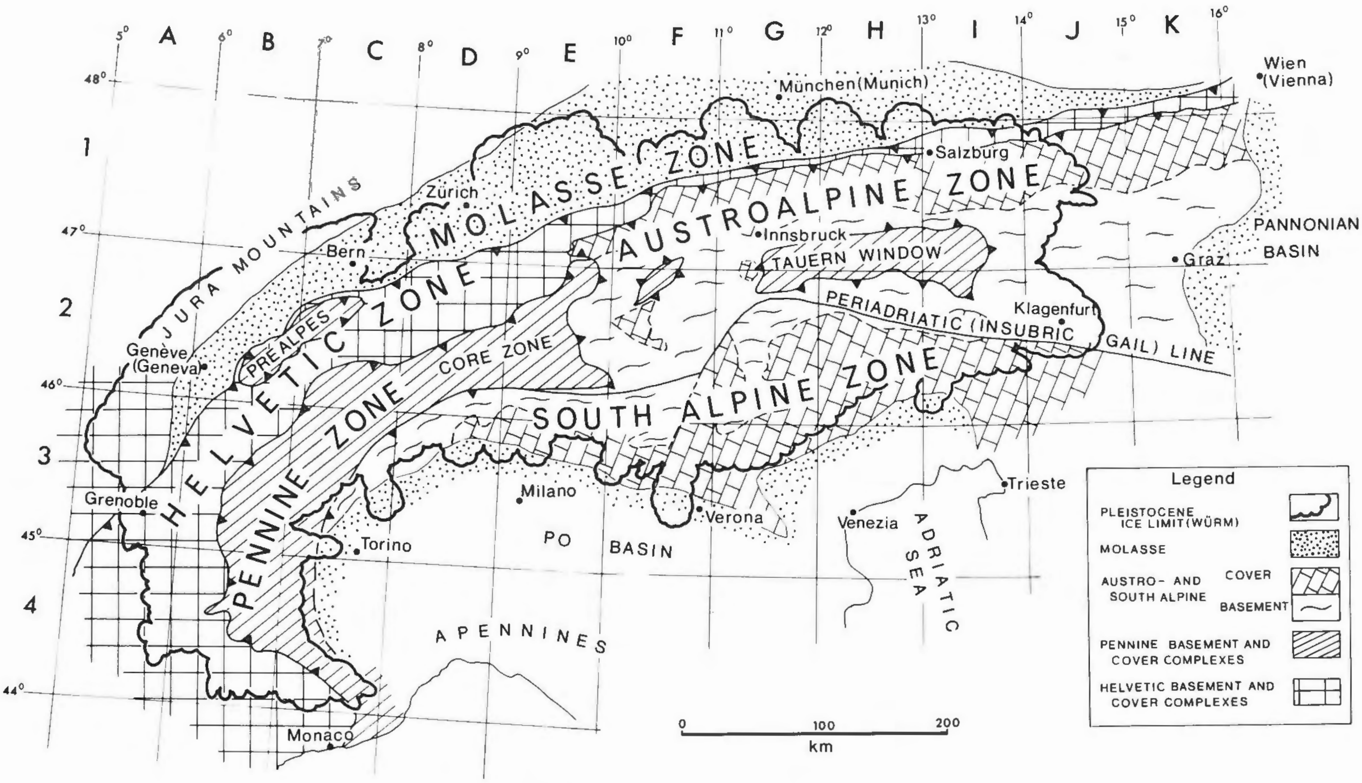


Figure 41: Geological sketch map of the Alps outlining the distribution of the five major tectonic-lithological zones, their thrust-fault contacts (barbs on the upper plate), and the extent of late Pleistocene glaciation.

Pennine Zone were overridden by Austroalpine thrust nappes and therefore suffered extensive regional metamorphism, particularly in the Pennine core zone of southern Switzerland and in the Tauern Window of central Austria. Dip slopes underlain by Pennine slates and schists are susceptible to deep-seated creep and large-scale sagging (Huder, 1976).

The Helvetic Zone (Dauphinois in France) is underlain by a Paleozoic basement complex of granite, gneiss, and quartzose clastic rocks. The cover succession consists of Mesozoic and Tertiary shale, carbonate and flysch. The cliff-forming Malm and Urgonian limestones fail along spectacular dip slopes and scarp faces. Many slopes underlain by evaporitic shale, slate, and flysch tend to undergo deep-seated slumping and creep that trigger embankment failures along many torrents. The term flysch (= flow), which is used to describe monotonously interlayered successions of sandstone and shale, owes its origin to chronic slumping evident on many mountainsides of the Helvetic Alps.

The Molasse Zone consists of massive conglomerate, sandstone, and mudstone which represent the indurated product of the early erosion of the rising Alpine chain. The generally southeast-facing dip slopes in the Tertiary Molasse of the northern foreland of the Alps are susceptible to large-scale bedrock failures and to debris slides involving deeply weathered blocky bedrock and colluvium.

The Alps attain their greatest elevations in the uplifted basement massifs of the Helvetic Zone (Mont Blanc Massif-4807 m) and in the metamorphic core complex of the Pennine Zone in southern Switzerland and central Austria. Separating the main ranges are fault-controlled longitudinal valleys and intramontane troughs drained by large rivers (Rhone, Inn, Drau). Bedrock in these depressions is generally buried under thick surficial deposits of Pleistocene age.

Parts of the longitudinal depressions (e.g. Rhone-Rhine valleys in Switzerland, Inn Valley and Mur-Mürz valleys in Austria, Val Sugana in Italy) are seismically active (Baratta, 1901; Schorn, 1902; Pavoni, 1977; Vogt, 1979; Oberhauser, 1980). The present seismicity of the Alps is a manifestation of a north-northwest-trending neotectonic compression in a stress regime which continues to thrust the Alpine ranges over its southern foreland along a north-dipping subduction zone (Fig. 41). This southern foreland zone has experienced the most devastating earthquakes in the Alps (e.g. the Friuli region in 1976).

Pleistocene history

During the Pleistocene large mountain glaciers and ice fields covered most of the Alps (Fig. 41). Flowing from high central ranges through deep intramontane depressions lobes of ice reached beyond the front ranges onto the foreland.

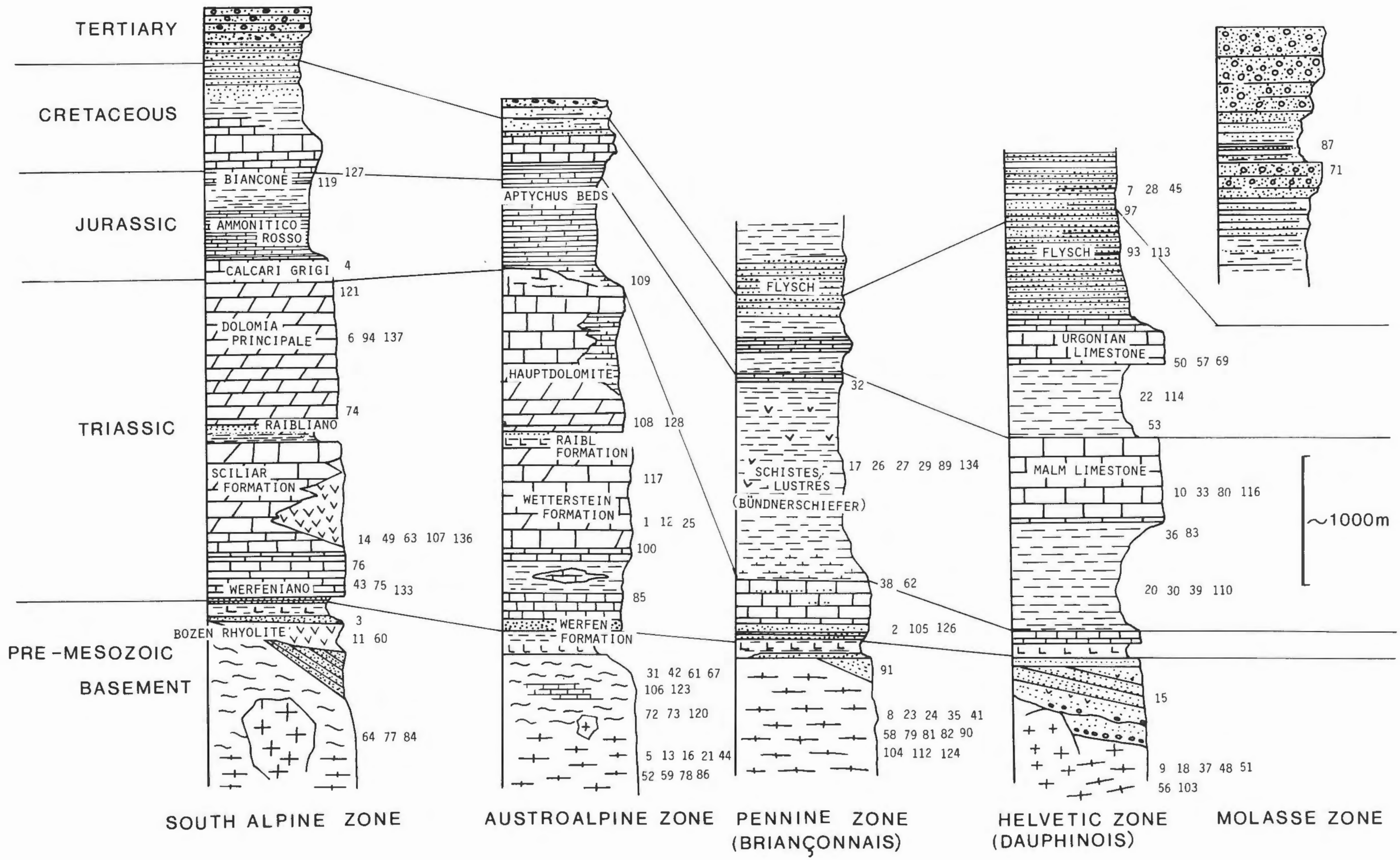


Figure 42: Generalized stratigraphic columns of the five Alpine geological zones shown in Figure 41. Case histories in the Appendix are identified with respect to their position in the stratigraphic columns.

Although deposits related to older glaciations are preserved locally most of the surficial deposits are related to the final stages of the last glaciation (Würm Glaciation). This last great ice sheet attained its maximum extent between 20 000 and 17 000 years ago and its retreat was effectively complete by about 9500 years ago (*see* Trümpy, 1980; Van Husen, 1981).

Downwasting of valley glaciers left behind a variety of coarse colluvial, lacustrine, deltaic, and alluvial deposits, commonly preserved as terraces along the main valleys. In many small tributary basins, particularly those underlain by metamorphic bedrock, terraces of relict late Pleistocene ice margin deposits are persistent sources of debris for high-gradient mountain torrents. Debris flows derived from these deposits have contributed to the growth of large fans and cones fringing the valleys.

Deglaciation locally triggered collapse of rock slopes. Evidence for late glacial rock avalanches is particularly common along high carbonate cliffs of the Helvetic, Austroalpine, and South Alpine zones (Abele, 1974).

Other steep slopes, particularly those underlain by low-grade metamorphic rocks (e.g. phyllite, slate, micaschist) responded to deglaciation by deep-seated creep or 'sagging'. In places, zones of broken rock up to a few hundred metres thick and extending for several kilometres along the valley trend, form what has been referred to as 'Bergzerreissung' (= mountain rupture; Ampferer, 1939), 'Talzuschub' (= valley closure; Stini, 1942), or simply 'Sackung' (= sagging; Heim, 1932; Zischinsky, 1969). This metastable state of equilibrium along many mountainsides, initiated by deglaciation, has remained a very significant factor with regard to historical mass movements in the Alps: excessive infiltration of water into sagging slopes can reactivate or accelerate their movement setting up potentially dangerous blockages of torrent channels in the toe zone.

Little Ice Age

The period of climatic deterioration during the Middle Ages, but particularly between approximately 1550 and 1850, is generally referred to as the Little Ice Age. This period was characterized by marked advance of the glaciers and by a regional climate that interfered seriously with human activity in the higher Alpine valleys. The Little Ice Age has been studied using historical records and analyzing terminal moraines deposited in the forefields of the present Alpine glaciers (Kinzl, 1932; LeRoy Ladurie, 1971; Patzelt, 1973; Messerli et al., 1975; Bachmann, 1978; Schneebeli and Röthlisberger, 1976).

Climatic deterioration and advancing glaciers began to affect human activity in the highest parts of the Alps as early as the 13th century. However, distinct successions of cold and rainy summers in central Europe, coinciding with poor harvests in the lowlands and advancing ice in the Alps, turned into a well perceived threat around the middle of the 16th century. At the end of the 16th century there were several glacier-related floods and debris flows of regional impact. Between about 1550 and 1850 periods of glacier advances, following years with cold summers, alternated with periods of minor retreat (Messerli et al., 1975). The Little Ice Age was followed by a long period of glacier retreat and climatic amelioration which lasted from the 1850s to the mid-1960s. Since then the mass budgets of many Alpine glaciers have been positive, and once more the termini of many glaciers are advancing (Bachmann, 1978).

Climate

In a broad sense the climate of the Alps reflects three major weather systems: one originating in the Atlantic Ocean ('west weather'), another in the Mediterranean Sea, and a third in the continental interior of eastern Europe. On a local scale the weather in the Alps is modified greatly by aspect and elevation of individual mountain ranges. Precipitation, the most important factor for slope stability, is greatest along the northern and southern front ranges of the Alps (Fig. 43). There, blockage of moist air from the Atlantic and the Mediterranean triggers sustained catastrophic rainstorms (e.g. 1733, 1770, 1851, 1868, 1882, 1910, 1966). Local cloudbursts or downpours during hot 'continental' summers occur in many intramontane basins and along the eastern foothills of the Alps. In most of the Alps, precipitation is greatest in the autumn and summer (Fliri, 1977). Destructive mass movements tend to follow the same pattern: most mass movements, other than those released by earthquakes occur during rainstorms, delayed springtime snowmelt coinciding with warm rains, local cloudbursts, glacier-related floods, and ice-debris avalanches from the snout of hanging glaciers. A plot of all case histories described in the Appendix according to month of occurrence shows that most serious mass movements take place between June and September, the months of greatest mean precipitation (Fig. 44). This graph supports previous, more local, observations (e.g. Heim, 1932; Nussbaum, 1957).

On a regional scale potentially destructive debris flows and landslides seem to cluster near areas with strong climatic and topographic gradients (e.g. Tagliamento-Piave basins, Ticino region, Mont Blanc Massif).

The relationship between destructive mass movements in the Alps and long term climatic variability is more difficult to assess. In Figure 45C nonseismic destructive mass movements described in the Appendix are grouped in ten-year intervals. In Figure 45B the thoroughly documented chronology of minor and major debris flows in a single large transverse valley of the central Alps (Ötz Valley) is plotted in the same fashion (Leys, 1977). Figure 45A shows the fluctuations of the snout of the Grindelwald Glacier in central Switzerland (Messerli et al., 1975), providing an indirect measure of climatic change.

A very strong cultural-historical 'noise' induced by the necessarily fragmented data base can be expected to weaken any broad relationship that might exist between climate as expressed in retreats and advances of major Alpine glaciers and the incidence mass movements. Therefore, any vague parallels emerging in Figure 45 should only be taken as hints. However, times of short-term climatic deterioration coinciding with cold and rainy summers, first in the late 16th century, then between 1760 and 1780, and again between 1810 and 1860 were also periods with enhanced mass movements.

In detail the effect of individual intense storms, such as those in the 1430s, 1560s, 1660s, 1740s, 1760s, 1790s, 1820s, 1850s, 1880s, and 1960s probably had greater impact on human settlements and transportation routes then the gradual change of precipitation pattern referred to above. The recent events of 1965 and 1966 serve as notable examples: although a majority of glaciers advanced slightly in response to cooler and wetter summers since the mid-1960s, the damage suffered from the extreme storms events (particularly those of 1966) far outweighed that from the subtle adjustments of the general atmospheric circulation pattern. The experience gained during sporadic intense meteorological events therefore has been and still is the main basis for appropriate land use decisions and the design of effective remedial measures (Aulitzky, 1968).

Land use history

In the Alps, as elsewhere, destructive mass movements result from both natural causes and human activity. The extremely long history of clearing, colonization, and intense land use in the Alps up to elevations of 2500 m make it difficult to separate natural from induced mass movements.

The clearing of forests in high basins can be traced back to the activities of Celtic miners and Roman military men. However, the main phase of deforestation of Alpine mountain valleys dates to the early Middle Ages when a scarcity of available farm and pasture land in the lowlands of Europe forced colonization of the mountains. At this time the extensive tracts of subalpine forest were converted into seasonal pasture. Clearcutting and overgrazing locally triggered severe erosion. Exhaustion of thin soils caused regional population shifts within the Alps during the 13th century (e.g. 'Walser' migrations) and intensified some of the problems brought on by the initial clearing and overgrazing. During the 15th and 16th centuries the requirements for timber and charcoal in districts with newly opened mines, salterns, and smelters brought another phase of deforestation. Finally, wars and early railroading strained the shrinking forest and land resources.

Most of the early settlers of the Alps tried to avoid flood-prone valley bottoms and preferred locations on bedrock benches and debris cones. Sporadic bursts of debris from upland basins onto debris cones of normally tranquil mountain streams soon brought home some of the simple relationships between land use and debris transport in steep mountain terrain. Nevertheless, raging giants and infuriated dragons long remained the most common explanations for

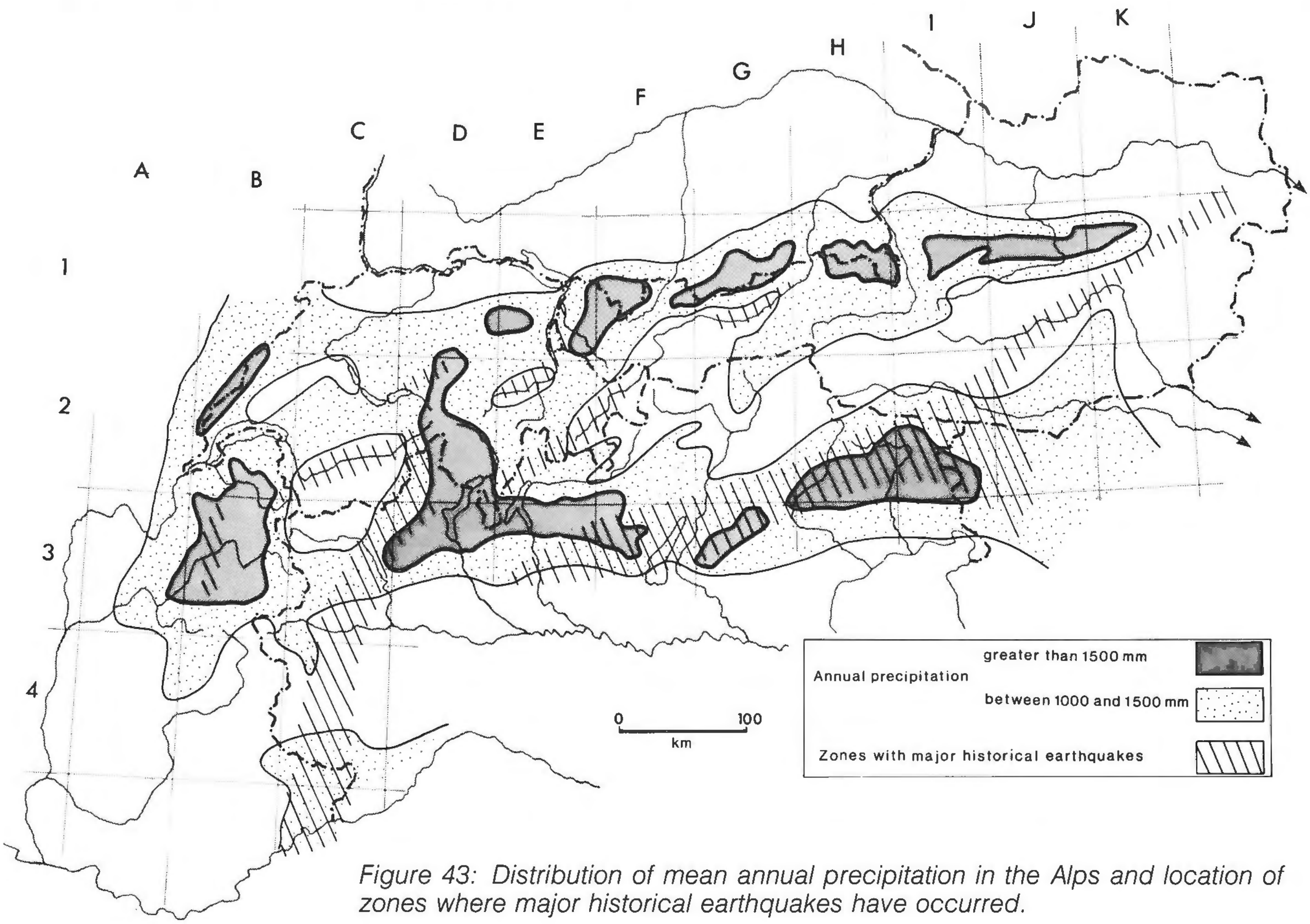


Figure 43: Distribution of mean annual precipitation in the Alps and location of zones where major historical earthquakes have occurred.

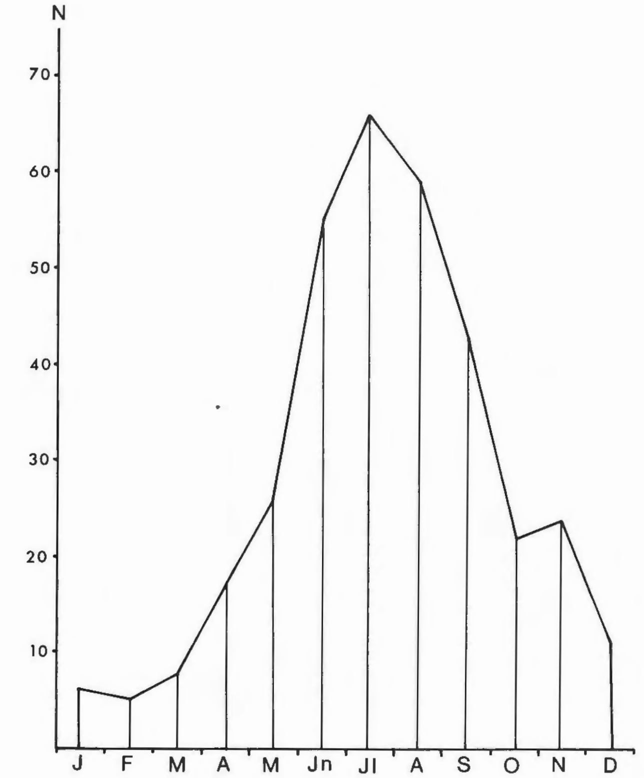


Figure 44: Diagram showing the month of occurrence for non-seismic mass movements presented in the Appendix; N = number of case histories in a certain month.

the sudden roar in the gorges and the violent eruption of rubbly debris onto fields and communities.

In spite of these superstitious explanations the repeated exposure to rockfalls, debris flows, and snow avalanches led to simple restrictive community rulings as early as the 14th and 15th century. Critical forest land ('Bannwald') was set aside in the source areas and on the cones of hazardous torrents. Dykes and walls were constructed to deflect occasional spurts of debris away from threatened communities.

Continued deterioration of the uplands in the 18th and 19th centuries due to clearcutting and overgrazing, coupled with serious competition from mechanized farming activity in the lowlands of central Europe led to depopulation of many high valleys, particularly in the western Alps. The first systematic regional government legislation dealing with torrent control, the protection of existing forests, and reforestation was enacted after major natural disasters in the second half of the 19th century (Eisbacher, 1982).

The present intense development of Alpine valleys is mainly related to tourism and hydroelectric power generation. This development has seriously taxed the long-neglected renewable resource base of the Alps (i.e. forests, water, soil) and has led to a reappraisal of traditional and modern land use patterns (Aulitzky, 1968, 1972, 1974).

Reforestation or engineering works along torrents, on unstable slopes, and near known snow avalanche tracks recently have been supplemented by programs emphasizing monitoring and hazard zoning of areas with high population densities.

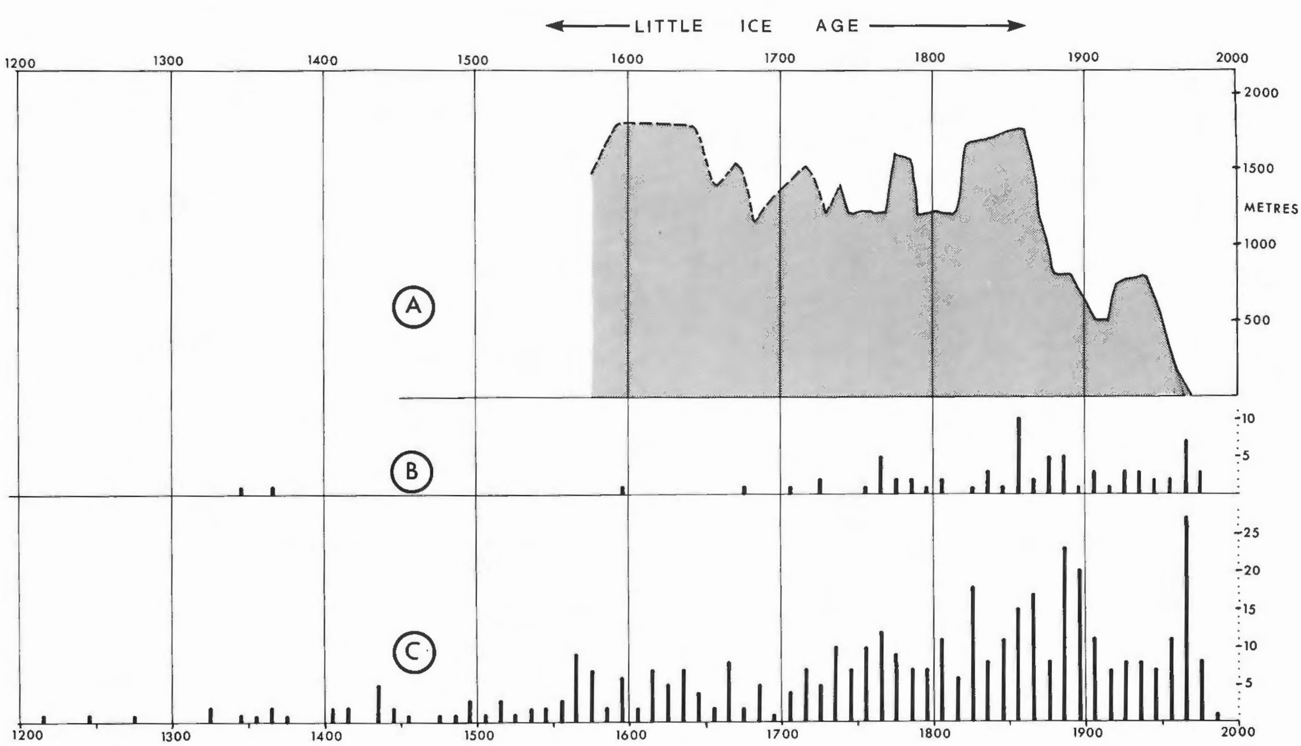


Figure 45: A: Historical fluctuations of the snout of the Grindelwald Glacier relative to its position in 1970, measured in metres along the glacier bed (from Messerli et al., 1975). B: Number of historical destructive mass movements in the Ötz Valley, western Austria, grouped in ten-year intervals (data from Leys, 1977). C: Number of destructive mass movements presented in this Appendix, grouped in ten-year intervals.

Reisskofel (A1)

Location: Gail Valley, Kärnten (Carinthia), Austria (I2)[1]
Date(s): 328 A.D.

The Gail Valley follows a major west-trending fault zone, the Gail Valley Line, which separates the South Alpine and Austroalpine basement-cover complexes. Numerous debris cones flank the valley and the Gail River receives a voluminous bedload from its tributary torrents. Between the villages of Reisach and Grafendorf (630 m), a huge debris cone rises northward from the bottom of the valley to an apex at an elevation of 900 m (Fig. 46); from the apex a wide track of angular carbonate debris branches upwards into several bedrock ravines of the Reisskofel south face. This scarp face of the massif is composed of thick-bedded carbonates dipping between 40 and 80° to the north. The cliff-forming carbonates overlie recessive redbeds of the Austroalpine basement-cover transition. Steep fractures in the carbonates dip towards the Gail Valley and separate highly unstable rock towers from the main Reisskofel Massif (2371 m).

Figure 47: View north from the apex of the Reisach cone to the jagged carbonate scarp face of the Reisskofel Massif. Note recently built hotel. (GSC 204165-L)

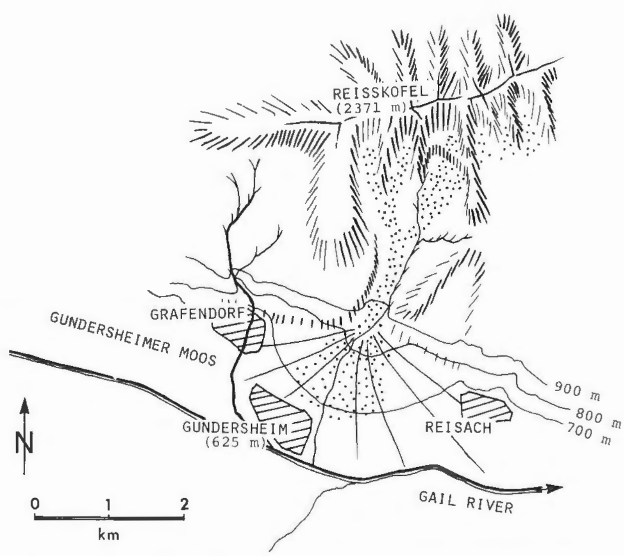


Figure 46: Index map of the Reisskofel Massif and the huge cone of Reisach. Notice that traditional settlements cluster along the periphery of the cone.

According to the 'Memorabilienbuch' (chronicle) of Grafendorf the debris cone was once the site of a Roman gold mining town, named Risa. In 328 A.D. this settlement was buried by debris which also dammed the east-flowing Gail River (Till, 1907). A local legend adds that an earthquake had shattered parts of the Reisskofel before a rainstorm mobilized the loose blocks into a flow that swept over the blooming but 'sinful' village.

Today most of the rubbly surface of the upper cone is covered by dense forest. Angular blocks of carbonate protrude through the thin soil. A small creek drains the cone. In recent years settlement has encroached onto the upper cone, and a large hotel complex has been erected across the ancient debris track (Fig. 47). The former lake on the west side of the debris cone in the course of time silted in and this Gundersheimer Moos (= swamp) was reclaimed for agricultural activities during the comprehensive regulation of the Gail River. Although in recent years there has been no major rock avalanche from the bowl-shaped carbonate walls of the Reisskofel, fresh blankets of talus indicate continued rockfall activity from the tilting carbonate slabs below the summit ridge.

[1] Location in grid of Figure 40.

Aime (A2)

Location: Tarentaise, Savoie, France (B3)
Date(s): 5th century A.D. (also 2nd century A.D., 8 September 1579, 14 September 1733, 10 June 1764, 26 October 1778, 18 July 1834, 1 November 1859, 20 April 1893)

Near the town of Aime (600 m) the Isère River follows the regional trend of major bedrock structures including a complicated southeast-dipping zone of thrust faults involving carbonate rocks, calcareous flysch, shale and evaporite of the Helvetic (Dauphinois) and Pennine (Briançonnais) cover complexes. Debris fans and cones on the north side of the valley for centuries have been inhabited in preference to adjacent unstable slopes and floodplain. Thus Aime developed on the fan of the Ormente Torrent and the villages of Villette and Centron on the flanks of the Nant Agot cone (Fig. 48). Catchment and source areas for both torrents extend to about 2500 m elevation and include unstable dip slopes of calcareous flysch and terraces of relict colluvium.

Excavations on the Ormente fan have revealed walls of the Roman town of Axima underneath a 3 m blanket of debris. The layout of the dwellings indicates that the occupants had been exposed to repeated debris flows before much of the town disappeared under masses of rubble sometime in the 5th century A.D. Use, and possibly, abuse, of the land in the upper Ormente basin also seems to date back to this early period (Chavoutier, 1979, p. 11). During the medieval growth of Aime at the site of the buried Roman settlement, unstable slopes in the Ormente basin contined to launch debris flows into the Isère Valley. Mass movements frequently coincided with rainstorms brought on by the feared 'vents du midi', warm winds from the south. Aime was seriously damaged by debris flows during regional rainstorms on 8 September 1579 and 14 September 1733. On the latter date a swath of debris, 50 m wide tore across the town (Fig. 49). A local cloudburst and sudden snowmelt on 10 June 1764, released another destructive pulse of debris onto the town. On 26 October 1778, following an extended period of rain, a series of debris flows demolished or seriously damaged a total of 36 houses. On 1 November 1859, snowmelt due to unseasonal air temperatures up to approximately 20°C and furious squalls of rain brought new havoc (Mougin, 1914, p. 770).

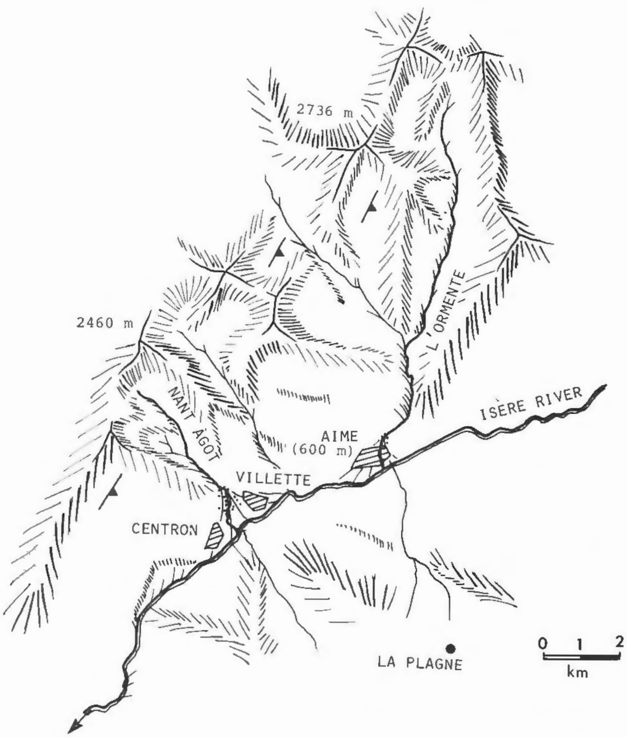


Figure 48: Index map of the Isère Valley in the vicinity of Aime; slaty cleavage in argillaceous rocks and bedding dip southeast, favouring slump-generated debris flows and embankment failures along south-facing slopes.

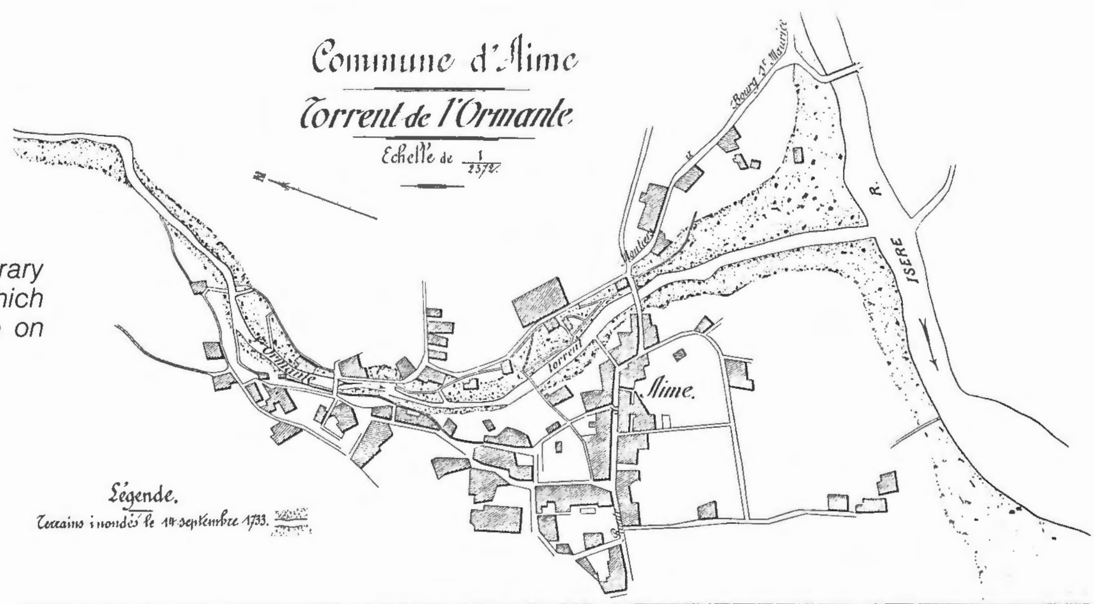


Figure 49: Contemporary map of a debris flow which invaded parts of Aime on 14 September 1733.

The steep cone of the Nant Agot, downstream from Aime, also hosted some kind of Gallo-Roman settlement which probably disappeared underneath bouldery debris in the 2nd century A.D. A debris carpet, 6 m thick, was found to cover foundations of residential buildings. The later settlement of Centron on this cone also suffered repeatedly from destructive flows during periods of sudden snowmelt (10 June 1764, and 20 April 1893) and intense rainstorms (18 July 1834). Repeated destruction of human works and loss of life have forced traditional building activity onto the sides of the cone (Mougin, 1914, p. 773).

Today Aime is a prosperous centre of tourism, serving the modern ski developments on the southern uplands of the Tarentaise (e.g. La Plagne). The fan of the Ormente Torrent is completely built over and its channel along the axis of the fan is confined by strong stone-masonry embankments. Parts of the formerly ravaged uplands have been reforested. The cone of the Nant Agot, much steeper than that of the Ormente, has not been developed to the same extent. An inconspicuous creek bed meanders across the boulder-strewn and forest-covered apex and then skirts the settlement of Centron, which is gradually expanding towards the axial sector of the cone.

Meran — Merano (A3)

Location: Südtirol (Alto Adige), Italy (G2)
Date(s): Between 785 and 815 A.D. (also 1372)

The coalescent debris fans of the torrential Passer River and Naifbach Torrent form a gently inclined surface on the eastern bank of the Adige (Etsch) River (Fig. 50). Meran (320 m) spreads across the lower part of the surface. This beautiful city grew from clusters of agricultural communities whose roots can be traced to pre-Roman time. An extremely pleasant climate and an adequate elevation above the floodplain of the Adige River were the principal reasons for the early growth of these settlements.

The catchment basin of the Naifbach Torrent includes the steep western face of the granitic Ifinger Massif (2500 m); the deeply incised gorge of this torrent follows a westerly trending fault zone involving shattered granite, redbeds, and rhyolite of the South Alpine basement-cover transition. Slump scars and rockfall chutes abound along the walls of the gorge. Lobate deposits of granitic debris blanket the northern side of the gorge.

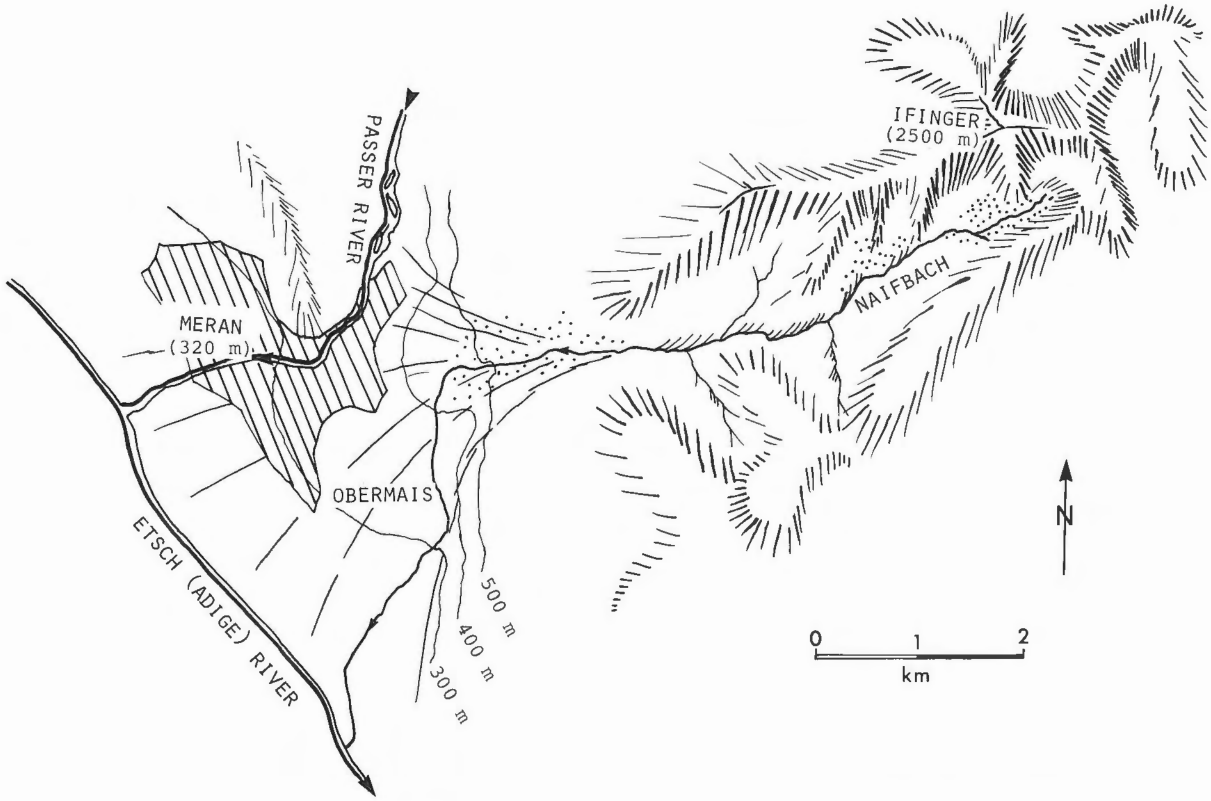


Figure 50: Sketch map of the large cultivated debris cone of the Naifbach Torrent near Meran. Rare bursts of debris, launched from bedrock failures in the narrow gorge and along the crest of the catchment basin, pushed the torrential Passer River against a bedrock spur along the west side of the valley during the early settlement of the region. Notice that the cone is disproportionately large relative to the catchment area.

Sometime between 785 and 815 A.D. a major rock-slide-debris flow apparently laid waste to the Roman town of Maia, then located near the present community of Obermais. Excavations during the 17th and 18th centuries encountered Roman walls and cellars deeply buried in debris; human skeletons of possibly more recent age were also unearthed (Aretin, 1808, p. 73). Another destructive outburst of debris over the Naifbach fan occurred in 1372. These and other smaller debris flows from the Naifbach gorge gradually pushed the channel of the Passer River north towards a bedrock spur near the area that was to become the heart of the city of Meran. During the later debris floods of the Passer River which originated from the bursts of the Passeier Wildsee (*see* Ganderberg — Passeier Wildsee (A16), 1419 flood), the impact near the channel restriction was catastrophic (Dalla Torre, 1913, p. 285-286; Stacul, 1979, p. 63-68).

Early protective works along the Naifbach Torrent attempted to confine the channel to the southern uninhabited sector of the cone by means of simple diversion dams. Subsequent erosion by the torrent into the fan was brought under control by check dams and revetments.

Beginning in 1882 segments of the fan that had been ravaged regularly by debris floods were reforested. In the uplands check dams were inserted into gullies that had been susceptible to chronic slumping of redbed formations.

Today vineyards, orchards, vacation homes, and tourist facilities mingle with old estates on the surface of the fan. A few large blocks of granite are scattered through the orchards of the upper cone, serving as reminders of the sudden debris flows from the gorge. The narrow Naifbach channel on the fan now is inconspicuous and a major hotel-tram station complex has been erected across its apex, serving the ski area of the Ifinger Massif. A protective forest clings to the steep bedrock slopes along most of the Naifbach gorge.

Lavini di Marco (A4)

Location: Val Lagarina, Trentino, Italy (G3)
Date(s): 883 A.D. or 21 July 365 (or 369) A.D.

Val Lagarina and other fault-controlled valleys parallel to it (including the depression of Lago di Garda) are flanked by precipitous mountain ranges of massive Mesozoic carbonate rocks of the South Alpine cover complex (Fig. 51, 52). A distinctly Mediterranean vegetation and white walls composed of Dolomia Principale and Calcari Grigi formations provide the spectacular background for the towns along the sides of the valley. Many bedrock slopes parellel the north-northeasterly trend of bedding, forming either dip slopes or vertical scarps. In detail rock slope morphology is also influenced by local high-angle fracture zones and faults. Earthquakes and intense autumn rains contribute to incipient instabilities on many cliffs. In historical times the region has suffered strong earthquakes on 21 July 365 (or 369) A.D., in 543 A.D., on 3 January 1117 (or 1111), in September 1419, and in 1457. Rockfalls have accompanied these and other earthquakes.

The region hosts impressive rock avalanche deposits ('marocche'), bare lobes of carbonate slabs that have nurtured legends about disappeared villages and towns. Some of the rock avalanches have resulted from dip slope failures ('scivolamenti'), others from collapse above or detachment along composite fracture surfaces ('frane di crollo').

In western Trentino the best-known prehistorical and historical slide masses are Monte Spinale (500 × $10^6 m^3$), Lago di Tovel (250 × $10^6 m^3$), Lago di Molveno (300 × $10^6 m^3$), Marocche di Sarca (350 × $10^6 m^3$), San Giovanni (25 × $10^6 m^3$), Torbole (40 × $10^6 m^3$), Lavini di Marco (200 × $10^6 m^3$), and Castel Pietra (1 × $10^6 m^3$). The low inclination of some rupture surfaces led some early investigators to reject a rock avalanche origin for the rubbly deposits in favour of a morainal origin. However, from an investigation of the breakaway zones Schwinner (1912) demonstrated that large-scale cliff collapse was the most probable mechanism. More recently, it has been suggested that some of the rock avalanches fell onto stagnating late Pleistocene glaciers, an interpretation that has gained considerable support (Abele, 1974).

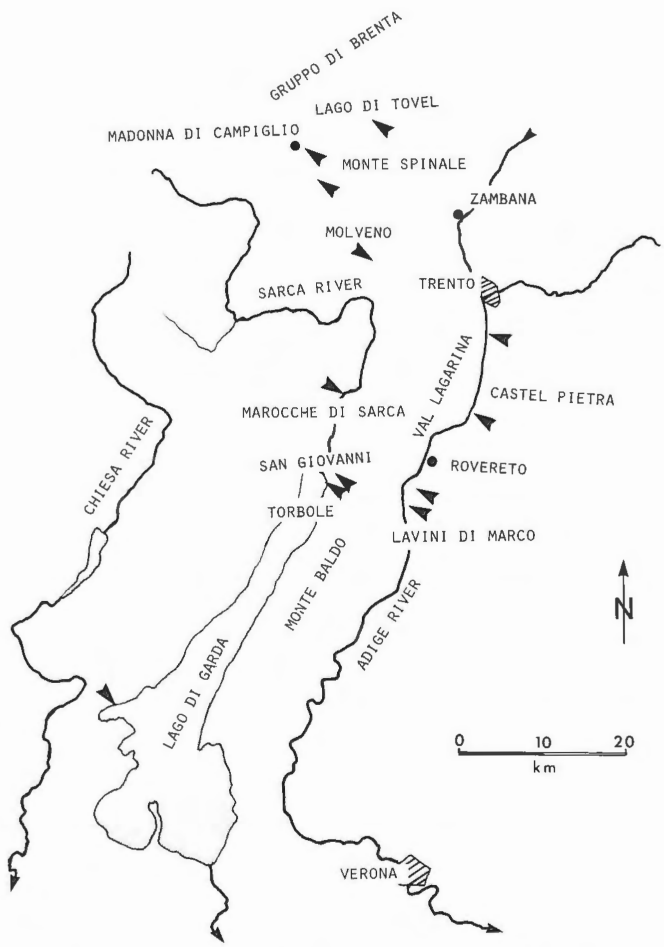


Figure 51: Regional index map of the central Trentino region, showing the location of major prehistorical and historical rock avalanches (arrows).

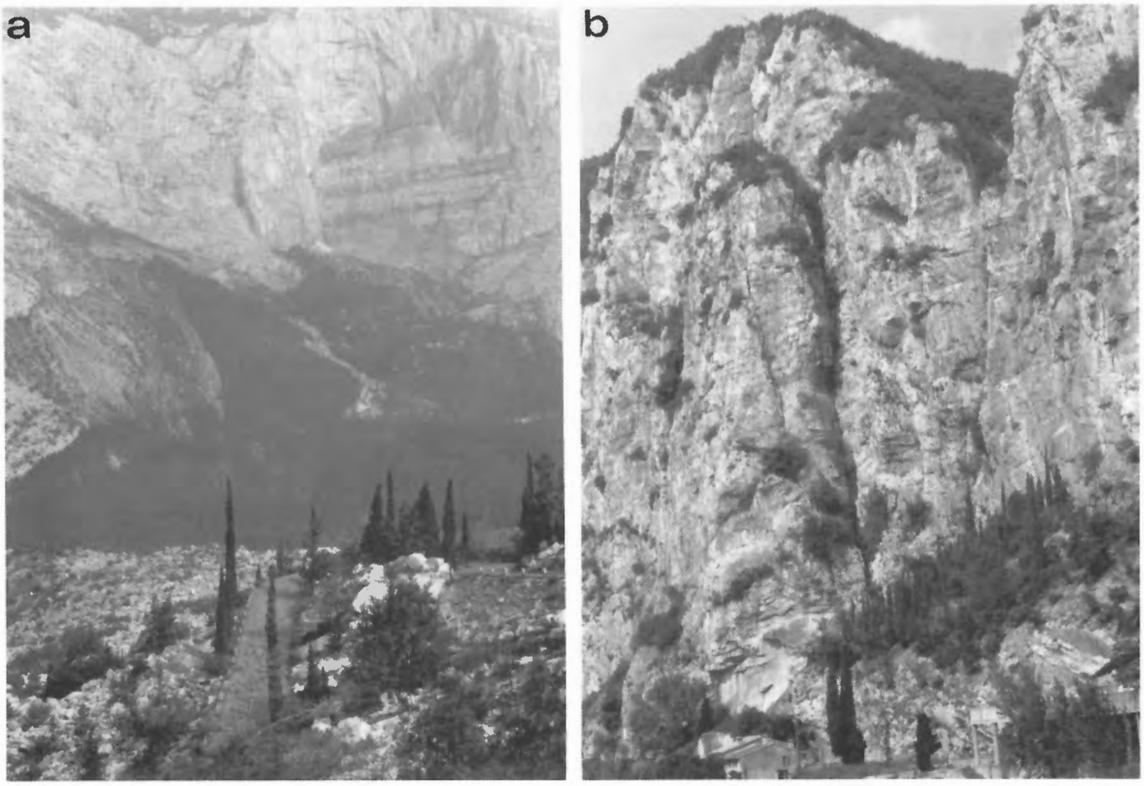


Figure 52: a) View across the Marocche di Sarca onto the concave zone of rupture from which recurrent rockfalls and avalanches have spread over the valley floor; (GSC 204165-M). b) Incipient detachment of joint-controlled rock towers in gently dipping massive carbonate strata along the west side of Lago di Garda; most of the highway is protected from rockfalls by strong concrete galleries, tunnels, and slope reinforcements (GSC 204165-N).

The Lavini di Marco, south of Rovereto, has attracted most attention. The first reference to the Lavini di Marco dates back to 1270 (Gorfer, 1977, p. 154). However, the fame of the Lavini di Marco rests mainly on a particularly vivid passage in Canto 12 of Dante's (1265-1321) Inferno:

'. . . As is that rubble, which struck the Adige in its flank, on this side of Trento, by earthquake or defective support-that from the mountain top from where it moved, to the plain the rock is so inclined that it might offer passage to one that were above . . .'.

Legend, speculation, and serious study have left a somewhat confused picture of the Lavini (Fig. 53). Some investigators have linked the blocky lobe to an event recorded in the Annals of Fulda, according to which a mountainside collapsed across the south-flowing Adige River in 883 A.D., blocking its course and depriving the citizens of Verona of their main source of water for several days. Others attribute the collapse of the mountain to the great earthquake of 21 July 365 (or 369) A.D. and also relate the rock avalanche to the disappearance of the town of Lagaris. Whatever the interpretation, some ruined human works have been found at a depth of 15 m below the surface of the debris (*see* Gorfer, 1977, p. 154-155).

The main rock avalanche originated by the failure of possibly up to $200 \times 10^6 m^3$ of massive oolitic limestone and interbedded marl (Calcari Grigi Formation). The rupture surface ('lasta' = dip slope) is inclined 23^0 and is a composite of bedding planes dipping 18^0 to the west and a set of crosscutting near-vertical fractures trending north-northeast (Fuganti, 1969, p. 7-22). It rises south-southeasterly from the bottom of Val Lagarina (200 m) to the peak of Zugna Torta (1250 m). The movement of the failed rock mass was guided by the dip of the basal rupture surface and by a fault-controlled northwesterly trending ridge (Costa Stenda) halfway up the mountain (Fig. 53). The front of the debris lobe was propelled to the far side of the valley and there probably blocked the flow of the Adige River.

Stands of black pine, vineyards, dilapidated stone walls, and encroaching urbanization today have muted its former forbidding bleakness. Minor recent rockfall activity along the southern sector of the lasta does not threaten buildings or structures along the highway which winds along the foot of the mountain.

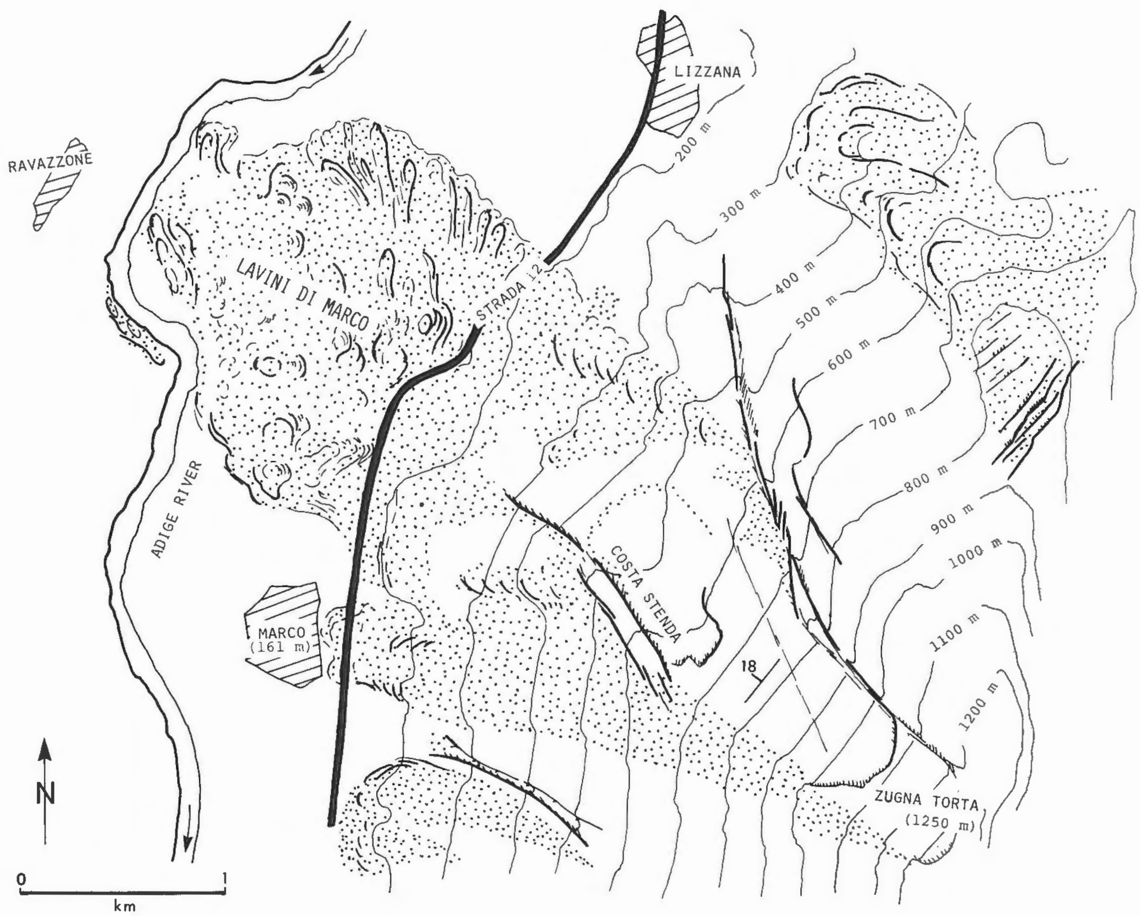


Figure 53: Sketch map of the Lavini di Marco. Detachment surface of the slide mass is a composite carbonate dip slope along a homocline dipping about 18° to the northwest.

Lienz (A5)

Location: Lienz, Osttirol, Austria (H2)
Date(s): 1113 (?) (also spring 1787, 3 September 1965, 6 November 1966)

The city of Lienz (680 m) occupies a depression between a steep debris cone and a gentle fan at the confluence of the Isel and Drau rivers (Fig. 54). This modern capital city of Osttirol developed from several precursor settlements located on or near the cone-fan complex of the Schleinitzbach, Zauchenbach, and Debant torrents. These torrents drain basins underlain by gneiss and mica schist of the Austroalpine basement complex. Uplands reach heights of almost 3000 m and bedrock slopes show evidence of sagging and slumping. Lower mountain slopes have extensive Pleistocene colluvial veneers and ice margin terraces.

It is probable that the lower segment of the Schleinitzbach cone once hosted a Roman or early medieval community by the name of Luenzina. About 1113, a massive slope failure in the steep upper parts of the catchment basin triggered debris movements onto the cone that lasted for more than 100 years. Luenzina was engulfed and destroyed by one of the early flows (Strele, 1936, p. 126-127). A chapel honouring St. Helena was later built on a rock spur overlooking the cone; there were annual processions to the chapel for centuries thereafter, reflecting the unease with which people of the region looked towards the mountain. Today, two harmless creeks cross the bulging Schleinitzbach cone and the community of Oberlienz spreads far up its western sector. A few blocks of gneiss protrude through the intensely cultivated surface of the cone.

The Debant fan, southeast of Lienz, is much gentler than the Schleinitz cone. It has been a site of active debris accumulation only in recent centuries, mainly in response to intensified erosion of colluvial embankments along the Debant Torrent. The upper segment of the fan used to host the Roman station of Aguntum (Fig. 55). Ruins of this impressive town were visible until the 16th century. Then, possibly in response to extensive clearcutting in the uplands,

Figure 54: Index map of the surroundings of Lienz; the huge debris cone on the west probably originated by massive bedrock failures on metamorphic upland slopes along the crest of the Schleinitz Massif. The Debant fan on the east owes its origin to more frequent but less voluminous debris floods and flows triggered by failure and scour of embankments cut in relict surficial deposits.

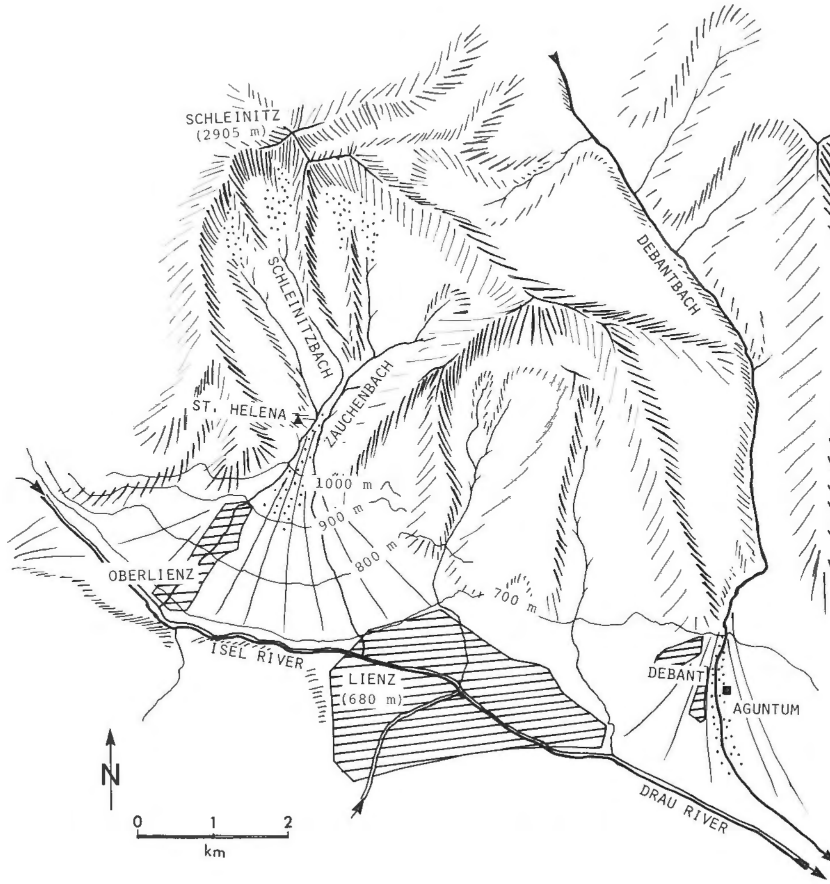


Figure 55: The recently excavated ruins of the Roman town of Aguntum near the axis of the Debant fan. The strip of forest in the background indicates the position of the dyked braided channel of the Debant Torrent which is now several metres above the level of the Roman streets. (GSC 204165-O)

more than 4 m of debris accumulated on the fan, completely burying the Roman buildings. In the spring of 1787 a large debris flow threatened two hamlets astride the fan (Sonklar, 1883, p. 89). During the great rainstorms of 3 September 1965, and 6 November 1966, the Debant fan received $0.2 \times 10^6 m^3$ and $0.15 \times 10^6 m^3$ of debris respectively (Kronfellner-Kraus, 1975).

The channel of the Debant Torrent is now confined by masonry dykes 2 m high, leaving plenty of lateral play for the torrent, which, during debris floods would otherwise threaten the growing suburban community of Debant.

Masière de la Vedana (A6)

Location: Belluno, Veneto, Italy (H2)
Date(s): 3 January 1117 (also 29 June 1873)

The southern front ranges of the Alps near Belluno are composed of Mesozoic carbonate rocks of the South Alpine cover complex. The main bedrock structures in this region are folds and thrust faults which trend east-northeasterly; they are transected by high-angle north-trending fault zones (Fig. 56). The courses of the Piave River and its tributaries are controlled by these bedrock structures. The region is also within a broad east-northeasterly trending belt of historical seismic activity.

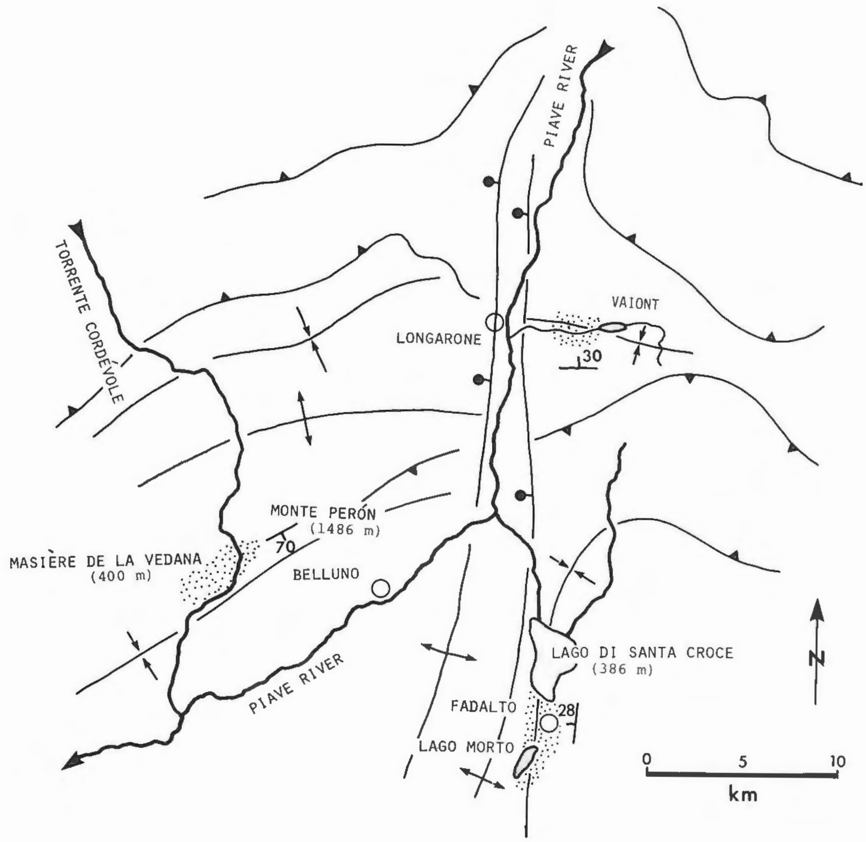


Figure 56: Index map of the Belluno region, showing major synclines (converging arrows), anticlines (diverging arrows), and thrust faults (barbs on the upthrown panel). The narrow gorge of the Piave River and its southward extension towards Lago di Santa Croce are controlled by high-angle transverse fault zones. Three major prehistorical to historical rock avalanche lobes are indicated by the dotted pattern.

On 3 January 1117, a major earthquake, felt throughout the eastern Alps, occurred in the Veneto-Trentino region. Intense shaking seems to have released numerous rockfalls and probably also rock avalanches (Schorn, 1902, p. 9-13). It is possible that during this earthquake two communities, Cornia and Cordova, in the area known as Masière de la Vedana, were destroyed by a rock avalanche (Montandon, 1933, p. 283-284). The extensive debris lobe of the Masière exceeds 100 $\times$ $10^6 m^3$ in volume and consists of angular slabs of carbonate rocks derived from the south face of Monte Perón (1486 m). The spoon-shaped rupture surface is a composite fracture-bedding plane in thick-bedded carbonate rocks that dip approximately 30 to 40^0 to the south. Remaining lateral bedrock spurs guided the movement of the gigantic debris stream.

Another major dip slope in the region, which probably has been a source of recurrent rockfalls during earthquakes such as that in 1117, rises along the southeast shore of Lago di Santa Croce (386 m) at an average angle of 28° to a ridge 1400 m high. Most of this slope is underlain by thick-bedded, west-dipping carbonate strata. Prehistorical slide masses with a total volume of approximately 200 $\times$ $10^6 m^3$ have impounded the lake and rerouted regional drainage to the north. During the most recent damaging earthquake in the Belluno area on 28 June 1873, several small rock avalanches broke away from this cliff (Schwinner, 1912). The northernmost section of this dip slope, rising on the east shore of Lago di Santa Croce, has not yet failed; a highway tunnel paralleling the trend of the strata pierces the toe of this slope (Fig. 57). Some difficulties were encountered during construction.

Many other dip slopes and fracture-controlled ravines along scarp faces in the region have been the focus of recurrent rockfalls. In spite of potential local hazards many talus cones below bedrock cliffs have been built over, like the debris lobes of Masière de la Vedana and Lago di Santa Croce.

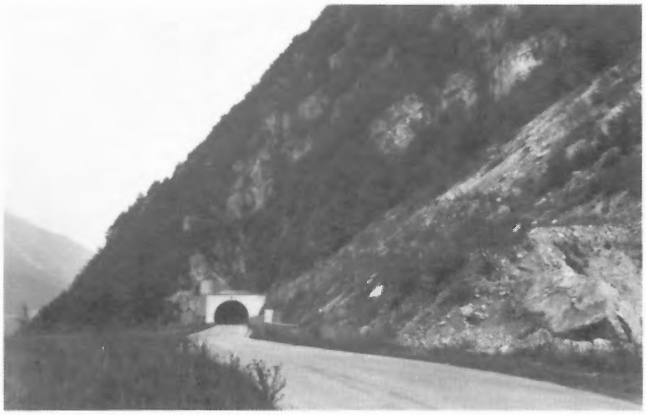


Figure 57: View to the north along the shore of Lago di Santa Croce. A recent highway tunnel pierces the northern flank of the Fadalto slide scar following the strike of the dip slope in massive carbonate rock. (GSC 204165-P)

Semsales (A7)

Location: North of Vevey, Fribourg, Switzerland (B2)
Date(s): 13th century (also 1880)

The village of Semsales (850 m) is located south of the apex of a large debris fan that arises eastward from a gently rolling foothill valley into the deeply dissected Niremont Massif (1513 m) of the Swiss Alpine front ranges (Fig. 58). The range is underlain almost entirely by argillaceous flysch formations of the Helvetic cover complex.

In the 13th century the old village of Semsales was annihilated by a large debris flow which had become detached from a slope failure in the uplands along the Mortivue Torrent. The village was later rebuilt south of its original site. However, debris flows continued to sweep across the cone and in 1880 parts of the village were once again laid waste. Since that time six different projects have been carried out to counteract erosion of unstable embankments by the torrent. The slow sustained flowage of the slopes bordering the torrent accentuated by intermittent sudden failure of the flysch terrain during violent rainstorms is very difficult to control by check dams; the dams therefore have to be replaced from time to time (Getaz, 1977).

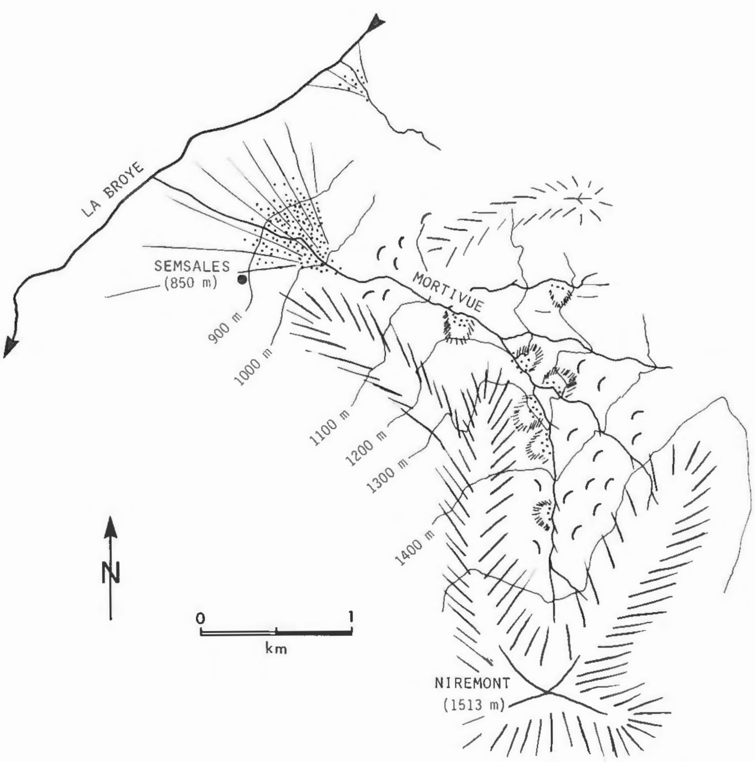


Figure 58: Index map of the Mortivue basin showing the approximate extent of local instabilities (dots) and areas of deep-seated creep (crescents) in argillaceous-sandy bedrock.

Mottec (A8)

Location: Val d'Anniviérs, Valaise (Wallis), Switzerland (C2)
Date(s): 13th century

Val d'Anniviérs rises from the Rhone River southward into the high ranges of the Pennine core zone of the Swiss Alps. Drained by the north-flowing Navisence Torrent, this narrow valley is bordered by steep bedrock slopes composed of south-dipping gneisses and schists. Several high-gradient debris cones in Val d'Anniviérs attest to the action of sporadic rockfalls, debris flows, and snow avalanches. Most of the settlements cling to protected bedrock ledges along otherwise inhospitable slopes.

Sometime in the 13th century a mass of schist and gneiss broke away along a near-vertical fracture zone in bedrock at approximately 2200 m elevation below the Corne de Sorebois (Fig. 59). A rock avalanche with a volume of $2 \times 10^6 m^3$ cascaded onto the hamlet of Saledo (1500 m), demolished the settlement, and impounded the Navisence Torrent into a lake which extended some 500 m upstream. The few survivors of the catastrophe left the site and re-established themselves in another part of Val d'Anniviérs (Montandon, 1933, p. 286).

After the catastrophe the lake gradually decreased in size as the Navisence Torrent cut a channel across the blocky debris lobe. The former lake bottom was eventually cultivated by the people of Pralong. On the slide mass itself, the hamlet of Mottec was established and has grown in recent years by the addition of several vacation homes (Fig. 60). The valley upstream from the slide mass presently is being used as a desilting basin for a major hydroelectric water intake. Behind the bedrock spur east of the breakaway zone linear ridges and depressions indicate incipient instability; no permanent dwellings exist directly below this cliff.

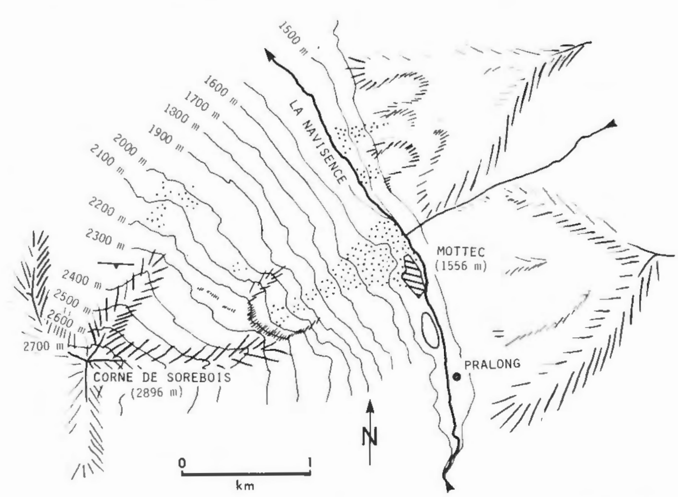


Figure 59: Sketch map of a section of Val d'Anniviérs in the vicinity of Mottec, showing the deposits and detachment zone of the historical rock avalanche below the Corne de Sorebois spur.

Figure 60: View of the hamlet of Mottec; blocky wedge-shaped rock avalanche deposits in the background. (GSC 204165-Q)

Plaine d'Oisans (A9)

Location: L'Oisans, Isère, France (B3)
Date(s): 14 September 1219 (also 10 August (?) 1191, 4 August 1465, 7 August 1612, 18 July 1666)

Near Bourg d'Oisans (720 m) the Romanche River occupies a broad valley flanked by steep bedrock slopes composed of sedimentary and metamorphic rocks of the Helvetic (Dauphinois) basement and cover complexes (Fig. 61). Prior to the 11th century a shallow lake (Lac de St. Laurent) occupied most of the valley from Bourg d'Oisans to Aveyna, where it drained across a bedrock sill. Farther west is the gorge of the Romanche River, cut into metamorphic rocks. A few hundred metres west of Aveyna, the debris cones of two high-gradient torrents, the Vaudaine on the north and the Infernet on the south constrict the river. The catchment areas of both torrents encompass bare cliffs of amphibolitic bedrock rising to elevations between 2000 and 2500 m. From these heights rockslides used to sporadically descend onto the debris cones crossing the channel of the Romanche and thus raising its bed to the level of the outlet.

On 10 August (?) 1191, a large rock avalanche-debris flow elevated the narrow channel of the Romanche to a height above that of the bedrock sill at the outlet of Lac de St. Laurent. Soon the lake rose 10 to 15 m and lapped against the lowest buildings of Bourg d'Oisans. In subsequent years the Romanche failed to deepen its channel across the blocky cones and the people of the Oisans became accustomed to the new lake level. Then, on 14 September 1219, possibly during an autumn rainstorm, the downstream face of the debris barrier was deeply eroded, leading to the collapse of most of the slide mass. A gigantic wave of water and debris swept into the gorge below, carrying away most of the small settlements along the river. When the floodwaters fanned across the city of Grenoble, 30 km downstream, aggrading debris shifted the channels of both the Romanche and Isère rivers. A large part of the city was flooded and a surge of water continued as far as the Rhone delta on the Mediterranean Sea. Casualties reached into the thousands.

From that time onward, the inhabitants of the Oisans district closely watched Lac de St. Laurent. In 1449 debris flows again choked the outlet of the lake, and on 4 August 1465, after a catastrophic cloudburst and debris flows onto the cones, enough material accumulated to once more back up the river over the Plaine d'Oisans, but subsequent deepening of the channel seems to have been harmless. The spectacle of debris flows, rising cones, and an expanding lake repeated itself on 7 August 1612, and again on 18 July 1666. However, at this time cause and effect of the inundations were understood, and a channel was quickly excavated across the cone before the impounded waters could either damage the land upstream or surge into the gorge below.

In the last three centuries the Romanche River and its upstream tributaries have gradually converted the lake bed into a swampy valley flat breached by braided fluvial channels. Beginning in the 19th century a system of dykes and drainage ditches were constructed, changing most of the plain into fertile agricultural land; nevertheless, the threat of serious flooding remained far into the 20th century (Allix, 1929, p. 28-31; Mougin, 1931, p. 39-42).

Today, rock faces and upland ravines still supply debris to the uninhabited cones of the Vaudaine and Infernet torrents; regular snow avalanches prevent a lasting advance of vegetation into the upland basins. Only a major rock avalanche-debris flow would raise the level of the cones to the extent that a new lake could form and thus threaten communities in the gorge of the Romanche River below.

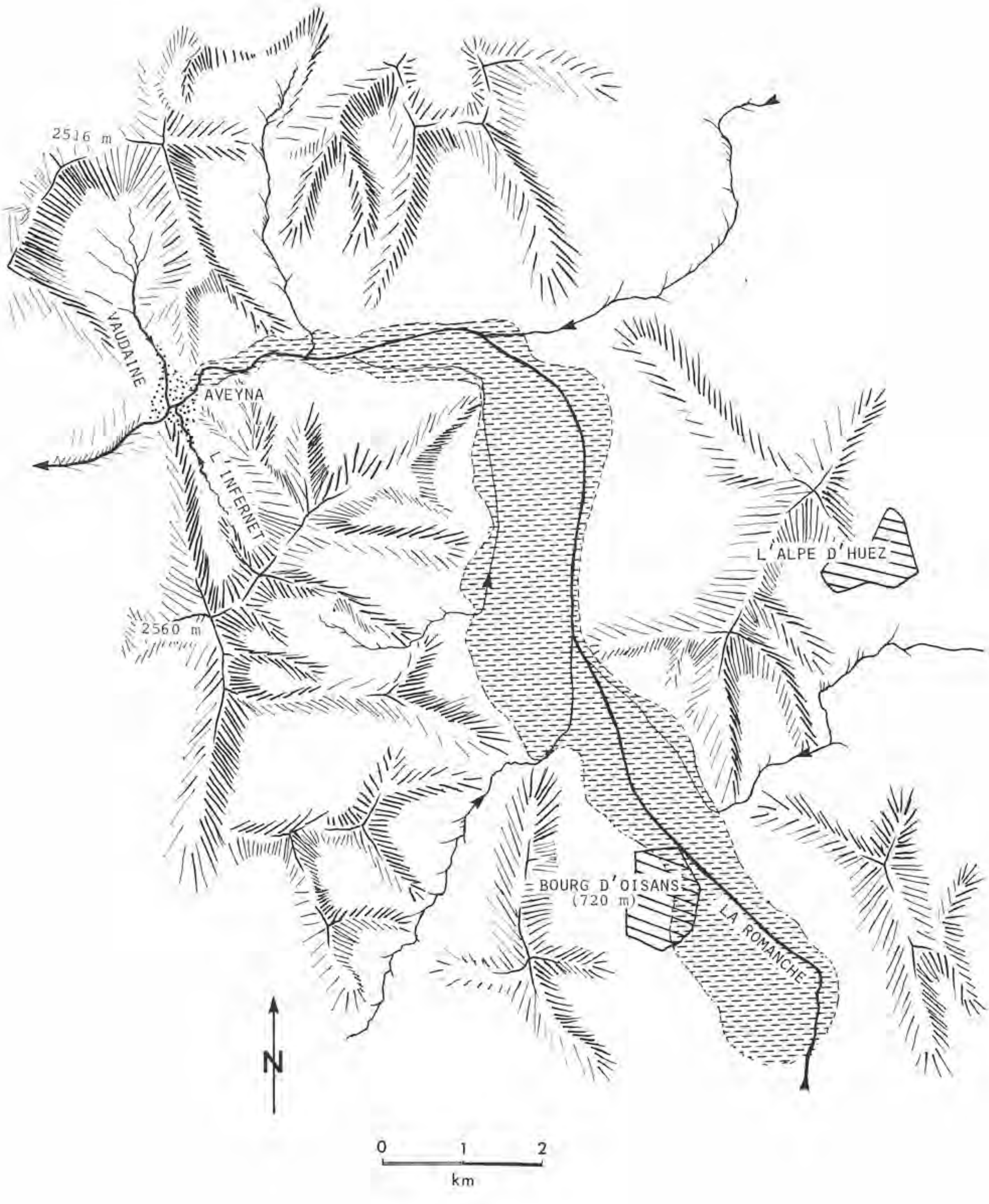


Figure 61: Sketch map of the Romanche Valley near Bourg d'Oisans showing the approximate outline of the former Lac de St. Laurent which formed behind the debris accumulations on the cones of the Vaudaine and Infernet torrents.

Mont Granier (A10)

Location: 10 km south of Chambéry, Savoie, France (A3)
Date(s): 24 November 1248

The prominent Mont Granier Massif (1933 m) is part of the calcareous front ranges of the French Alps. The northern prow of this bedrock ridge rises abruptly from a low saddle northeast of the Isère River (Fig. 62). From bottom to top, the north wall of Mont Granier exposes a recessive calcareous shale unit, a ledge of siliceous limestone, another calcareous shale, and the cliff-forming Urgonian limestone. These units belong to the Helvetic (Dauphinois) cover complex. Bedding along the cliff dips 10 to 15° north-northeast.

On 24 November 1248, following a period of heavy rains, the northern segment of the mountain failed along a surface defined by bedding of the lower calcareous shale and a near vertical fracture zone of the backwall. Approximately $500 \times 10^6 m^3$ of rubbly debris spread as far as 7.5 km to the northeast, covering an area of 15 to 20 km^2 (Fig. 62). The rock avalanche deposits, characterized by a central and frontal zone made up of huge blocks of the cliff-forming Urgonian limestone, buried the town of St. André and possibly as many as 16 hamlets in the surrounding countryside (Goguel and Pachoud, 1972). Estimates of the number of casualties range from 1500 to 5000 (Guillomin, 1937, p. 586-587).

The hummocky surface of the slide deposits, known as Abîmes de Myans, has been greatly modified by human activity during recent centuries. The community of St. André, several small hamlets, surrounding vineyards, and vacation homes interlace with lakes and protruding knobs of Urgonian limestone. The steep breakaway wall (Fig. 63) which remains precariously unstable has been a locus of several rockfalls; one of the largest occurred on 6 June 1953. The zone immediately below the main historical rupture surface has been left undeveloped and is covered by forest.

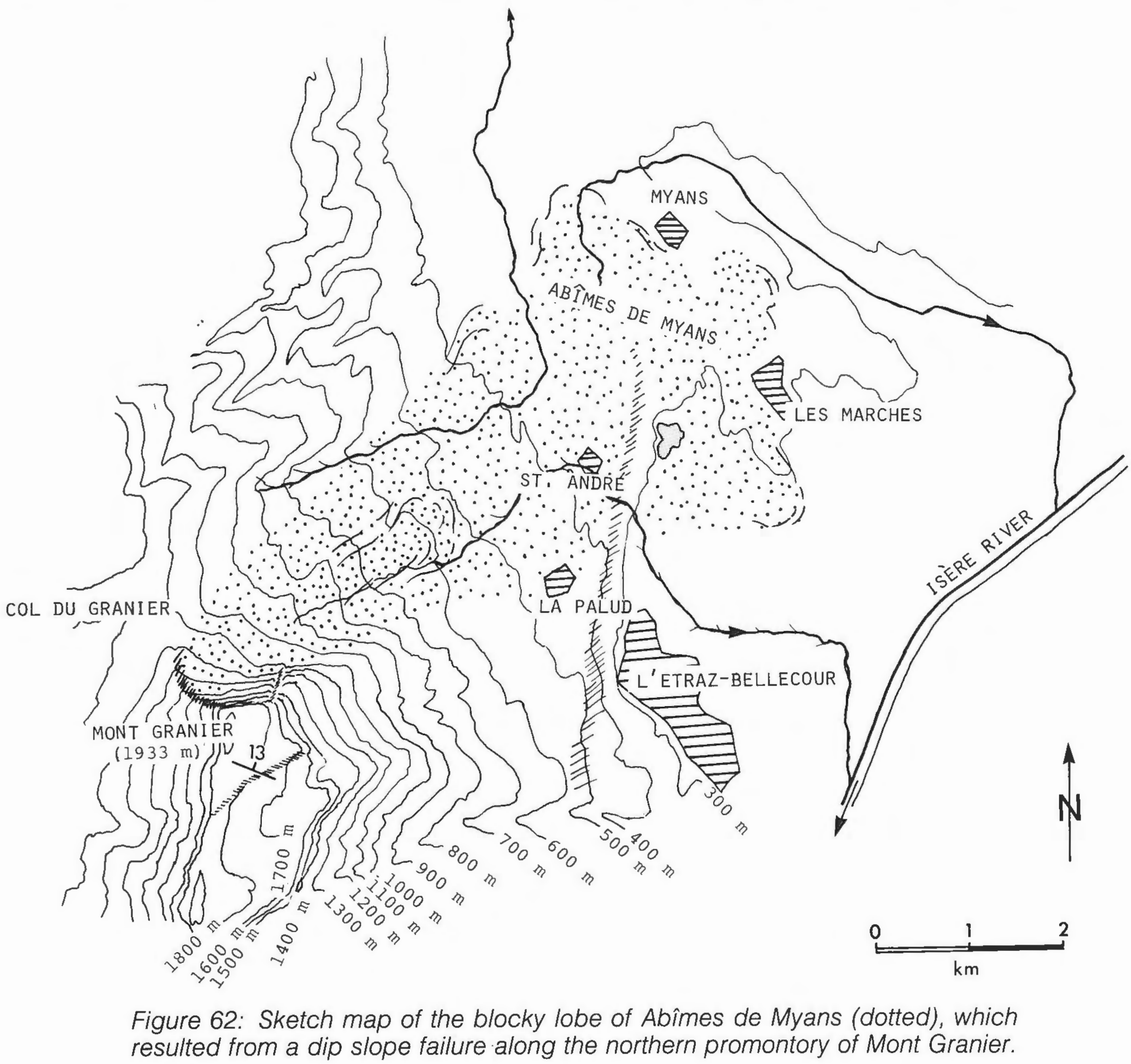


Figure 62: Sketch map of the blocky lobe of Abîmes de Myans (dotted), which resulted from a dip slope failure along the northern promontory of Mont Granier.

Figure 63: View of the joint-controlled vertical headwall of the Mont Granier rockslide detachment zone. (GSC 204165-R)

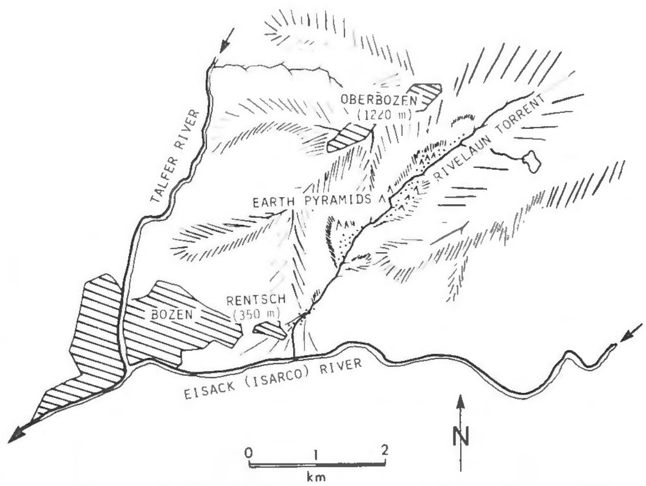


Figure 64: Index map of the Rivelaun basin east of Bozen (Bolzano) showing the location of the Rivelaun cone and the earth pyramids in the source area of the Rivelaun Torrent.

Rentsch — Rencio (A11)

Location: Bozen (Bolzano), Südtirol (Alta Adige), Italy (G2)
Date(s): 13 July 1327 (also between 1215 and 1321, 24 August 1806, August 1957)

The small community of Rentsch (350 m) nestles against the western flank of the debris cone of the Rivelaun Torrent, just above the confluence with the Eisack (Isarco) River (Fig.64). The Rivelaun Torrent, like several other tributaries of the Eisack River in the region, originates by the confluence of small tributary creeks below the escarpment of an extensive rhyolite plateau of the South Alpine basement complex. Along the rim of the plateau, at elevations between 1000 and 1200 m, pre-Pleistocene topographic depressions are filled with semi-indurated morainal deposits of Pleistocene age. Where the torrents drop from the plateau into steep bedrock gorges, moraines and relict colluvium are deeply dissected. Along the upper Rivelaun Torrent morainal terraces have been sculptured by erosion into the famous earth pyramids of Bozen (Fig. 65). The cone at the mouth of the torrent consists largely of debris from the gullies carved between the pyramids.

Local chronicles report that between 1215 and 1321 the village of Rentsch suffered from violent hailstorms, torrential rains, and debris flows. On 13 July 1327 the settlement finally disappeared under a blanket of bouldery rubble (Strele, 1936, p. 128). It is probable that excessive clearcutting on the upland plateau in the vicinity of Oberbozen (1220 m) initiated this devastating cycle of erosion and mass movement. Damaging debris flows also invaded Rentsch on 24 August 1806, after a long and rainy summer, and in August 1957 (Stacul, 1979, p. 91).

Figure 65: Earth pyramids carved from Pleistocene morainal deposits along the flanks of the Rivelaun Torrent. View downstream along the torrent. (GSC 204165-S)

Today the Rivelaun cone hosts several old estates dispersed among highly productive vineyards. The straight channel of the torrent has been fitted with masonry linings, sills and check dams. Meadows, farmland, and forests surround the much visited earth pyramids along the escarpment near the thriving agricultural and tourist community of Oberbozen.

Dobratsch (A12)

Location: Gail Valley, Villach, Kärnten (Carinthia), Austria (12)
Date(s): 25 January 1348

The imposing scarp face of the Dobratsch Massif (2000 m) soars above the north bank of the Gail River (500 m), 5 km west of the town of Villach (Fig. 66).The Dobratsch is composed of intensely fractured carbonate strata which overlie a succession of recessive calcareous shales and redbeds of the Austroalpine cover complex. Bedding dips 30 to 60° to the north. During late Pleistocene downwasting of the Gail Valley glacier large sections of the carbonate cliff lost their footing and collapsed; a debris lobe of approximately 300 × $10^6 m^3$ spread across stagnant glacier ice near the present community of Arnoldstein. Complete melting of this ice produced hummocky terrain characterized by numerous conical mounds, known as 'Alte Schütt' (Abele, 1974, p. 110-112).

On 25 January 1348, a strong earthquake flattened the town of Villach and several neighbouring villages. The earthquake occurred without significant foreshocks and apparently was centred below the Periadriatic Lineament (Gail Line), a major high-angle fault following the Gail Valley. Loss of life in the communities of the region was heavy and in part a result of rockfalls and avalanches from the Dobratsch. Shaking during the earthquake caused several rock slabs to break away from the Dobratsch scarp face. The largest of these wedges (30 × $10^6 m^3$) failed along a composite fault surface dipping approximately 45° to the south ('Rote Wand' = red wall) and avalanched some 4 km across hummocky terrain underlain by late Pleistocene slide deposits in the Gail Valley. The carbonate rubble blocked the Gail River, flooding several hamlets upstream from the dam. A short time later the Gail River burst across the weakest part of the slide barrier and devastated the valley below. It is probable that the rock avalanche itself buried a few settlements (Till, 1907).

The large number of rockfalls triggered by the 1348 earthquake created a thick cloud of dust which enveloped the region for days. This cloud probably accounts for the fact that villages flattened by the earthquake were initially thought to have been struck by the Dobratsch rock avalanche; snow avalanches may have compounded the confusion. The dust cloud was also blamed by some for the outbreak of the plague, which, precisely at this time, reached central Europe.

Today the main rock avalanche lobe of 1348 is known as 'Junge Schütt' (= young debris) or 'Steinernes Meer'

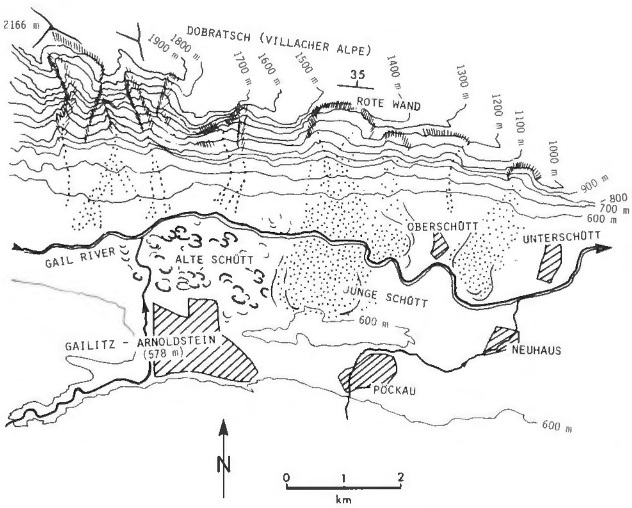


Figure 66: Sketch map of the Dobratsch scarp face and rock avalanche deposits related mainly to the 1348 earthquake (dots). The faulted carbonate strata dip approximately 35° to the north. Alte Schütt is a rock avalanche deposit originally laid down on the downwasting late Pleistocene Gail Valley glacier; Junge Schütt denotes the Rote Wand rock avalanche of 1348.

a

b

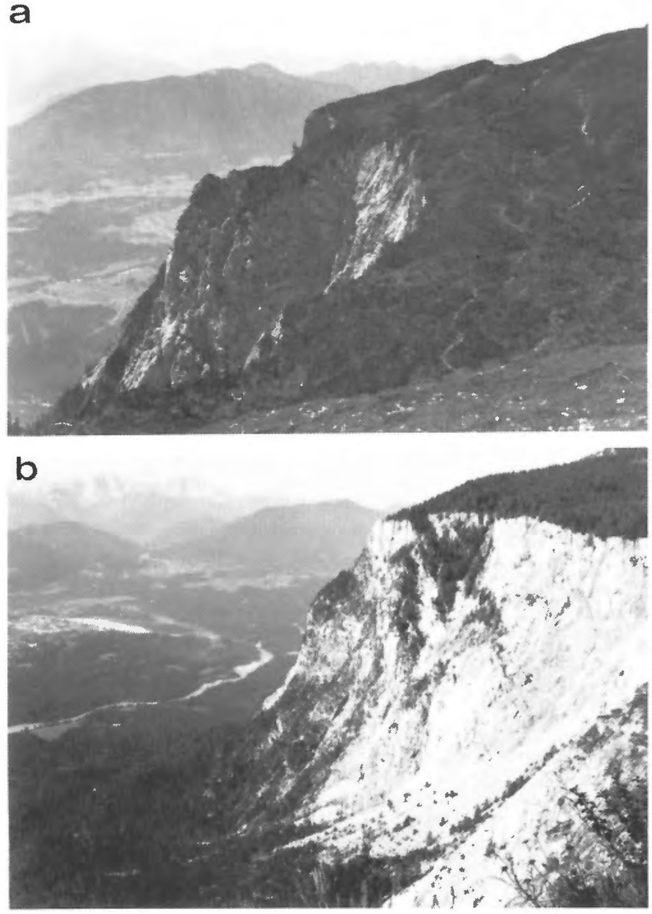


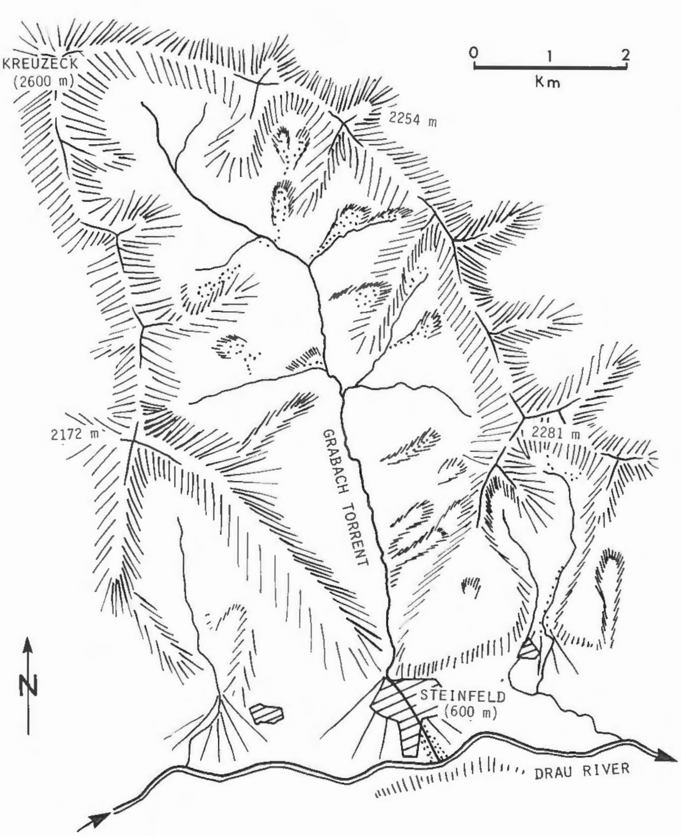
Figure 67: a) Fracture-controlled linear depression behind the crown of the late Pleistocene failure zone along the Dobratsch south face (GSC 204165-T). b) Detachment zone of the Rote Wand rock avalanche of 1348; note intense fracturing of the carbonate rocks which dip to the right (GSC 204165-U).

(=stony sea). Two hydroelectric installations exploit the hydraulic head created by the debris lobe across the Gail River. The prehistorical slide deposits of the 'Alte Schütt' are being progressively invaded by the expanding communities of Gailitz-Arnoldstein, Oberschütt, and Unterschütt (Fig. 66). New housing developments have advanced steadily in the direction of the highest cliff of the Dobratsch south face. Behind this face are several large depressions, aligned sinkholes, and gaping fissures which indicate that a future failure of this face is likely (Fig. 67).

Steinfeld (A13)

Location: Drau Valley, Kärnten (Carinthia), Austria (I2)
Date(s): 1365 (also 1323, June 1827, 1848, 2 November 1851, 5 November 1966)

In the area of Steinfeld (600 m) the channel of the Drau River is constricted by several conspicuous debris cones and fans (Fig. 68). The largest are associated with south-flowing torrents originating in the uplands of the Kreuzeck Massif (2600 m). The bedrock of this mountain range is composed of mica schists and gneisses of the Austroalpine basement complex. Foliation and schistosity generally dip to the north. Bowl-shaped catchment basins at elevations between 1200 and 2000 m have a discontinuous mantle of late Pleistocene relict colluvium characterized by deep erosional gullies and slump scars. Below these debris sources torrents plunge into steep-walled bedrock gorges and then emerge onto the aprons of debris which host the communities of the Drau Valley.



During the 14th century the community of Schönfeld (=fine fields), became a regional administrative centre, spreading over the cone of the Grabach Torrent. Beginning in 1323, and probably due to deforestation of the uplands related to extensive mining activity, severe debris flows issued from the gorge of the torrent. In 1365 several pulses of debris laid waste to the whole town. According to a legend, many people were killed. The abandoned site of Schönfeld became known as Steinfeld (=stone field). For 500 years oral traditions of the bursts of debris from the Grabach Torrent kept the fan free of any kind of building. However, eventually a new community, named Steinfeld began to grow on the ruins of Schönfeld (Fig. 68).

The winter of 1826-27 brought exceptional snowfalls to the mountains of the region and spring was delayed by months. An abrupt snowmelt in June 1827 reactivated many erosional scars in the uplands of the Drau region. Then, during a succession of relatively wet years between 1840 and 1850, thick debris lobes blocked the channel of the torrent above the narrow bedrock gorge. In 1848, a local cloudburst mobilized this debris, causing severe damage to buildings in Steinfeld. On 2 November 1851, a regional rainstorm onto frozen ground deluged the uplands of the Drau Valley; most of the torrents carried debris flows of massive proportions and Steinfeld was laid to waste again (Strele, 1936, p. 133).

Towards the end of the 19th century, check dams were erected across tributary ravines of the upper Grabach Torrent (Fig. 69a). Many of the works proved beneficial for decades and even fulfilled their function during the great storm of 3-6 November 1966. Although the impact of debris flows in 1966 was serious on other cones in the valley and several lives were lost, there were only minor pulses of debris down the Grabach Torrent and these were channeled through Steinfeld by two dykes without inflicting much damage.

In the years since 1966, new protective and control structures have been added to the upland basin of the Grabach Torrent (Fig. 69b) and observation points have been established along the gorge to be occupied during potentially dangerous storms. Recently, runoff from the Grabach basin has been diverted through long tunnels towards the Möll River hydroelectric complex. The town of Steinfeld once again prospers (Fig. 69c).

Figure 68: Index map of the Grabach basin and the debris fan of Steinfeld; debris sources are mainly shallow scars in relict colluvium.

Figure 69: a) Old stone-masonry check dams in the upper Grabach basin, rendered ineffective by slumping of surficial deposits along the torrent embankment (GSC 204165-V). b) Modern concrete check dam in the Grabach channel; note gently rising wings and armoured discharge section which allows passage of minor debris flows (foreground) while confining intermittent storm runoff to the centre of the channel (GSC 204165-W). c) Dyked and walled channel of the Grabach Torrent on the fan of Steinfeld; note the forested protected zone at the confluence of the torrent with the Drau River and the uninhabited area of the lower fan which would be inundated by the river in the event of its blockage by debris flows (GSC 204165-X).

Ziano (A14)

Location: Predazzo, Val di Fiemme, Trentino, Italy (G2)
Date(s): 15th Century (?)

The village of Ziano di Fiemme (950 m) borders a steep talus slope at the foot of a steep carbonate cliff on the north bank of the Avisio Torrent west of Predazzo (Fig. 70). Above the town the carbonate strata, which are part of the South Alpine cover complex, dip gently north and are dissected by ravines which also cut into the rim of a smooth upland basin underlain by volcanics. Just east of Ziano a huge bouldery cone known as Mosene (= debris), projects southward from the generally dry ravine of Val Bonetta. Towers of carbonate, detached along vertical fractures, lean into the narrow gorge, and accumulations of rockfall rubble indicate sporadic collapse of these spires. Open fissures in the mountains above Ziano have been attributed to earthquakes (Dalla Torre, 1913, p. 333).

Excavations on the Mosene cone have unearthed ruins of prehistorical and Roman settlements. A tradition relates that in the remote past the town of Cornelijan (or Cunelian) was completely swallowed up by a rock avalanche that fanned over the Mosene cone (Gorfer, 1977, p. 583).

The cone has been avoided as a building site for many centuries in spite of the fact that Ziano has had to contend with repeated floods of the Avisio Torrent. In recent years a solid protective concrete dam, approximately 10 m high, has been constructed across the mouth of the Val Bonetta ravine (Fig. 71), and the first vacation homes have sprung up below. The old village of Ziano faces another more direct threat: a precariously tilted pinnacle of dolomite which has separated from the main rock wall hangs over the town. A large cross has been erected on the top of this spire.

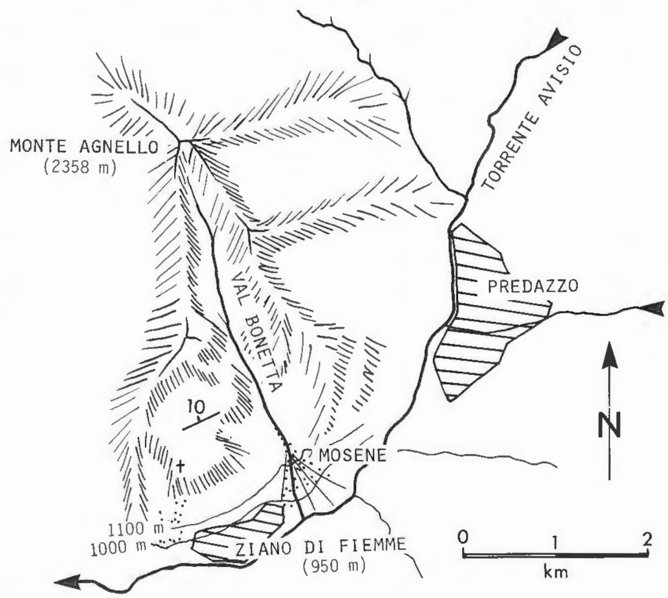


Figure 70: Sketch map of the Mosene cone and the Val Bonetta ravine which is flanked by steep carbonate walls; rockfall zone behind the village is indicated by a cross.

Figure 71: Protective steel-concrete dam at the mouth of Val Bonetta. The lined channel permits nondestructive discharge of small debris floods spilling over from major flows that would be retained behind the structure. Note drainage bores at the base and selective discharge section ('filter dam') at the crest. (GSC 204165-Y)

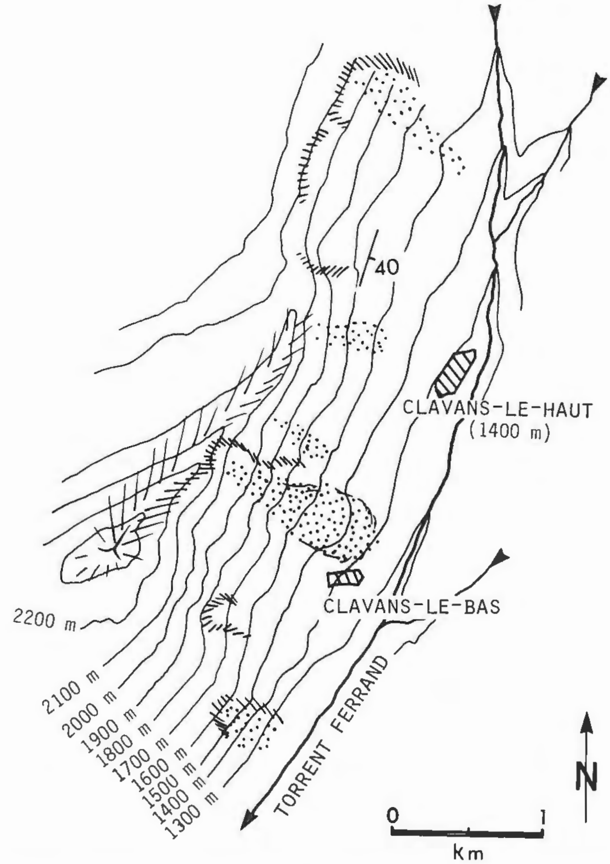


Figure 72: Sketch map of the intensely fractured dip slope above Clavans defining the northwest wall of the Ferrand Valley. Historical slope failure at Clavans-le-Bas is indicated by dots.

Clavans (A15)

Location: Freney d'Oisans, Isère, France (B3)
Date(s): 1418

The two communities that make up Clavans (1400 m) are perched on a narrow terrace above the gorge of the Ferrand Torrent, which follows a north-northeasterly trending strike valley in sandstones and calcareous shale terrain of the Helvetic (Dauphinois) basement-cover transition (Fig. 72). The dip slope above Clavans is underlain by fractured basement gneisses and quartzites dipping approximately 40° to the east.

In 1418 a third of the community of Clavans was destroyed by a rockslide which also caused the death of several people (Allix, 1929, p. 38, 189, and 204). The lobate slide mass, approximately $3 \times 10^6 m^3$ in volume, can still be seen between the two groups of buildings constituting the small settlements of Clavans-le-Bas and Clavans-le-Haut. (Fig. 73). Fresh rockfall deposits on the valley wall behind the two hamlets attest to continued slow deterioration of the bare upland ridge.

Figure 73: View of the slide lobe composed of quartzitic rubble behind the community of Clavans-le-Bas. (GSC 204165-Z)

Ganderberg – Passeier Wildsee (A16)

Location: Passer Valley, Südtirol (Alto Adige), Italy (G2)
Date(s): 22 September 1419 (also 1404, 14 September 1503, 30 September 1512, 21 May 1572, 18 June 1721, 17 September 1772, 22 October 1774, 13 October 1787)

The Ganderberg (2330 m) rises along the east side of the upper Passer Valley (1200 m). The Passer River drains the steep south slope of the Ötztal Massif and joins the Adige (Etsch) River at Meran (Fig. 74). The region is underlain by metamorphic rocks of the Austroalpine basement complex; in the area of the Ganderberg the metamorphic foliation dips steeply to the north. The Ganderberg slope is a sagging rock mass involving as much as 500 to $600 \times 10^6 m^3$ of gneiss, schist, and amphibolite. The westward dipping zone of detachment of the sagging slope follows a composite fracture cutting across the trend of the regional foliation. Halfway up the slope is a distinct scarp that separates the lower part of the sagging terrain from the upper part of the broken rock mass.

In 1404 a large rockslide from the toe of the Ganderberg (or 'Gspellerberg') blocked the flow of the Passer River and created a lake more that 50 m deep and approximately 1km long (Fig. 74). For the almost 400 years of its life this lake was to become known as Passeier Wildsee (= wild lake) or Kummersee (= lake of sorrow). During the early years, the lake overflowed without seriously effecting the stability of the slide mass below the outlet. However, on 22 September 1419, a major section of the debris barrier failed and a flood of boulders, sand, and water burst through the gorge below. At least 400 people died as the surge of debris overwhelmed small settlements along the lower Passer River and swept through the town of Meran, 25 km south of the Wildsee.

Continued movements along the toe zone of the sagging slope maintained the lake. Other catastrophic bursts, generally coincident with regional rainstorms, were recorded on 14 September 1503, when the city walls of Meran were completely demolished, on 30 September 1512, when the city tower of Meran collapsed, and on 21 May 1572, when the reconstructed walls were breached again. Then the fearsome floods ceased for almost 150 years, although the lake remained. On 18 June 1721, a section of the debris dam failed again and $5 \times 10^6 m^3$ of water were discharged through the narrow gash at the outlet. Following this disaster the first plans were made to design a protective structure across the outlet of the lake. However, they were not carried out.

On 17 September 1772, a regional rainstorm deluged the south slope of the Ötztal Massif, flooding most tributary torrents of the upper Passer. The level of the Wildsee rose dramatically and its outlet was plugged by a log jam carried into the lake by raging tributary torrents. It is possible that movement of the Ganderberg slope also accelerated in response to the three days of rain and warm temperatures. In any case, the waters of the lake soon breached the narrow

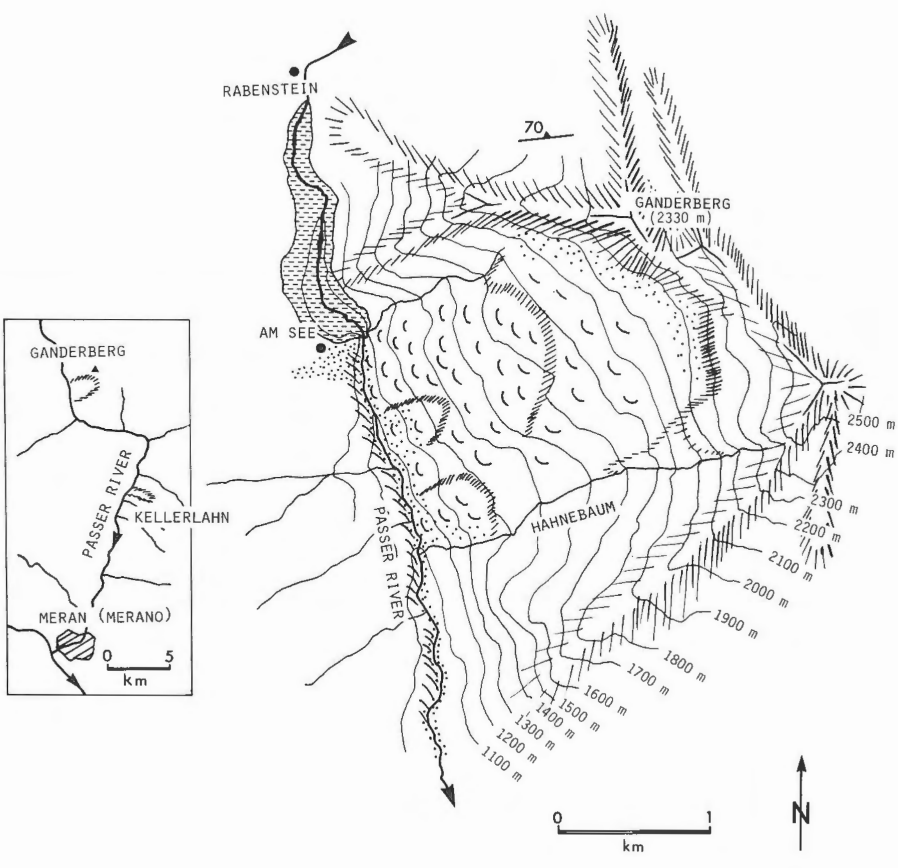


Figure 74: Index and sketch map of the sagging Ganderberg slope in the upper valley of the Passer River. The outline of the historical Passeier Wildsee (= wild lake) is shown by the broken line pattern. Note the large inactive arcuate scarp near the crest of the mountain and smaller active slumps involving fractured bedrock and surficial deposits along the toe zone.

outlet and $3 \times 10^6 m^3$ of water and debris swept down the valley, bringing back memories of earlier tragedies. Shortly afterwards the lake was visited by J. Walcher, a professor of mechanics at the University of Vienna (Fig. 75). Walcher recognized that the only lasting remedy against future debris floods would be to permanently lower the level of the lake (Walcher, 1773, p. 87-96). In 1773 a reinforced drainage channel and flood gates were constructed at the narrowest section of the outlet in order to control runoff during critical periods of sudden snowmelt or rainstorms. However, on 22 October 1774, encouraged by preliminary success with the sluice gates, the supervisor of the operation opened the gates too far and the waters of the rain-swollen lake poured into the channel. As a result the undercut embankments collapsed and the lake drained in 12 hours, spreading death and destruction in the valley below. In Meran, the flood wave demolished several buildings (Staffler, 1846, v. 2, p. 744-745; Sonklar, 1883, p. 87-88; Dalla Torre, 1913, p. 294). After the last of the debris floods the former lake bottom became the only flat stretch of land along the otherwise precipitous gorge of the upper Passer River.

Today the sagging Ganderberg slope is still in motion; the narrow highway that crosses it high above the river bears evidence of this: retaining walls display numerous gaping fissures, 5 – 10 cm wide. The gorge of the Passer River along the toe of the sagging slope continues to close by slumping and an extraordinary amount of blocky bedload is supplied to the river here (Fig. 76a). Concrete check dams and a protective dam have been erected to prevent the river from undercutting its eastern embankment and to retain massive pulses of debris, but slow creep of the whole mountainside is uncontrollable. Bedload in the channel of the Passer River is still very coarse as far as Meran (Fig. 76b). Over the years a great variety of training dykes, masonry linings, and check dams have been added, and their maintenance continues to be a costly problem.

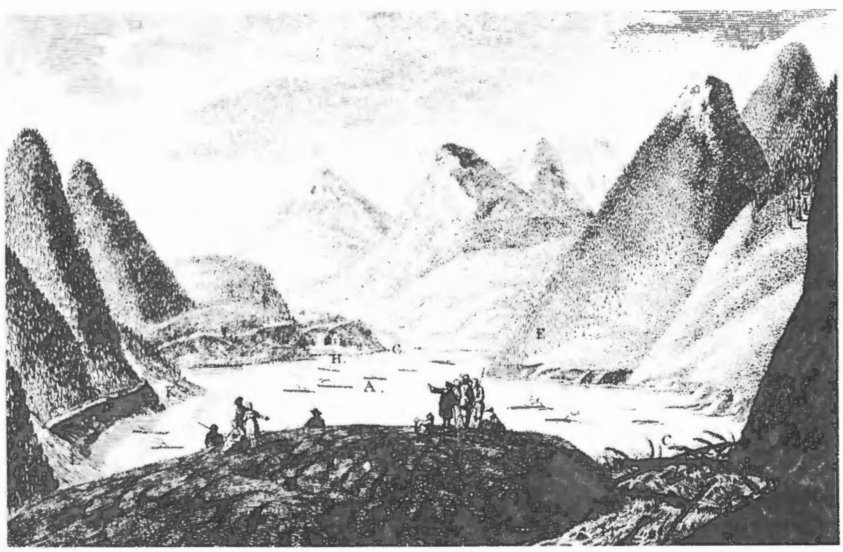


Figure 75: View of the Passeier Wildsee, looking northward, as depicted by Walcher (1773) after his tour of inspection; note shoreline formed during a higher stand of the lake.

Figure 76: a) View upstream along the chaotic toe zone of the sagging Ganderberg slope. The Passer River is eroding into embankments of surficial deposits opposite the sagging bedrock slope (GSC 204166-A). b) Coarse bedload in the channel of the Passer River where it enters the city of Meran (Merano) (GSC 204166-B).

Cervières (A17)

Location: Briancon, Hautes Alpes, France (B4)
Date(s): June 1431 (also 1407, June 1434, 1448)

The village of Cervières hugs a ledge of fractured carbonate rocks and colluvial debris along the north bank of the Torrent de la Cerveyrette, a tributary of the Durance River (Fig 77). At Cervières (1636 m) the channel of the Cerveyrette Torrent is confined by the debris cone of a southern tributary, the Ruisseau de Blétonnet. The floor of the upper Cerveyrette Valley is a swampy alluvial flat which has aggraded behind Pleistocene end moraines at the locality La Chau (1859 m). Between La Chau and Cervières the Cerveyrette Torrent flows in a deep ravine carved into Pleistocene moraine and outwash deposits (Fig. 78). The flanking mountain ridges are composed of carbonate and slate of the Pennine (Briançonnais) facies complex.

During the early Middle Ages practically the whole catchment basin of the Cerveyrette lost its forest cover due to clearing and overgrazing. As a result enhanced peak discharges of the torrent scoured deeply into the moraine and outwash above Cervières. In 1407 a number of buildings in Cervières were invaded by debris for the first time. In June 1431, some 30 to 50 houses were demolished by voluminous debris flows. In June 1434, another part of the settlement, which then was probably much larger than today, was covered by debris. In 1448 20 more buildings collapsed in the onslaught of debris flows (Mougin, 1931, p. 51). It is probable that the present village grew around a cluster of old buildings located on sloping terrain far above the channel of the torrent.

In recent years the channel across the village has been lined by strong stone-masonry revetments.

Figure 78: View of the deeply scoured morainal deposits above Cervières (GSC 204166-C).

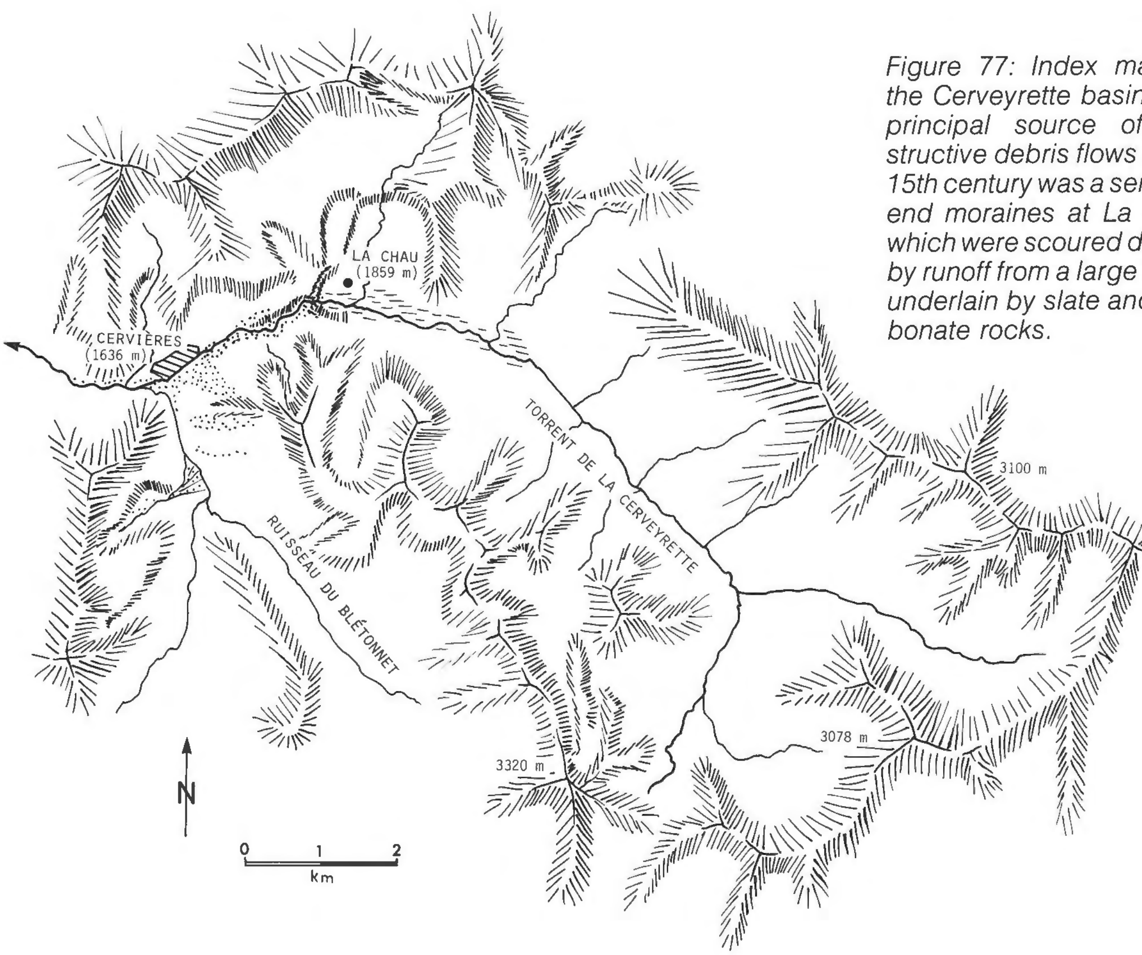


Figure 77: Index map of the Cerveyrette basin. The principal source of destructive debris flows in the 15th century was a series of end moraines at La Chau which were scoured deeply by runoff from a large basin underlain by slate and carbonate rocks.

La Chapelle (A18)

Location: Arc Valley, Savoie, France (B3)
Date(s): 10 August 1431

La Chapelle (430 m) is a small village at the toe of a bulging debris cone above the east bank of the Arc River (Fig. 79). The small creek that crosses the cone is fed by several tributary rivulets which drain a steep bedrock bowl along the west slope of Le Grand Mas (2235 m). The mountainside is underlain by southeast-dipping gneiss of the Helvetic (Dauphinois) basement complex.

On 10 August 1431, a massive debris flow or slide mass completely erased an old village on the cone of La Chapelle (Mougin, 1914, p. 1118). The cirque-like morphology of the upland bowl between Le Gran Mas and Plan Verney suggests that in prehistorical time (?) a major section of the bedrock ridge may have failed along a valleyward-dipping fracture zone. The detached rock masses of the scarp slope eventually gave rise to massive debris flows, leaving behind the concave bedrock bowl and adding a substantial volume of debris to the bulging cone in the valley.

Boulders strewn along the apex of the La Chapelle cone suggest recurrent historical debris flow activity. The broadly convex surface of the cone is now agricultural land. A chapel near the apex of the cone is dedicated to St. Bernard and commemorates the great deluge of 1431.

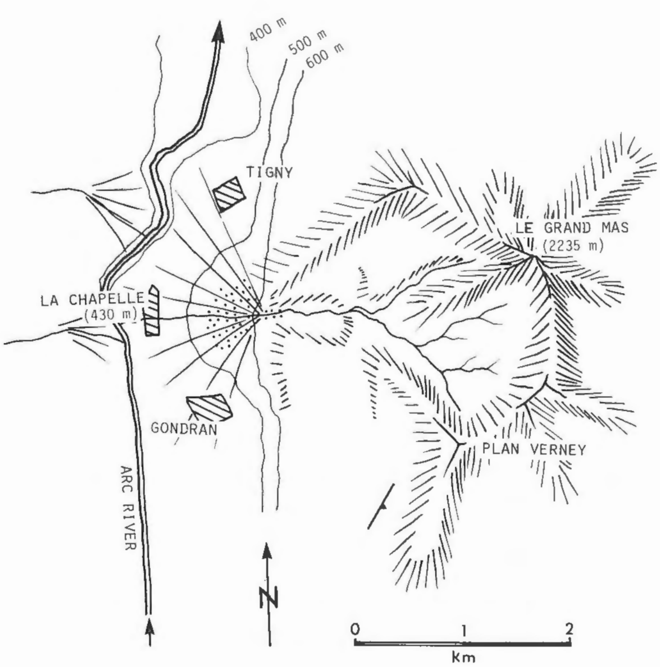


Figure 79: Sketch map of the cone of La Chapelle below the metamorphic bedrock scarp slope of Le Grand Mas; note location of villages between the floodplain of the Arc River and the hazardous apex of the cone.

Zug (A19)

Location: Zug, Switzerland (D1)
Date(s): 4 March 1435 (also 1592 to 1594, 5 July 1887)

The prosperous town of Zug covers most of the delta of the Lorze Torrent at the north end of Lake Zug (Fig. 80). The Lorze Torrent, which crosses unstable terrain of the Molasse Zone of the Alpine foothills, has in the past blocked the gorge by embankment failures, causing sporadic debris floods on the triangular fan delta west of Zug. These floods deposited layers of gravel on top of a poorly consolidated wedge of lacustrine sands, silts, clays, and organic matter.

During the early growth of Zug tall stone buildings, piles, and shoreline installations overloaded the sedimentary wedge underneath the town. On 4 March 1435, the shorefront of the city, consisting of 26 houses (and 60 inhabitants), suddenly slumped into the lake. At the time many strange ideas were advanced to explain this totally unexpected disaster, including one which blamed the fishes in the lake for having nibbled away at the mud of the shoreline!

Subsidence of the shoreline of Zug continued and between 1592 and 1594 several houses were severely damaged. On 5 July 1887, another wedge of sediment amounting to $0.15 \times 10^6 m^3$ slumped into the lake, carrying a row of houses and claiming the lives of several people (Heim, 1932, p. 52).

Following the last of the major failures, subsidence was monitored carefully. Between 1889 and 1944 a downward displacement between 1 and 11 mm/year was measured adjacent to the failure of 1887 (Von Moos, 1948). Today buildings are set back approximately 100 m from the shore, leaving space for an attractive marina and a shorefront promenade. The channel of the Lorze has been brought under control by stone revetments.

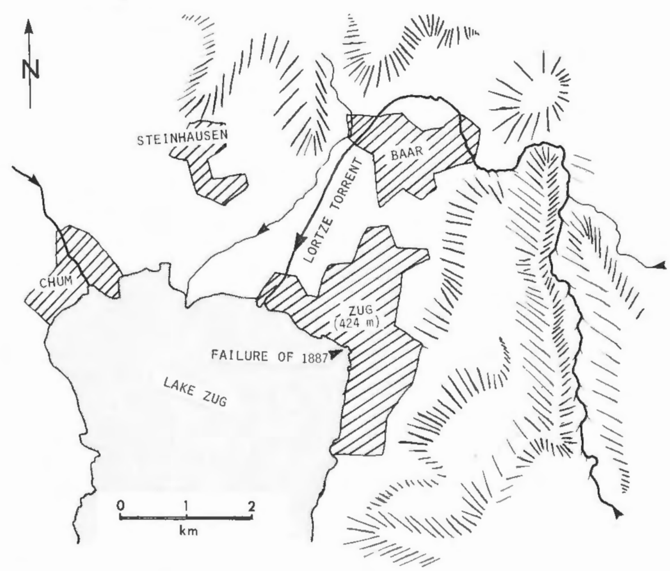


Figure 80: Sketch map of Lake Zug showing the location of the major shoreline failures on the eastern fringes of the Lorze fan delta.

St. Jean-de-Maurienne (A20)

Location: Arc Valley (Maurienne), Savoie, France (B3)
Date(s): 1 February 1439 (also February 1618, 22 December 1740, 31 July 1816, 18 July 1824, 25 July 1862, 20 July 1871, 24 July 1872, spring 1965)

The town of St. Jean-de-Maurienne and nearby communities are perched on debris fans fringing the narrow northwest-trending Arc Valley (Fig. 81). In this part of the Maurienne the Arc River crosses a zone of intensely deformed Helvetic (Dauphinois) and Pennine (Briançonnais) basement gneisses, evaporites, shale, limestone, and flysch, known as 'Faisceau des salins'. This zone is approximately 10 km wide, trends north-northeasterly, and dips to the east-southeast. From the Arc River (500 m) steep ravines, torrent channels, and ridges rise to peaks approximately 2500 m elevation. During the Middle Ages the high catchment areas lost most of their forest cover; bedrock slopes are now unstable; many are characterized by deep-seated creep and incipient slumps. Composite bedrock-debris slumps ('glissements') have lower tongues which launch slow- or fast-moving debris flows ('coullées'). The mechanism of these composite flows has been documented by Malatrait (1975).

One of the largest sagging slopes is the Glissement de Jarrier, west of St. Jean-de-Maurienne (550 m). It is a bowl-shaped depression superimposed onto a dip slope of calcareous slate. The layer of disrupted bedrock and colluvium is approximately 50 m thick and its total volume is $650 \times 10^6 m^3$ (Malatrait, 1975, p. 162-167). Between 1435 and 1439, precipitation in the region of St. Jean-de-Maurienne apparently was abnormally high; debris flows occurred in areas immediately east of the city. On 1 February 1439, during an intense winter rainstorm coupled with unseasonal snow melt, lobes of debris from the Jarrier slide mass displaced the mouth of the Bonrieu Torrent towards the town centre of St. Jean. A massive debris flow which subsequently followed the new channel destroyed many buildings and killed 75 people. The unstable debris lobes in the Jarrier bowl again launched destructive debris flows in February 1618, on 22 December 1740, and on 31 July 1816 (Mougin, 1914, p. 1121-1124; Chabert, 1978).

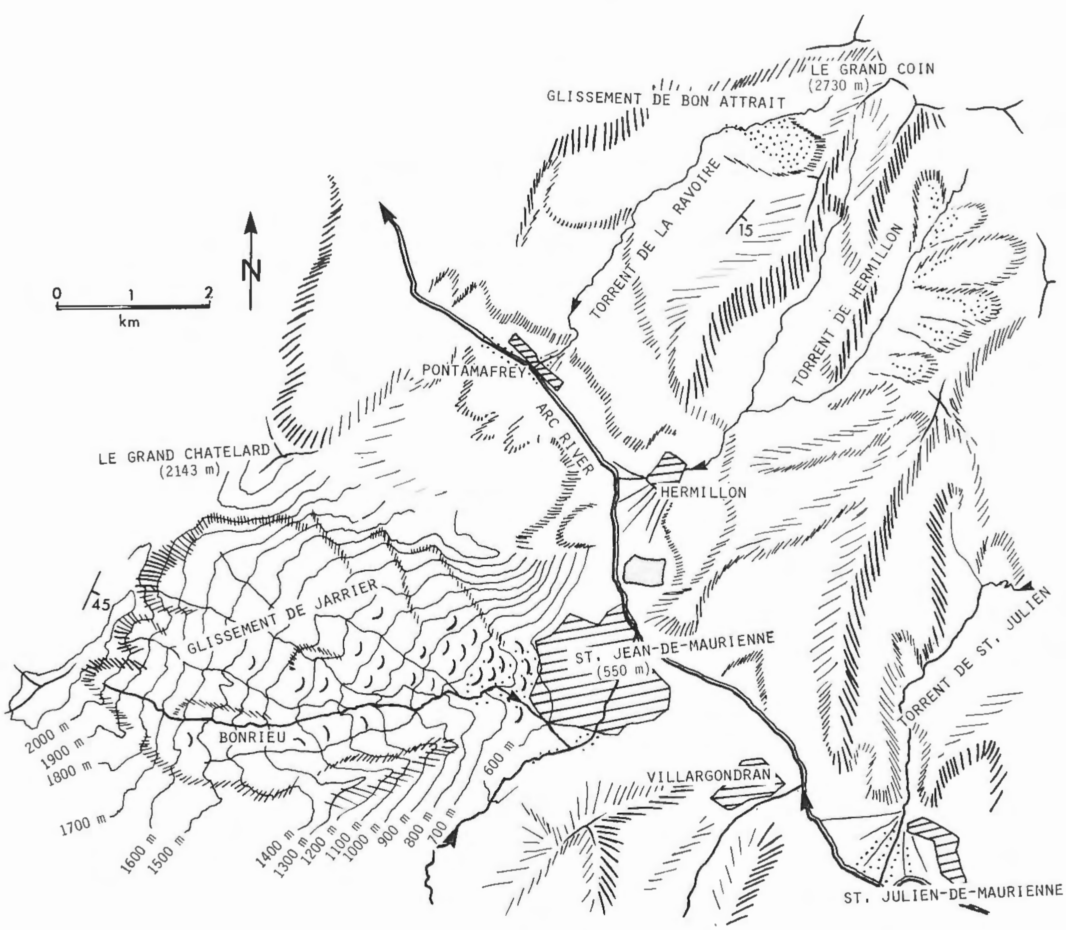


Figure 81: Sketch map of the Arc Valley in the vicinity of St. Jean-de-Maurienne; note the northeasterly structural trend which controls the location of major deep-seated detachment zones such as those of the Glissement de Jarrier (dip slope) and the Glissement de Bon Attrait (scarp slope).

Figure 82: a) Walled channel of the Ravoire Torrent at Pontamafrey along the banks of the Arc River (centre of photo) (GSC 204166-D). b) View southward from above the apex of the Hermillon fan towards St. Jean-de-Maurienne; note flood retention basin in the background; the channel of the Hermillon Torrent is in the built-over area in the foreground (GSC 204166-E).

Between 20 and 22 December 1740, when the Glissement de Jarrier was reactivated, many other basins in the Maurienne also experienced large debris flows. These slope movements were caused by heavy downpours and snow melt accompanied by an unseasonally warm 'vent du midi'. In the basin of the Hermillon Torrent, east of the St. Jean-de-Maurienne, unstable flysch-evaporite terrain, that had shown ominous signs of instability in the decade before 1740, failed and released a debris flow which crashed through the village of Hermillon, demolishing 22 houses, and killing 7 people. On 25 July 1862, Hermillon again was struck by a debris flow, this time triggered by a local cloudburst; mills and bridges were demolished and one person was killed.

Unstable terrain also occurs in the catchment basin of the Torrent de St. Julien. There, on 18 July 1824, a cloudburst triggered major pulses of debris from pre-existing slide masses; a massive flow onto the cone erased 30 buildings and killed 2 people in the village of St. Julien-de-Maurienne. Other mid-summer cloudbursts brought destructive debris flows to St. Julien on 20 July 1871, and on 24 July 1872 (Mougin, 1914, p. 1076-1081).

Another torrent with a history of recurrent debris flows is the Torrent de la Ravoire. In spring 1965 a chronically unstable west-facing scarp slope composed of evaporites and carbonates slumped across the channel of this torrent (Glissement de Bon Attrait in Fig. 81); debris flows mobilized from the toe of this slide, demolished transportation routes in Pontamafrey (Goguel, 1968).

During the last hundred years reforestation and the construction of engineering works along the most dangerous torrents have somewhat alleviated the threat of debris flows to the communities in the valley. Some of the settlements are protected by dams above the cones; in others the channels of the torrents have been lined with high concrete revetments (Fig. 82a). In recent years, residential buildings crowd ever larger proportions of the cones (Fig. 82b). The high basin adjacent to the Glissement de Jarrier now also hosts a major ski development.

The notoriously unstable bedrock slopes in this region also pose formidable problems in the maintenance of the transportation corridor between France and Italy which passes through the Maurienne. A few kilometres east of St. Jean-de-Maurienne the route skirts the bouldery debris cone of the Torrent du Poucet whose intensely fractured upland source area has been found impossible to control by technical works. During heavy rains debris flows from the Poucet basin tend to impound the Arc River and push its bed southward against the transportation routes. One of the most dramatic blockages of the Arc occurred during the storm of 20-22 December 1740, when a lake impounded by debris from the Poucet basin drowned several buildings. Flooding of roadways and bridges also has occurred frequently since then. In 1955 the impounded waters of the Arc even entered the railroad tunnel. Other threats to the modern highway come from sagging slopes. Temporary road closures and rerouting of traffic have been accepted in the management of this very difficult environment (Jail, 1975; Azimi et al., 1980; Jail and Marnezy, 1980).

Zarera (A21)

Location: Val Laguné, Graubünden (Grisons), Switzerland (F2)
Date(s): 13 June 1486

The town of Zarera (1500 m) used to be located at the junction of Val Laguné and Val da Camp (Fig. 83), at the foot of the southwestern scarp face of Cima di Cardan (2793 m). Cima di Cardan is a bedrock ridge composed of northeast-dipping amphibolitic gneiss of the Austroalpine basement complex.

On 13 June 1486, a section of the Cima di Cardan ridge collapsed along a composite fracture surface. The disintegrating rock mass, with a volume of approximately 0.5 to $0.8 \times 10^6 m^3$, cascaded more than 700 m down the slope, spreading into a thin lobe 750 m long and 250 m wide. Zarera was obliterated and 300 people lost their lives (Heim, 1932, p. 123).

Today the busy Bernina Pass highway skirts the haunted surface, which is strewn with angular slabs of gneiss and is used mainly as alpine pasture.

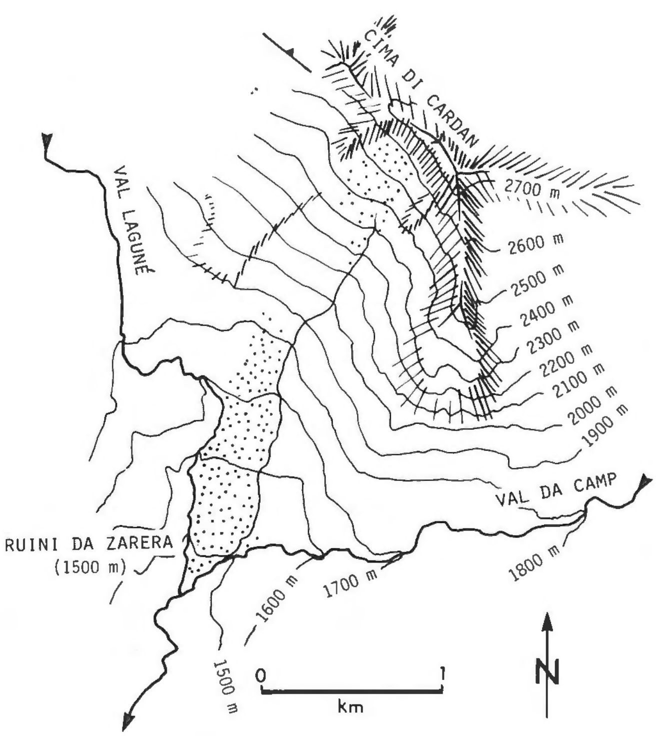


Figure 83: Sketch map of the fracture-controlled detachment zone and the block stream of Zarera (dots).

Brienz (A22)

Location: Brienzer See (Lake Brienz), Bern, Switzerland (D2)
Date(s): 1499 (also 1529, 1542, 1616, 1624, 1797, 2 November 1824, 1894, 26 May and 23 August 1896, 14 April 1901)

Lake Brienz (570 m) fills a depression along the west-trending Aare Valley (Fig 84). From the eastern end of the lake a bedrock slope rises northward to the mountain ridge of the Brienzer Rothorn (2350 m). This slope is underlain by north-verging folds of thin-bedded carbonate-shale units of the Helvetic cover complex and is dissected by several high-gradient torrents including the Trachtbach, Schwanderbach, Lammbach, and Eistlenbach torrents. Large aprons of debris have accumulated at the mouths of these torrents and grade into the fan delta of the Aare River along the eastern shore of Lake Brienz. During the Middle Ages large parts of the catchment basins of the torrents lost their forest cover, and retrogressive bedrock slumps and debris avalanches scarred the torrent embankments. To avoid the regular floods of the Aare River the settlements of Brienz, Kienholz, Schwanden, and Hofstetten had searched out the flanks of the debris cones, but were now menaced by increasingly serious debris flows.

In 1499 a violent rainstorm triggered a large slide (or several slides) in the upper reaches of the Lammbach Torrent. A mass of loose debris blocked the gorge, but eventually yielded to the pressure of impounded runoff; a flow of rock and mud then erupted onto the town of Kienholz, which, at the time of the disaster was an important regional administrative centre. Some 10 m of debris covered its buildings and about 400 people perished (Montandon, 1933, p. 295). The accumulated debris displaced the shoreline of Lake Brienz almost 100 m to the west (Schmidt, 1896, p. 60). Smaller debris flows inflicted damage in the area in 1529, 1542, and 1616. Then, in 1624, the community of Hofstetten was ravaged by a major debris flow from the Eistlenbach Torrent. In 1797 debris flows erupted simultaneously from the gorges of Lammbach, Schwanderbach, and Eistlenbach torrents, demolishing a total of 37 buildings.

On 2 November 1824, failure of a carbonate dip slope below the summit ridge of the Brienzer Rothorn generated a rapid boulder flow down the Trachtbach Torrent. Several buildings of Brienz disappeared underneath the blanket of carbonate rubble. This happened on a part of the cone where ancient (Roman ?) baths had probably suffered a similar fate (Montandon, 1933, p. 315). The Trachtbach Torrent again discharged a massive debris flow in 1894; buildings and roadworks in the upper parts of Brienz suffered heavy damage.

Towards the end of the 19th century a large bedrock slump developed along the eastern embankment of the upper Lammbach gorge. The detachment surface of this slump mass was a composite fracture-bedding plane dipping steeply towards the gorge. Incipient movement along the headscarp of the slump ('Rufisatz') had been noticed for more than 15

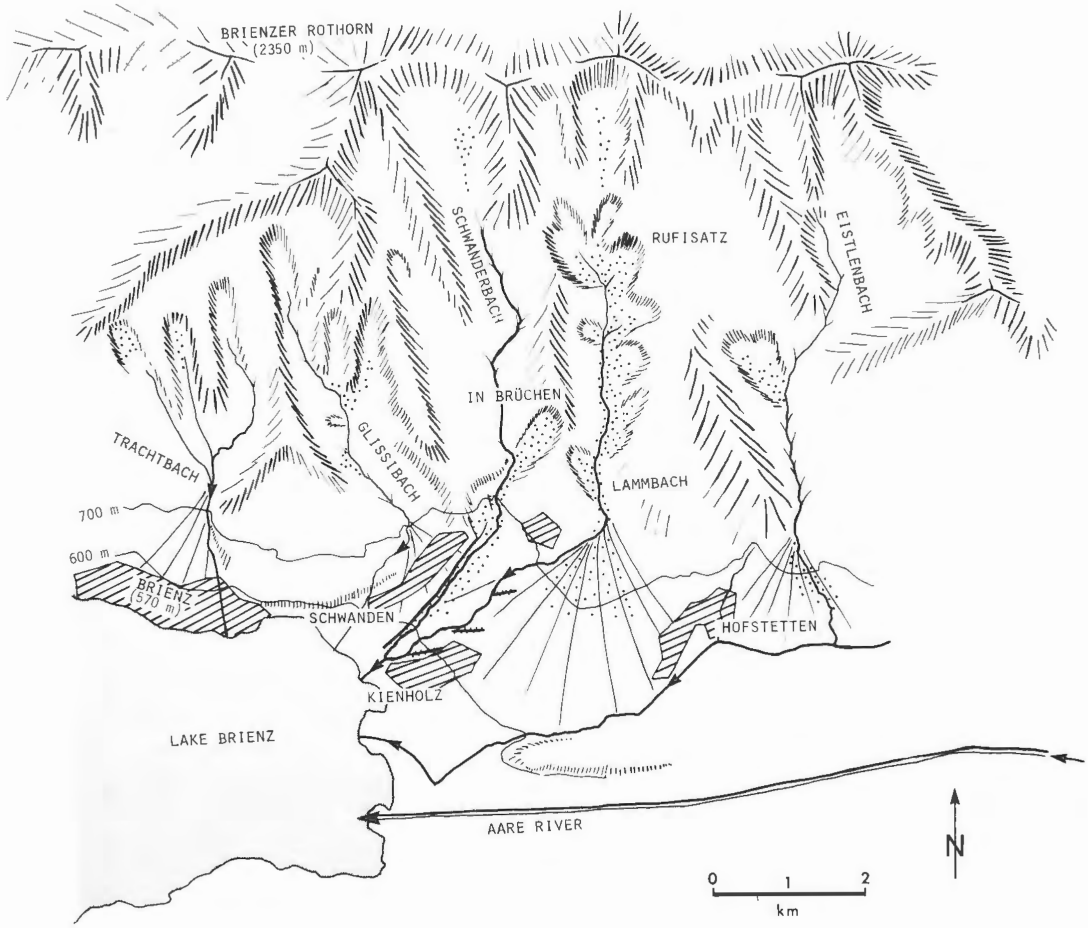


Figure 84: Sketch map of the unstable south slope of the Brienzer Rothorn; note the principal historical instabilities along the bedrock embankments of the torrents.

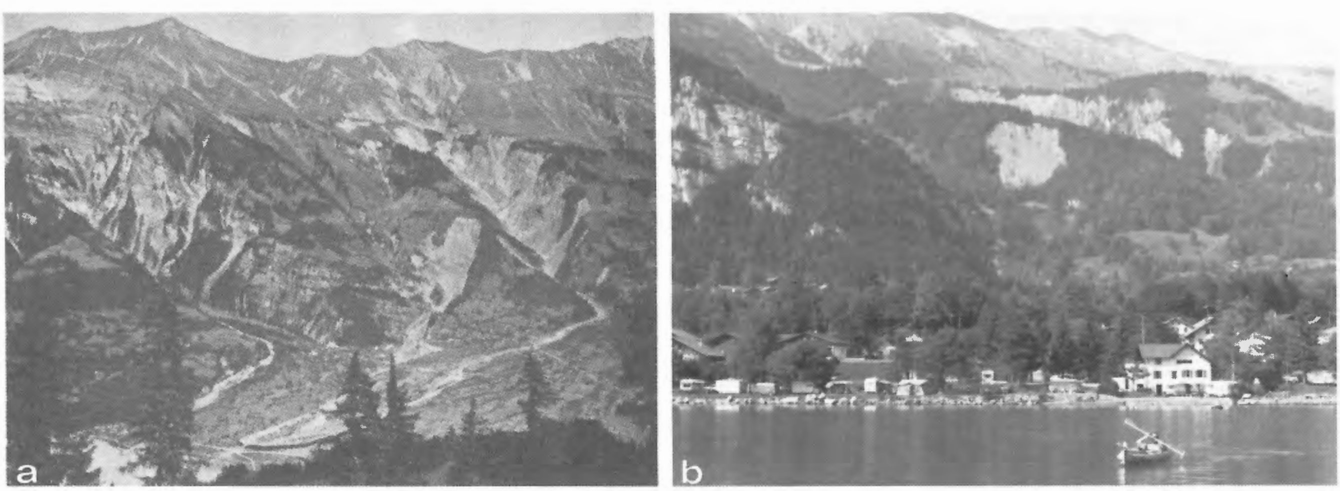


Figure 85: a) View of the Brienzer Rothorn south slope as it appeared about the beginning of the 20th century; note the deep Lammbach ravine on the right and the prominent slide scar 'In Brüchen' to the right of centre (from Salis, 1914). b) The Rothorn slope in 1981; note the difference in the amount of forest cover on the ridges flanking the former slide scars such as 'In Brüchen' in the centre of the photograph (GSC 204166-F).

years prior to 26 May 1896, when $0.3 \times 10^6 m^3$ of calcareous slate from the toe of the slide mass spread across the gorge and completely blocked the flow of the torrent. Soon this plug was saturated with water. On 31 May 1896, the frontal portion began to move down the Lammbach gorge at a rate of 2 m/s. Since the velocity of the flow was very low by the time it reached the cone, the farmers still had enough time to harvest their hay before

'....the metre-high wall of angular, cobblestone-sized limestone fragments advanced over the gentle slope, and by releasing water, came to a halt. The lobate margin of the flow rose over the grassland. The main mass of the flow subsided, so that the rims surrounded the blockstream as sharp-edged ridges, akin to moraines ...' (Schmidt, 1896, p. 63).

On 23 August 1896, a regional rainstorm mobilized the bulk of the Rufisatz in the uplands and debris engulfed a major part of Kienholz. These flows overloaded the fan delta of the Aare River, leading to a failure which launched a powerful turbidity flow towards the axis of Lake Brienz where a layer more than 1m thick was deposited (Sturm and Matter, 1978).

At the time of the Lammbach debris flows another slope failure, involving some $6 \times 10^6 m^3$ of carbonate and slate, developed along the eastern embankment of the Schwanderbach gorge ('In Brüchen'). On 14 April 1901, the toe of the slump developed into a flow, which came to a halt at a deflection wall that had been erected east of Schwanden in anticipation of the mass movement. Fortunately only a sixth of the bedrock slump was mobilized into the flow and artificial drainage of the slope subsequently slowed its movements (Heim, 1932, p. 158-162).

Beginning in the mid-19th century the first large debris deflection and retention dams were built along the torrents threatening the communities at the east end of Lake Brienz (Fig. 85). However, the width of the discharge sections along the channels soon were found to be insufficient to carry the accumulating debris through the towns. Following the disasters of 1894, 1896, and 1901 efforts were intensified to control the sources of the debris; stone-masonry check dams were built in the unstable upland ravines, aided subsequently by systematic reforestation (Salis, 1914, p. 24-42). This approach has proved successful and is being upgraded by modern concrete structures, gabion retaining-walls and dykes (Bauer, 1971).

Today the debris cones bear little direct evidence of the long history of destructive flows, and relatively calm creeks cross the communities in stone-concrete-lined channels.

Biasca (A23)

Location: Val Blenio, Ticino, Switzerland (E2)
Date(s): 30 September 1513 (also 20 May 1515, July (?) 1758, 27 September 1868, 4 October 1868)

The city of Biasca (300 m) overlooks the confluence of the Brenno River (Val Blenio) and the Ticino River (Valle Leventina). The mountains rise abruptly from the floodplain of the Brenno River to elevations between 2000 and 2800 m (Fig. 86), and are composed of Pennine core gneisses. Foliation in the high grade metamorphics dips between 20 and 25° northeast; steep fracture zones across the trend of the foliation control the location of sagging slopes and rockfall zones (Zeller, 1964). Traditional communities in Val Blenio and Valle Leventina are perched on steep debris cones or narrow ledges of bedrock, thus avoiding the flood-prone valley bottoms. One kilometre north of Biasca a conspicuously barren cone, the Buzza di Biasca, spreads across Val Blenio from the mouth of the bedrock gorge of Crenone; above this defile soars an imposing rock wall, the summit face of Pizzo Magno (2329 m).

On 30 September 1513, during intense autumn rains, an unstable rock wedge, 10 to $20 \times 10^6 m^3$ in volume, collapsed along a north-trending, near-vertical fracture zone below Pizzo Magno. The disintegrating slab fell westward into the Crenone gorge, then fanned across Val Blenio, pushing a frontal wave of debris 100 m up the opposite valley wall. The main rock avalanche was preceded by rockfalls that destroyed a number of buildings; the avalanche itself probably devastated parts of the northern outskirts of Biasca. However, the main catastrophe was still in the making. The conical debris mass dammed the Brenno River, forming a shallow lake which eventually extended 5 km upstream and attained a volume of more than $100 \times 10^6 m^3$. The rising waters drowned the hamlets of Malvaglia and Semione. For more than a year it looked as if the new lake was to become a permanent feature of the valley. However, on 20 May 1515, springtime overflow at the lower outlet of the lake eroded deeply into the debris plug, causing it to collapse. An explosive surge of debris and water engulfed Biasca, swept down the valley of the Ticino to Bellinzona, and finally set off a huge wave in Lago Maggiore (Heim, 1932, p. 172-173; Montandon, 1933, p. 295-296). About 600 people lost their lives. Long after the catastrophe the survivors in the valley below the Buzza di Biasca felt that their upstream neighbours had been responsible for the disaster. After all, the flood had restored the previously inundated villages and adjacent agricultural land. Deep feelings of distrust persisted for a long time (End, 1922, p. 121-128).

The rainstorm possibly triggered the Pizzo Magno rock avalanche on 30 September 1513, also released a debris or rock avalanche which destroyed a village named Campo Bargigno, near Cauco in Val Calanca, approximately 10 km to the east of Biasca (Montandon, 1933, p. 296).

Other destructive boulder flows down steep bedrock ravines of the region have occurred more recently. In July (?) 1758, a mass of rock broke away from the gneissic scarp face

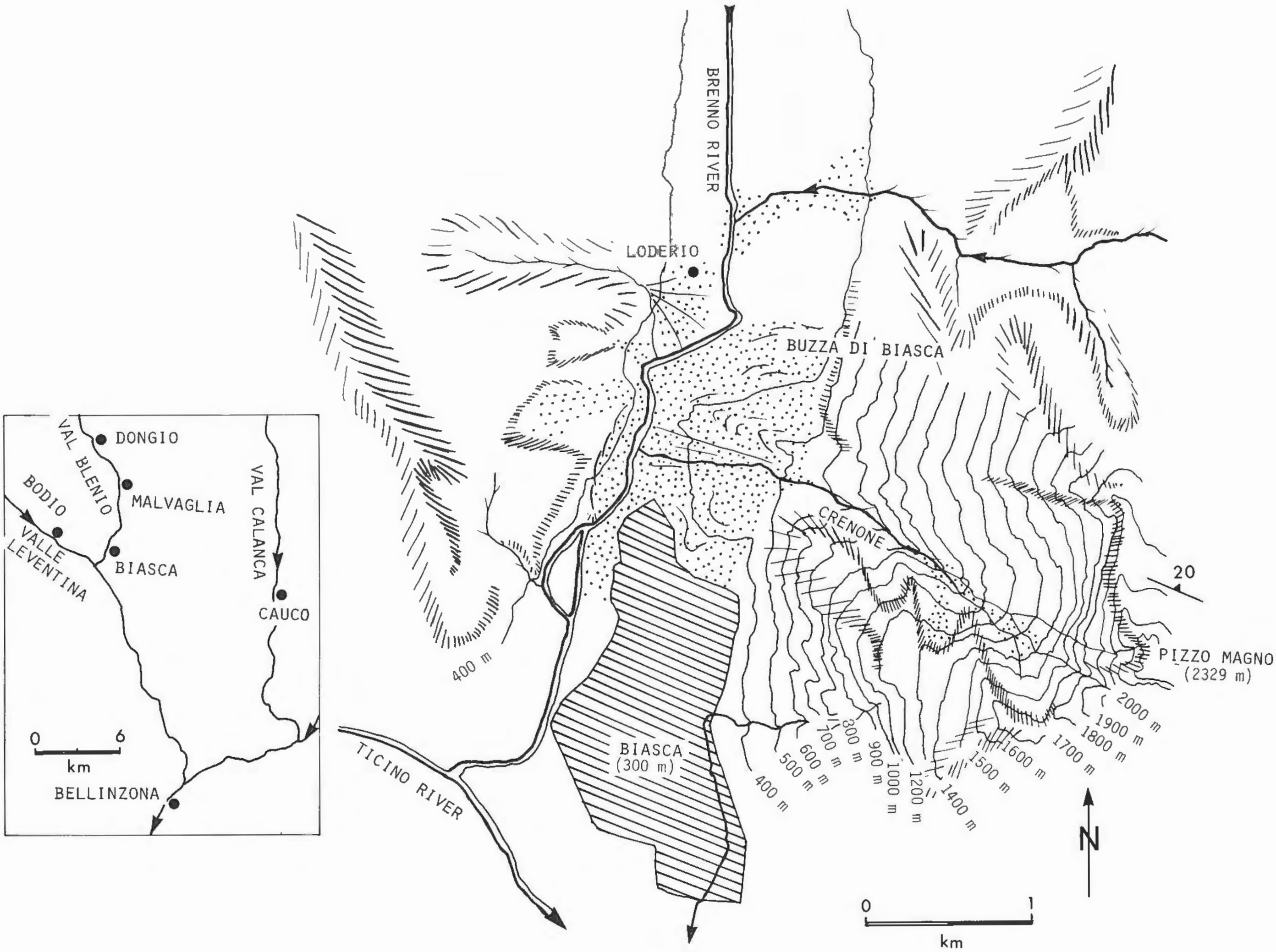


Figure 86: Index and sketch map showing the location of Buzza di Biasca and the detachment zone of the 1513 rock avalanche along the west flank of Pizzo Magno.

Figure 87: a) View of a recently developed northern suburb of Biasca on the lower segment of the Buzza cone (GSC 204166-G). b) Blocky surface of the upper part of the Buzza di Biasca (GSC 204166-H).

above Dongio, 10 km north of Biasca, surged over the settlement and claimed the lives of 50 people (Montandon, 1933, p. 308).

On 27 September 1868, following a rainstorm that deluged the south-central Alps, an avalanche of gneissic blocks cascaded down the southwest-facing ravine above Bodio (320 m) in the Valle Leventina. The stream of gneissic blocks flattened 70 buildings and killed 22 people. According to oral traditions this had not been the first time that Bodio was struck by this kind of disaster. On 4 October 1868, a massive debris flow devastated the hamlet of Loderio, north of Biasca, and closed the narrow passage between the cone of Loderio and the Buzza di Biasca. The flow of the Brenno River was stemmed once again and the floodplain north of the Buzza became a huge lake. Understandably, this created a panic in Biasca, where the church bell was run with such vehemence that it cracked! However, the Brenno River overflowed without causing significant damage along the toe of the Buzza di Biasca. The losses from debris flows and slope failures in this region in 1868 were heavy, including the destruction of Cumiasca, a hamlet west of Dongio, where more than 20 persons were killed. Many people emigrated from the district after the catastrophe (Montandon, 1933, p. 311).

In November 1951, when the southern Alps suffered one of the most serious periods of flooding in recorded history, many debris flows occurred in the Biasca area. At this stage, however, a regional program of dyking and torrent control had been in effect for some time and the control works lessened the impact of mass movements.

In recent years residential construction activity in the vicinity of Biasca has been intense. Training of the Brenno and Ticino rivers has opened new land for development; building activity has also invaded cones with known debris flow history. Suburbs of Biasca are now advancing towards the Buzza (Fig. 87).

Ardenno (A24)

Location: Valtelline, Lombardia, Italy (E2)
Dates(s): 9 May 1538

Ardenno (260 m) is one of several agricultural villages and small towns that nestle near the apices of debris fans and on narrow terraces above the floodplain of the Adda River (Fig. 88). The flanks of the U-shaped Adda Valley (Valtelline) are composed of steeply dipping gneisses and schists of the Pennine and Austroalpine basement complexes. Short, high-gradient torrents draining into the valley commonly flow across sagging bedrock slopes or terraces composed of late Pleistocene ice margin deposits.

On 9 May 1538, a massive slope failure in the steep uplands above Ardenno changed into a rapidly moving debris flow which demolished a number of buildings in the village killing seven people. A large number of inhabitants were spared a similar fate because they were busy in the fields at the time of the disaster (Montandon, 1933, p. 296).

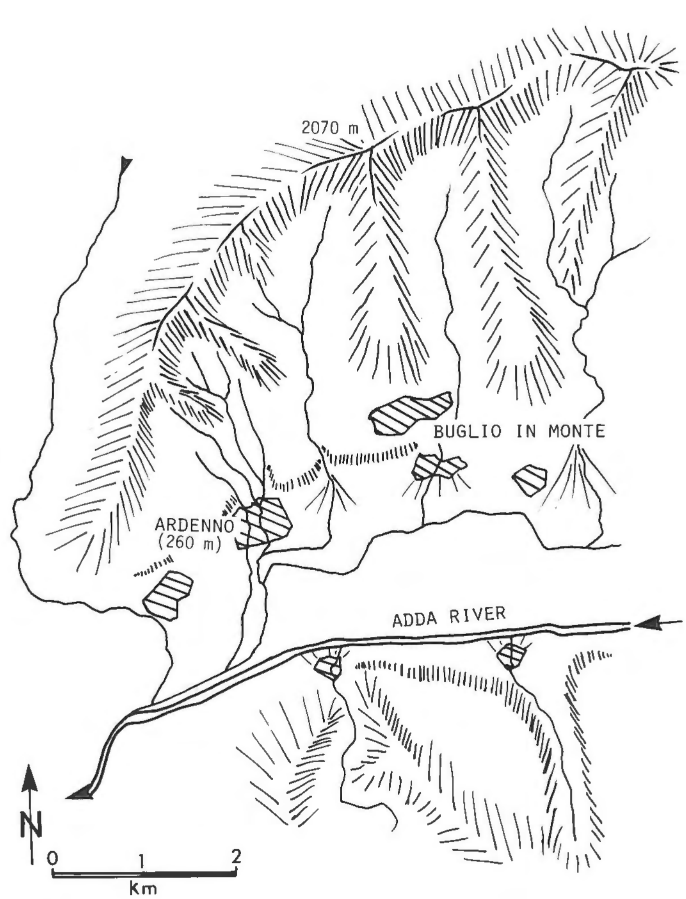


Figure 88: Sketch map of settlements above the floodplain of the Adda River (Valtelline).

Figure 89: View of the built-over cone of Ardenno as seen from the Adda floodplain; note vegetated ravines above the town. (GSC 204166-I)

Today much of the floodplain of the Adda has been brought under cultivation. Settlements continue to cluster near fans and cones along the sides of the valley. The small fan of Ardenno is now completely built over (Fig. 89).

Radmer an der Hasel (A25)

Location: Enns River, Steiermark (Styria), Austria (J1)
Date(s): 1540 (?)

The Radmer Valley is entrenched into rugged carbonate ranges of the Austroalpine cover complex on the south side of the Enns River. The upper Radmer basin is underlain partly by recessive shale-evaporite units and partly by carbonate cliffs attaining elevations of more than 2000 m. Copper and iron deposits found in the recessive strata along the base of the cliffs have attracted miners to this remote region for more than 3000 years.

During the Middle Ages the town of Radmer an der Hasel (900 m) developed into a thriving centre of copper mining, at times employing as many as 1000 people in the extraction and processing of ore. The mine workings extended far into the base of the steep scarp face which rises on the northwest side of the town to the summit ridge of the Lugauer Massif (2206 m). Partly due to subsidence of recessive strata below the carbonate rocks dipping approximately 40° to the northwest, the scarp face is laced with vertical extension fractures parallel to the valley (Fig. 90).

According to Stini (1938. p. 14), a rock avalanche buried most of Radmer in 1540 (?). A carbonate slab, 2 to 4 × $10^6 m^3$ in volume, failed along a steep composite fracture zone along a bedrock ledge known as Gspitzter Stein (1550 m). The debris lobe crossed the floor of the valley and, judging from several low terraces on its upstream side, probably blocked the flow of the Haselbach Torrent. Mining in the shales below the cliff may have precipitated the catastrophe.

Figure 91: View of the western half of the built-over rock avalanche lobe at Radmer an der Hasel; note the thoroughly shattered bedrock spur overlapped by fresh rockfall debris and the flanks of the historical detachment zone on the west (left background) (GSC 204166-J).

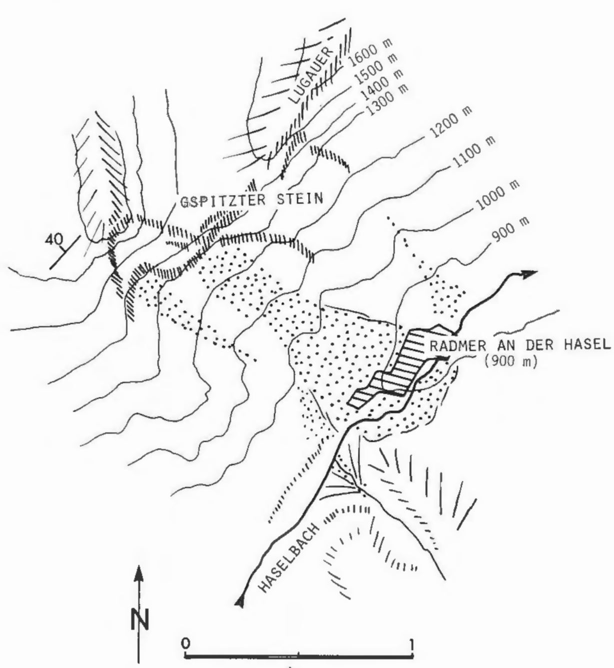


Figure 90: Sketch map of the detachment zone and rock avalanche deposits of Radmer an der Hasel.

Today the houses of the village and a castle rest on the rubbly surface of the slide mass (Fig. 91); above the settlement two fresh rockfall cones extend far into steep ravines cut into a shattered bedrock spur that borders the backwall of the 1540 slide.

Matrei (A26)

Location: Isel Valley, Osttirol, Austria (H1)
Date(s): 1550 (also 1347, 1702, 30 July 1723, 1752, spring 1827, 28 June and 22 July 1841, 18 August 1849, autumn 1882, 19 July 1933)

Matrei (974 m) is an old community situated on the debris cone of the Bretterwandbach Torrent, an eastern tributary of the Isel River (Fig. 92). The upland basin of this torrent is underlain by low grade metamorphic rocks of the Pennine Tauern Window. The foliation in phyllite, greenstone and lenticular carbonate bodies dips to the south. The terrain below the eastern crest of the basin (Kalsergrat) is characterized by numerous slumps and rockfall cones; the broad ridge to the west (Glunzerberg) is a composite sagging slope of at least 500 × $10^6 m^3$ (Zischinsky, 1969).

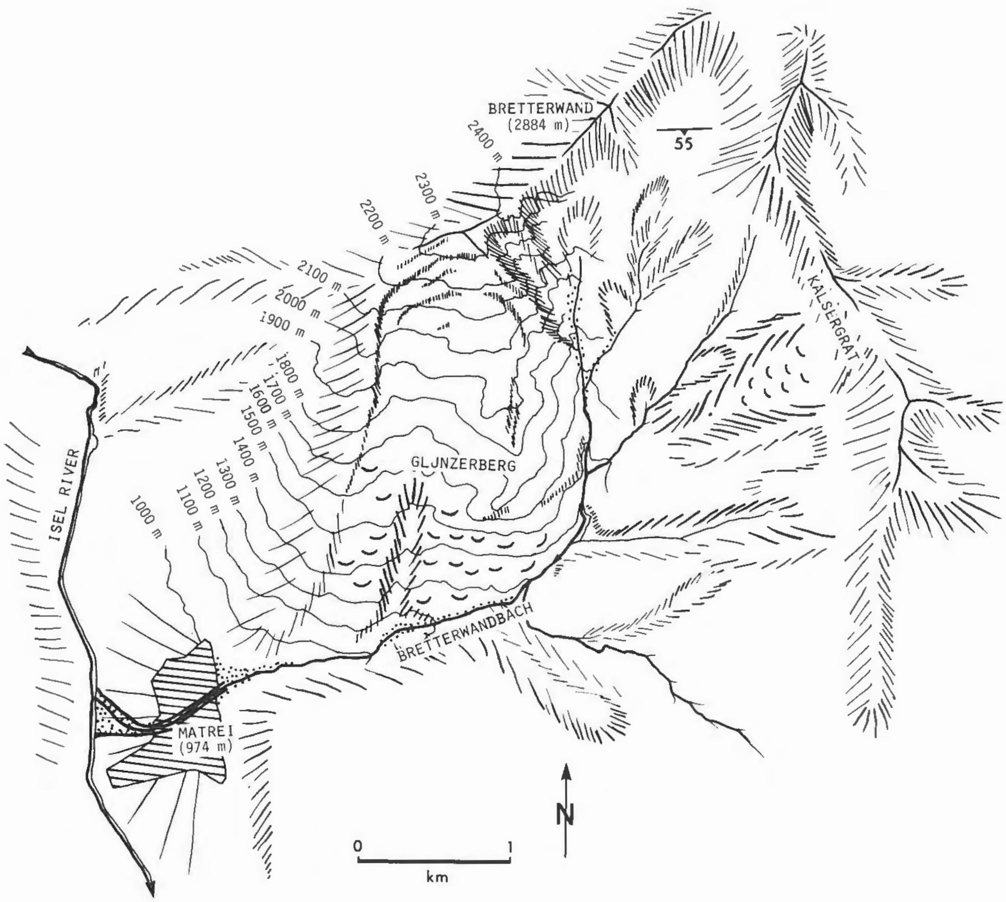


Figure 92: Sketch map of the Bretterwandbach basin near Matrei and the debris sources along the unstable bedrock slopes flanking the torrent; note the sagging Glunzerberg slope which involves low grade metamorphic rocks whose foliation dips about 55° to the south.

Figure 93: a) Concrete-masonry check dams spanning the lower part of the Bretterwandbach gorge above Matrei (person on the lower left for scale). Dense alder growth attests to present channel stability. In the background is a leaning limestone cliff forming part of the sagging Glunzerberg (GSC 204166-K). b) Protective masonry-concrete dykes flanking the Bretterwandbach channel in Matrei. The bridge can be raised mechanically during times of heavy runoff or threatening debris flows (GSC 204166-L).

Sustained deep-seated mass movements throughout the basin have been responsible for repeated destructive debris flows at Matrei, and more than once in the past the idea of relocating the entire town has been entertained. However, the tenacity of its people assured survival and growth.

A massive flow, probably in 1347, buried the locality completely (Hanausek, 1975, p. 108). In 1550, and during the decade that followed, debris flows destroyed most buildings and the church. Severe damage was inflicted in 1702, then again after a hailstorm-cloudburst on 30 July 1723, and once more in 1752. Several debris flows in spring of 1827 laid waste to half the town (Sonklar, 1883); between 28 June and 22 July 1841, the primitive dykes across the town were breached by debris which invaded residential buildings. Similar damage was registered on 18 August 1849, in autumn 1882, and on 19 July 1933 (Forcher, 1980, p. 109-112).

Debris flows from the Bretterwandbach gorge also have shifted the channel of the Isel River; thus programs to stabilize the channel of the torrent have paralleled efforts to train the receiving river. In the upper parts of the torrent a spectacular array of check dams has now been maintained for more than 80 years to prevent excessive erosion along the unstable bedrock embankments (Fig. 93a). A paved channel, flanked by strong stone-masonry walls 4 to 5 m high, escorts the torrent through Matrei to the dyked riverbank (Fig 93b). The two bridges crossing the torrent in Matrei can be raised mechanically to prevent blockage of debris and breaching of the walls during floods and debris flows. Matrei is a growing tourist centre at the entrance of a proposed national park. Residential buildings now cover almost the entire surface of the debris cone.

Aosta (A27)

Location: Valle d'Aosta, Italy (C3)
Date(s): 6 July 1564 (also 11th century, 1449)

The modern city of Aosta (580 m) developed from a fortress which guarded the route from the Dora Baltea River (Valle d'Aosta) to the northwestern provinces of the Roman Empire. The valley has always been densely populated, with most of the settlements accommodated on debris fans and narrow terraces above the floodplain of the valley (Fig. 94). The dip slopes on the north side of the valley are composed of low grade metamorphic rocks of the Pennine cover complex. During the relatively frequent regional rainstorms, slide masses in upland basins tend to develop directly into massive debris flows of devastating impact.

In the 11th century the town of Chambave (486 m), then situated to the west of the present community, was left in ruins after a mass of debris descended from the slopes below Becca d'Aver (2469 m). In 1449 another large debris flow from the slopes of Mont Mary (2815) flattened the village of Quart; the debris cone below the castle of Quart has since been only sparingly occupied and its lower segments are still shielded by a thick protective forest (Fig. 95). On 6 July 1564, a slide mass from the southeast slope of Becca France (2312 m) completely demolished the village of Thora, killing a large number of people. The small debris cone at the bottom of the mountain is now being gradually occupied by houses of the community of Sarre (Montandon, 1933, p. 283, 292 and 297).

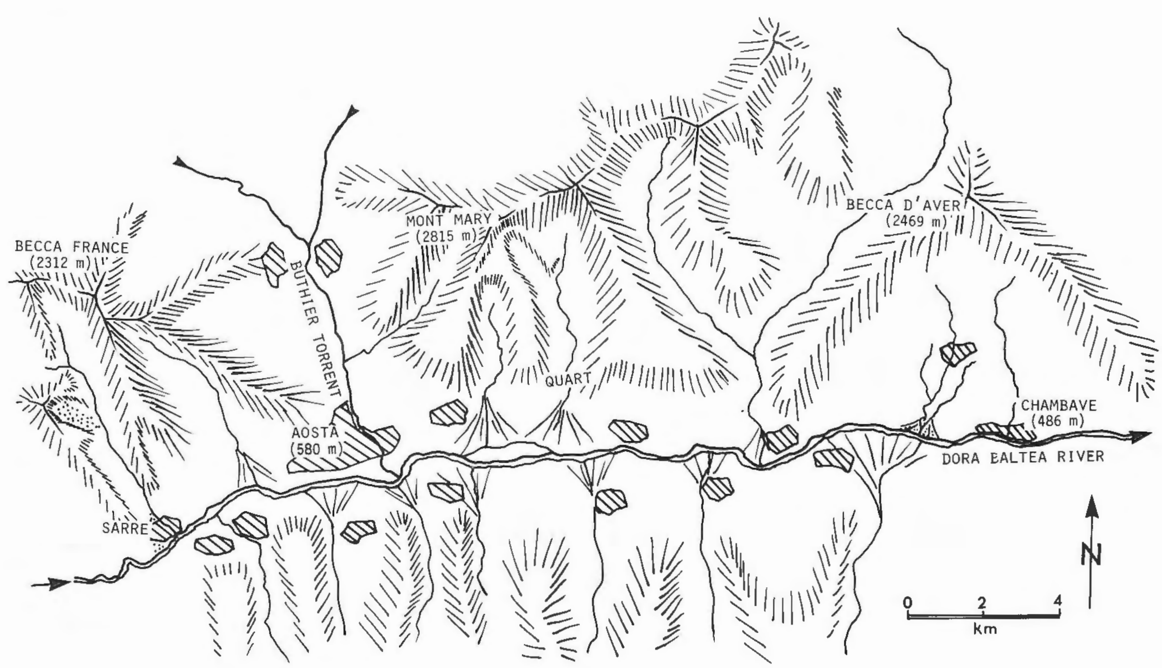


Figure 94: Index map of the densely populated Aosta region; note location of villages (diagonal ruling) astride debris cones of high-gradient torrents and above the floodplain of the Dora Báltea River.

Figure 95: View of the cone and catchment area above Quart east of Aosta; a protective forest covers most of the axial sector of the cone and sparse vegetation has reconquered the denuded uplands. Suburban residential development has begun to spring up on the flanks of the cone. (GSC 204166-M)

In spite of great development pressure in this region the axial sectors of the most notorious debris cones have been left undeveloped and several host protective forests. Modern residential construction skirts the fringes of bouldery debris flow deposits of the past. In recent years, extensive reforestation work has been carried out in some denuded upland basins.

Schlierengrat (A28)

Location: Sarnen Valley, Unterwalden, Switzerland (D2)
Date(s): 1565 (?) (also 1629, spring 1910, 5 August 1931)

The Schlierengrat (1700 m) is a southwest-trending bedrock ridge in the front ranges of the Swiss Alps (Fig. 96). Its southeast-facing dip slope above the Sarnen Valley is composed of argillaceous-sandy Schlierenflysch Formation of the Helvetic cover complex. Gently undulating bedrock surfaces and benches of surficial deposits flank the deeply incised torrent channel. Several retrogressive bedrock slides extend to the crest of the basins. Two major torrent systems drain the Schlierengrat dip slope: the Giswiler Laui and the Schlieren torrents; they have built substantial cones into the Sarnen Valley.

In the late 16th century (about 1565?) a large slide mass blocked the Grosse Schliere Torrent above its gorge to form a new lake. This lake persisted for almost a year before the debris dam failed and a massive flow of mud and rock swept over the cone and engulfed the community of Schoried. In 1629 a bedrock slump in the upper basin of the Giswiler Laui Torrent set up debris flows which destroyed parts of the village of Giswil (Montandon, 1933, p. 297 and 301).

In this early period destructive mass movements were caused in part by extensive removal of upland forests, overgrazing, and log haulage in ravines. Resource use has improved substantially during the modern era, but the unstable flysch terrain is still susceptible to erosion of torrent embankments and deep-seated creep during periods of excessive infiltration of water. For instance, in 1880 crown cracks and depressions indicated that a good part of the slope above the community of Sörenberg was in motion. During the exceptionally wet spring of 1910 some $1.5 \times 10^6 m^3$ of shale and sandstone began to move along a composite south-facing bedding-fracture surface. A frontal bulge of sandstone slabs impounded the torrent in the valley temporarily. After a harmless overflow the remaining slump mass stabilized (Heim, 1921, p. 436-437). The stop-and-go behaviour of many slopes in sandy argillaceous flysch ('flysch' = flowing terrain) depends greatly on the temporal pattern of precipitation and local runoff. Countermeasures, designed to neutralize the effect of large scale creep and small scale embankment failures in the Schlieren area include terracing and reforestation of major debris sources, emplacement of check dams and retaining walls along reaches with marked erosion, draining of swampy depressions, diversion of tributary torrents from unstable gullies into more stable channels, excavation of debris retention areas on the cone, and dyking of channels (Eicher, 1977). Nevertheless, sporadic setbacks, such as a massive debris flow from the Grosse Schliere Torrent on 5 August 1931, still have to be accepted during extreme rainstorms.

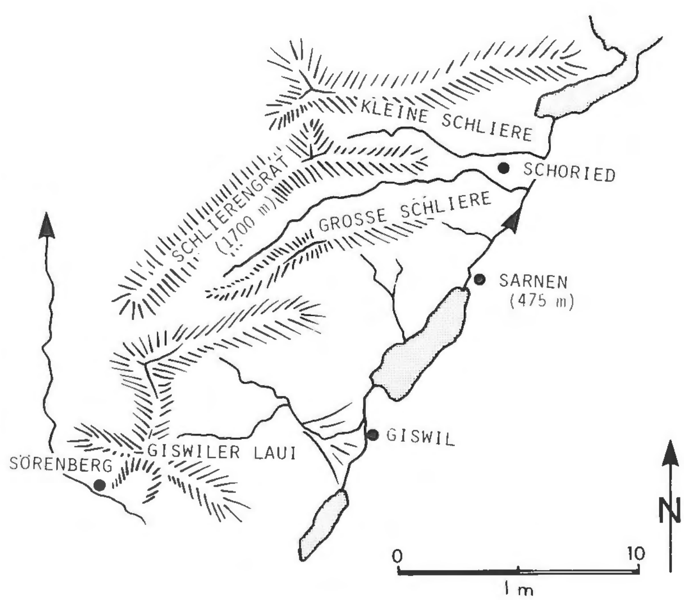


Figure 96: Index map of the Schlierengrat; most of the southeast-facing dip slopes are underlain by slide-prone sandy argillaceous flysch.

Gastein Valley (A29)

Location: Salzburg, Austria (I1)
Date(s): 14 June 1569 (also 1493, 1567, 29 June 1618, 10 October 1789, 21 August 1966)

The Gastein Valley, south of the Salzach River, is carved into low grade metamorphic cover and gneissic basement of the Pennine Tauern Window (Fig. 97). Foliation in phyllites, greenstones, and calcareous schists dips to the north.

Gold veins in the basement gneisses first attracted miners to the Hohe Tauern Range (2500 – 3000 m) some 3000 years ago. Early mining communities suffered repeatedly from floods, debris flows, landslides, snow avalanches, and even glaciers that advanced over mine adits. However, one massive slope failure actually created a lasting economic base for the community of Badgastein. Sometime in the Middle Ages thick relict colluvium and weathered bedrock slid away from the west side of the Graukogel Massif and bared gneissic bedrock, thus exposing several hotsprings which henceforth attracted health seekers from far and wide. This slide mass was reactivated during sporadic rainstorms, such as in 1493 and on 10 October 1789, but at Badgastein damage to buildings and baths was always repaired promptly.

Figure 98: View of Hofgastein from above the gorge of the Rastötzenbach. Most of the buildings near the apex of the fan are of relatively recent date. (GSC 204166-N)

Figure 97: Sketch map of the Gastein Valley showing the location of the unstable Graukogel debris slide east of Badgastein and the completely built-over fan of the Rastötzenbach Torrent at Hofgastein.

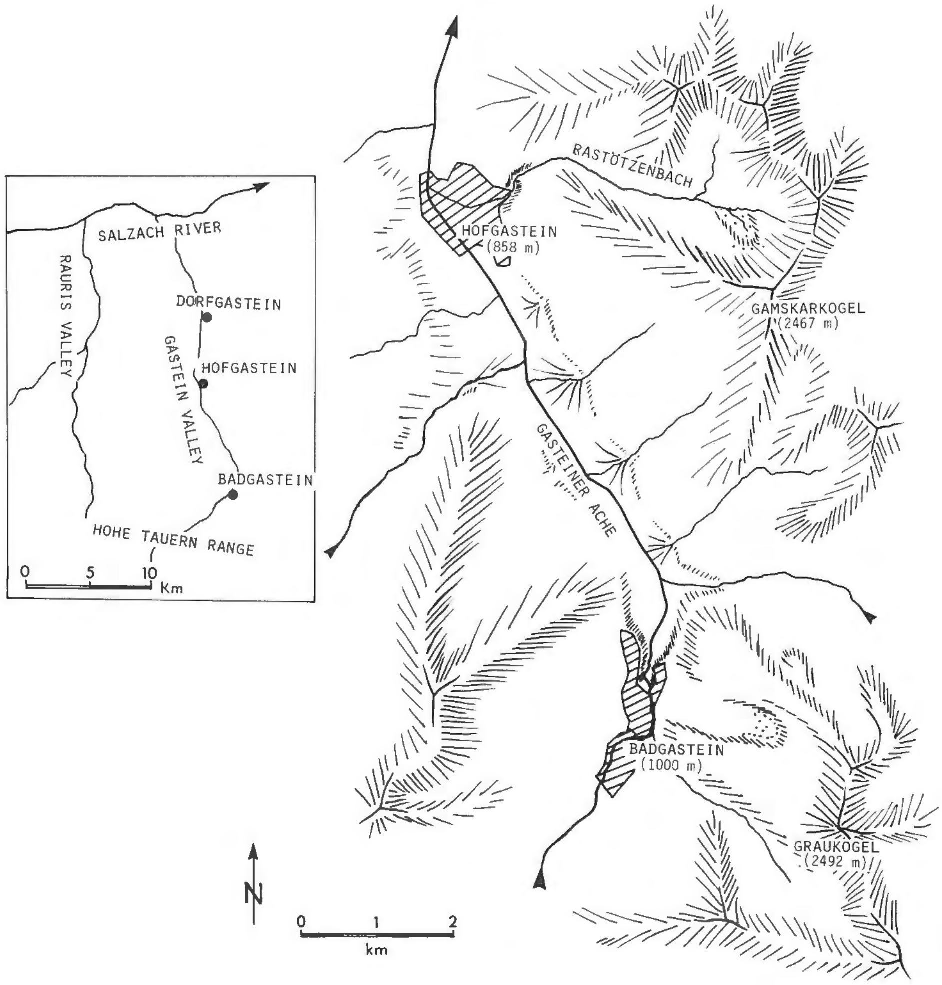


Figure 99: Typical slump in low grade metamorphic rocks of the Gastein region; note recently constructed tourist accommodations near the toe of the slide (GSC 204166-O).

By the 16th century many of the upland basins of the Hohe Tauern had lost their forest cover; slides on bare slopes blocked the channels of torrents and soon settlements in the main valleys began to suffer from debris flows and floods. In 1567, during a thunderstorm, a log-debris jam in the Rauris Valley failed explosively and a wave of mud, rocks, and trees devastated a mining town, killing approximately 100 people.

The largest of the debris flow disasters befell Hofgastein (858 m). This town is located on the debris fan of the Rastötzenbach Torrent, which drains a steep upland bowl underlain by unstable calcareous schists and Pleistocene colluvium. On 14 June 1569, a local cloudburst mobilized much debris in ravines of the upper Rastötzenbach basin. Trees and boulders soon blocked the narrow bedrock gorge above the town and then swept onto the cone, demolishing 52 buildings and claiming the lives of 147 people (Strele, 1936, p. 126). A similar rainstorm, followed by debris flows inflicted serious damage to the town on 29 June 1618.

The high population density of the Gastein Valley led to early countermeasures such as the establishment of protective forests and the construction of primitive masonry dykes. Nevertheless, slides involving metamorphic bedrock and colluvium continue to threaten some of the communities. For example, on 21 August 1966, after five days of rain totalling 370 mm, the Graukogel slide once again endangered the Badgastein hot springs, and several debris cones in the valley were ravaged by debris floods (Lauscher, 1973).

The tourist boom of the last few decades has spurred the construction of protective dams above several notorious cones, check dams in upland source areas, and training walls along the torrential Gasteiner Ache. Some debris cones (e.g. Hofgastein) are now almost completely built over (Fig. 98); unstable slopes along the valley flanks are locally close to new residential developments (Fig. 99).

Leytron (A30)

Location: Rhone Valley, Valais, Switzerland (C2)
Date(s): 1570 (also January 1906)

The town of Leytron (500 m) stretches along the foot of the stepped cliff of L'Ardève Mountain (1501 m), a ridge which soars above two bulging debris cones north of the Rhone River floodplain (Fig. 100). L'Ardève Mountain and adjacent unstable slopes are underlain by folded calcareous slate of the Helvetic cover complex. Bedding and cleavage dip approximately 35° to the south-southeast. The two lobate debris cones, now intensely cultivated, are the result of extensive prehistorical and historical mass movements. Narrow platforms at the base of the L'Ardève Mountain suggest that from time to time this bedrock spur has been mined along precariously placed adits!

In 1570 a massive failure along the rock face wrecked a good part of Leytron (Montandon, 1933, p. 297). The modern town has expanded onto tracts of vineyards underlain by slide rubble.

The two debris lobes on either side of the L'Ardève attest to more gradual retrogression of the dip slope of the L'Ardève spur. In January 1906 a slump involving 5 to 6 × $10^6 m^3$ slaty bedrock and surficial deposits above Chamoson generated several debris flows. Rapidly executed drainage works on the unstable terrain greatly retarded the movement of the flows and averted the threat to the village (Heim, 1921, p. 461).

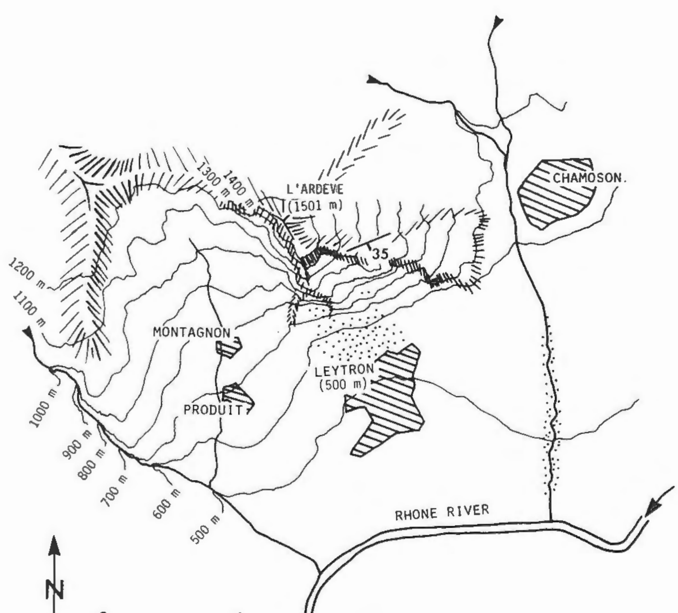


Figure 100: Sketch map of the region surrounding Leytron. This town is situated at the foot of a bedrock dip slope composed of calcareous slates and is flanked by large prehistorical debris cones.

Neukirchen (A31)

Location: Salzach Valley, Salzburg, Austria (H1)
Date(s): July (?) 1572 (also 1826)

The upper Salzach Valley is a densely populated east-trending glacial valley segmented by torrential tributaries of the Salzach River. The town of Neukirchen (850 m) extends onto the eastern flank of one of the most impressive of these debris cones at the mouth of the Dürnbach Torrent (Fig. 101). The catchment basin of the torrent rises to bare upland ridges more than 2000 m in elevation and is underlain by slide-prone south-dipping phyllite and mica schist of the Austroalpine basement complex. Sagging bedrock slopes constrict the torrent, giving rise to sporadic outbursts of debris on the fan below. A local legend recounts that the ancient village of Mitterdorf, straddling the axis of the Dürnbach cone, disappeared under masses of debris; the community of Neukirchen was later established east of the mired settlement. However, in July (?) 1572 Neukirchen itself was buried by slide masses – debris flows. Great devastation on the cone was also recorded in 1826 (Stini, 1938, p. 15 and 23).

Later attempts were made to control the upland debris sources by arrays of check dams along the torrent, but the sagging terrain posed exceptional problems. Eventually, check dams with flexible gabion cores were installed to absorb the lateral pressure from the embankments (Kronfellner-Kraus, 1974, p. 339-340). A protective forest and a set of dykes also separate the thriving town of Neukirchen from the lower channel of the Dürnbach Torrent.

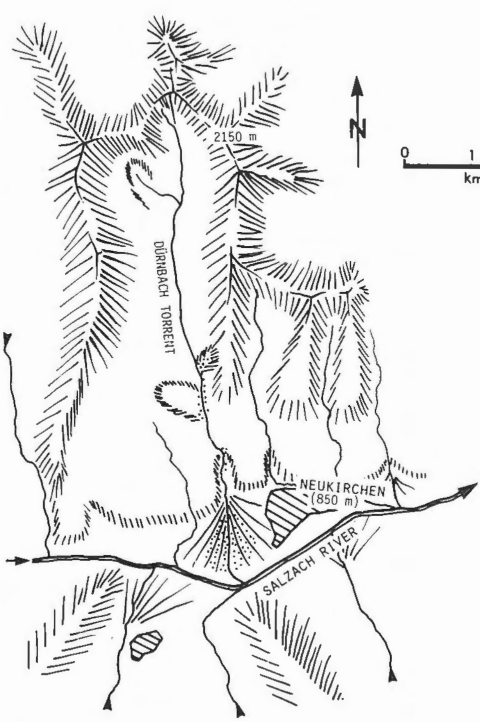


Figure 101: Sketch map of Neukirchen and the basin of the Dürnbach Torrent.

Corbeyrier — Yvorne (A32)

Location: Tour d'Ai, Vaud, Switzerland (C2)
Date(s): 4 March 1584

Tour d'Ai (2331 m) east of Lake Geneva is the highest point of a northeast-trending anticlinal ridge composed mainly of limestone of the intensely deformed Pennine klippen belt of the Swiss Préalpes (Fig. 102). The south-facing scarp slope of Tour d'Ai exposes northward-dipping recessive argillaceous limestones grading upwards into thick-bedded limestone which forms the castellate summit ridge. Below the scarp face a bench of Pleistocene surficial deposits forms the bowl-shaped terrace of Luan (1200 m) from which the Torrent d'Yvorne flows to the Rhone Valley (400 m).

The communities of Yvorne (445 m) and Corbeyrier (920 m) were well established agricultural villages when, on 1 March 1584, a severe earthquake jolted the area. A carbonate ledge on the south-facing scarp slope of the western summit ridge of Tour d'Ai collapsed and blocky debris ran out over the terrace of Luan. From 2 to 3 March (?) it rained and snowed almost without interruption. The Luan terrace, now overloaded by the rock avalanche lobe, soaked up large amounts of water from the rain and meting snow. On 4 March the weather improved and people again worked the fields below Luan. However, deep fissures opened at the terrace rim and slabs of surficial deposits began to skid and slump downhill. Near Corbeyrier cracks in the soil ejected mud and the earth trembled. Noisy toppling of trees should have suggested to the inhabitants of the village that much of the terrace was now in motion. However, most people remained in the fields. On the afternoon of 4 March incipient creep and ground subsidence changed into a well-defined stream 600 m

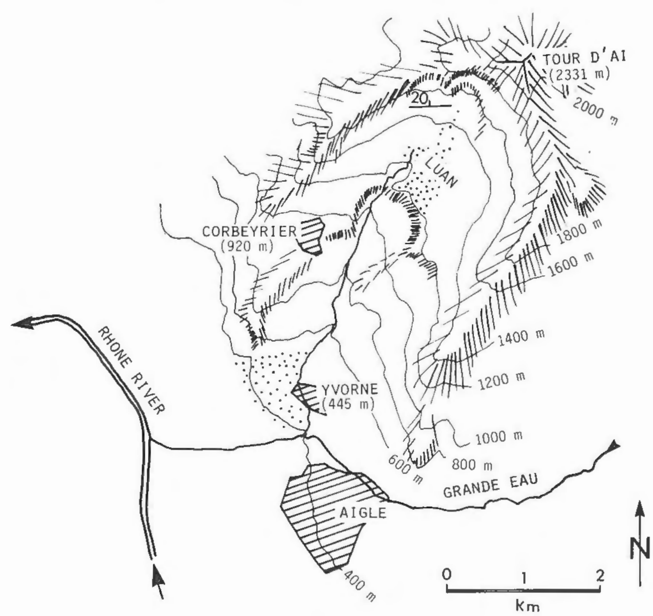


Figure 102: Sketch map of the south-facing scarp slope of Tour d'Ai; note the broadly concave detachment zone of the debris slide of 1584 below Luan.

Figure 103: Southward view from the top of the Tour d'Ai scarp face onto the forest-covered terrace of Luan; Rhone Valley in the hazy background. (GSC 204166-P)

wide, involving Pleistocene surficial deposits and superincumbent rock avalanche material. Along its western flank the slide mass tore away the main section of Corbeyrier. As it accelerated along its downward tapering track, the debris stream attained a volume of $10 \times 10^6 m^3$. Its velocity was so great that the front locally became airborne as it cleared slight elevations in the track. As the chronicler describes it, the debris. . . 'jumped over several plots and grape stands without doing any damage.' (Heim, 1932, p. 155). The slide mass overwhelmed the entire village of Yvorne and claimed the lives of 328 people (Heim, 1932, p. 155 and 186; Jeannet, 1918, p. 690-694).

The breakaway scar of the rock avalanche is still recognizable below the crags of Tour d'Ai. The steep grassy slopes below the cliffs are the starting zone of snow avalanches whose runout zones extend onto the terrace of Luan. The terrace itself is mantled by a thick protective forest, broken by a few clearings for summer chalets (Fig. 103). The conical debris deposits of 1584 on the north side of the Rhone Valley host the enchanting village of Yvorne and its vineyards.

Vorder Glärnisch (A33)

Location: Glarus, Switzerland (E1)
Date(s): 3 July 1594 (also 1701 to 1706)

The Glärnisch Massif (2900 m) southwest of Glarus (480 m) is part of the Helvetic cover complex (Fig. 104). Its rugged summit ridge is composed of massive carbonate ledges and recessive intervals of calcareous slate. The gently south-dipping panels of carbonate are cut by steep fracture zones parallel and perpendicular to the northeasterly structural trend. The Glärnisch ridge is segmented into castellate ledges and towers susceptible to sporadic collapse. Extensive late Pleistocene rock avalanche deposits cover the bottom of the Linth Valley between Schwanden and Netstal; they originated by failure of oversteepened cliffs in the vicinity of Glarus (Oberholzer, 1900).

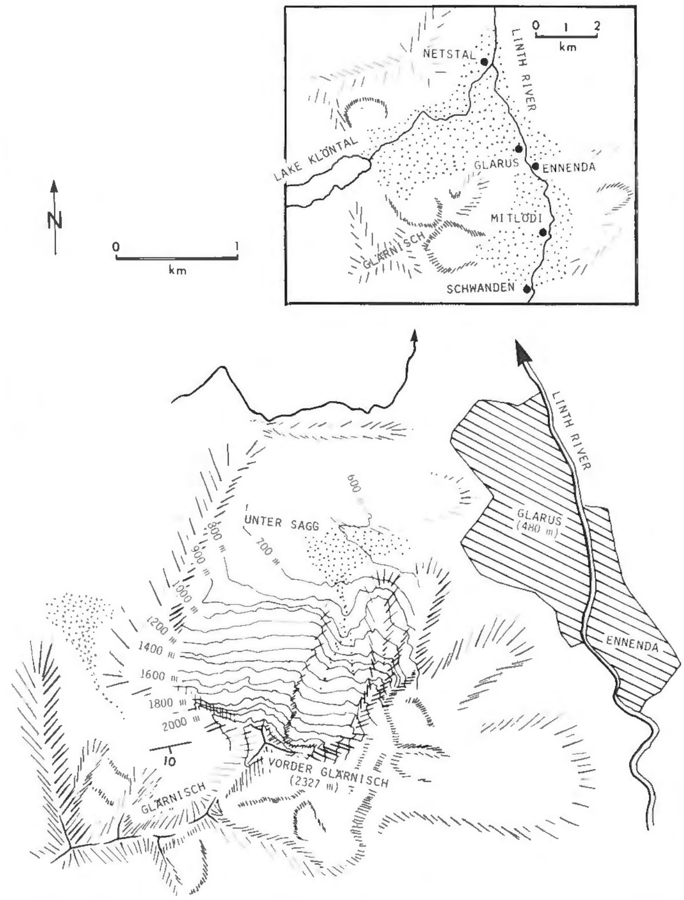


Figure 104: Index and sketch map of the Linth Valley near Glarus; most of the steep-walled valley and its tributary basins are covered with late Pleistocene rockslide debris (dots). Historical rockfall activity along the Glärnisch and other carbonate-shale cliffs of the area is concentrated along joint-controlled bedrock ravines and scarp faces.

The most unstable group of limestone towers formerly known as 'Drei Schwestern' was located just below the Vorder Glärnisch ridge (2327 m), an eastern buttress of the Glärnisch Massif. Collapse of the three spurs was probably preceded by slow subsidence of recessive calcareous slates along the scarp face below the Vorder Glärnisch, suggested by the reversal of a general southerly dip of the strata to a gentle northerly dip.

On 11 November 1593, a strong earthquake shook the canton of Glarus. As a direct result of the earthquake the central tower of Drei Schwestern collapsed; carbonate blocks tumbled some 1300 m down a narrow chute onto the terrace of Unter Sagg. Although blocks scattered over a considerable area, damage to the few buildings on the terrace was minimal. However, in the months that followed there were additional rockfalls from the same section of the cliff. On 2 July 1594, booming sounds akin to cannonfire were heard in the valley. The rock wall below the two remaining pinnacles began to bulge forward noticeably, and behind the pinnacles a large composite fissure opened over a length of 300 m. At this time people abandoned the terrace below. Their fears were justified: next morning the lowermost of the towers toppled and $0.1 \times 10^6 m^3$ of rock plummeted onto the alpine meadows and chalets, burying the springs that supplied water to Glarus. Because of the great height of fall and the steep gradient of the chute, individual blocks scattered over a large area (Heim, 1932, p. 83). Although direct damage was not great, nine days later, on 12 July 1594, the buried springs burst forth explosively from below the debris blanket and a flow of rubble engulfed the town of Glarus, inflicting considerable material damage (Baltzer, 1873, p. 32-33).

A series of rockfalls from the Glärnisch and other mountains in the region occurred during earthquakes between 1701 and 1703; this activity culminated in a small rock avalanche in 1706. Other minor falls from the cliff have occurred since then, for example on 30 July 1886, 16 April 1922, 18 April 1929, and 9 February 1930. These falls were noticed because each time an echo, resembling heavy thunder, reverberated from the walls of the narrow Klöntal gorge (Oberholzer, 1933, p. 575).

Today the Klöntal and its sparkling lake are a favourite tourist destination and the starting point for hikes and climbs into the Glärnisch. Relatively few buildings have been erected below the north face of the range although development pressures in the Linth Valley are enormous.

Val de Bagnes (A34)

Location: Martigny, Valais (Wallis), Switzerland (C3)
Date(s): 4 June 1595 (also 580 A.D., 13th or 14th century, 7 August 1549, September 1640, 16 June 1818, 28 June 1894, 25 July 1895, 17 June 1898)

Val de Bagnes is a southern tributary valley of the upper Rhone River (Fig. 105). It is drained by the Dranse River which heads in extensively glaciated upland ranges and steep bedrock basins carved into metamorphic rocks of the Pennine core complex. The Dranse River joins the Rhone River near the town of Martigny (470 m).

Several large debris cones constrict the natural channel of the lower Dranse River; the largest of these is the cone of the Merdenson Torrent near the village of Vollèges (840 m). In the 13th or 14th century, the community of Curalle (or Curru) located near the head of the cone was destroyed by debris flows (Montandon, 1933, p. 286-287). The source of these flows was the intensely fractured bedrock slope of the Pierre Avoi Massif (2335 m). The Merdenson and other torrents have continued to contribute coarse bedload to the Dranse River to the present day. The most destructive historical mass movements in the valley have been associated with ice floods (jökulhlaups) from glaciated upland basins.

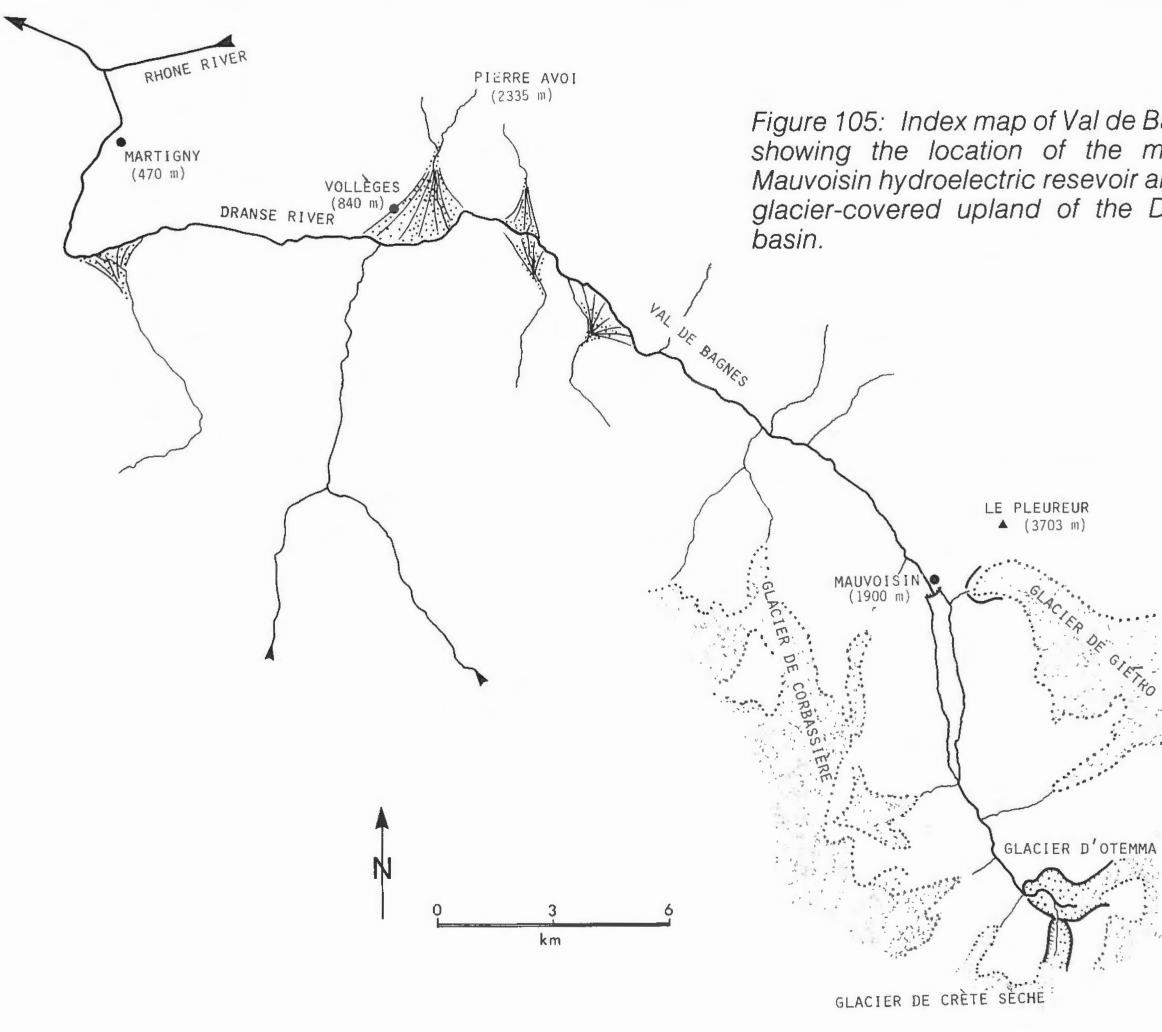


Figure 105: Index map of Val de Bagnes showing the location of the modern Mauvoisin hydroelectric resevoir and the glacier-covered upland of the Dranse basin.

The most serious debris floods of the Val de Bagnes originated at the foot of the Giétro Glacier, a hanging glacier above Mauvoisin (1900 m). During the Little Ice Age icefalls from the advancing glacier repeatedly blocked the flow of the Dranse River. Generally, a huge cone of ice accumulated in the Mauvoisin gorge whenever the snout of the glacier reached the steep defile at 2300 m, east of Mauvoisin (Fig. 106).

Outbursts of impounded water across this cone of remolded ice ('glacier rémanié') probably triggered floods in 580 A.D. and on 7 August 1549. The most dramatic buildup of ice, however, occurred during the advances of the Giétro Glacier in the last decade of the 16th century. In the spring of 1595 an enormous quantity of ice slid away from the snout of the Giétro Glacier and the level of the lake behind the Mauvoisin ice cone began to rise rapidly. On 4 June 1595, abnormally high air temperatures weakened the conical mass of ice and the lake emptied within an hour. Millions of cubic metres of water surged down the valley picking up boulders and loose debris along the way. The frontal wave of ice, water and debris swept away farms and roads. After crashing through the bedrock gorge above Martigny the deluge overwhelmed the town itself. The catastrophe claimed a total of 140 lives, half of them in Martigny; large tracts of the ravaged Val de Bagnes remained barren for many years. A similar but much smaller ice flood in the Val de Bagnes was registered in September 1640.

The Giétro Glacier again accumulated its cone of remolded ice across the gorge of Mauvoisin during the early 19th century (Fig. 106). In spring of 1818 the normally torrential Dranse River diminished to a trickle, an unusual situation for the season, immediately noticed by the townspeople in Martigny. A group of them went up the valley to investigate. Upon reaching Mauvoisin, they were confronted with the gigantic wedge of ice that by now had grown more than 100 m above the channel of the river. Aware of the catastrophe of 1595, the authorities immediately called upon Ignaz Venetz, an energetic public engineer, to excavate a tunnel through the ice and thus forestall the eventual overtopping and failure of the dam. Venetz and his 50 helpers worked frantically from both ends of the cone, using experience gained during the construction of rock galleries along the Simplon Pass route a few years earlier. Meanwhile, the lake continued to rise at a rate of about 0.5 m a day; the workers in the tunnel were threatened by icefalls from the glacier and by waves pounding against the upstream side of the cone. After about a month of incredibly hazardous work, the tunnel, now 200 m long, was nearly complete. Unfortunately, the upstream and downstream approaches did not meet at precisely the same level and the floor of the upper tunnel had to be lowered 6 m. The adjustments, which resulted in a downward tapering cross-section, were barely completed when the first rush of water from the lake entered the tunnel. Blocks of ice soon jammed the lower outlet of the

Figure 106: The ice cone at the foot of the Giétro Glacier at the narrows of Mauvoisin as depicted by H.C. Escher in 1818 (from Bachmann, 1978).

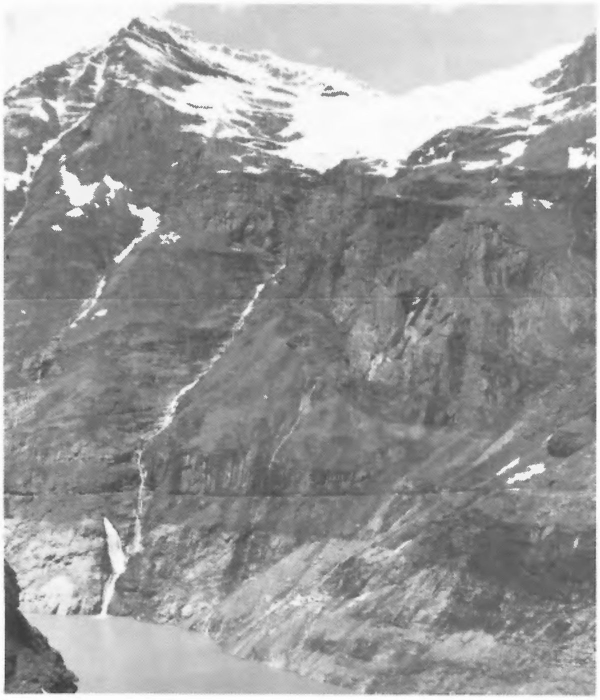


Figure 107: View of the Mauvoisin reservoir and Giétro Glacier in 1981 from the same position as that occupied by Escher in 1818. (GSC 204166-Q)

In the late 19th century a general retreat of Alpine glaciers eliminated the hazard of the Giétro ice cone, but created another set of problems farther upstream. Downmelting of the Crète Sèche and Otemma glaciers created a temporary ice-marginal lake at the confluence of these glaciers. This lake drained rapidly on 28 June 1894, on 25 July 1895, and on 17 June 1898. These floods released up to $1 \times 10^6 m^3$ of water into the boulder-strewn channel of the Dranse, destroying bridges and mills in the Val de Bagnes. Eventually the impounding morainal ridge between Crète Sèche and Otemma glaciers was artificially breached by a V-shaped cut to provide unimpeded runoff for the torrent of the receding glacier.

Between 1951 and 1957 the Mauvoisin hydroelectric dam was erected at the narrows that used to be the site of the threatening ice cone (Fig. 107). In recent years the snout of the Giétro Glacier again has approached the upper rim of the gorge and fragments of the glacier have tumbled into the reservoir. Its movement is now being monitored continuously by a cable device attached to the glacier. The measured displacement of the ice is recorded and relayed electronically to Mauvoisin. This monitoring system permits a timely lowering of the reservoir if major slippage of the ice mass is observed; however, even with a full reservoir an ice mass in excess of $0.7 \times 10^6 m^3$ would be required to generate a wave that would overtop the dam (Röthlisberger and Aellen, 1970; Röthlisberger, 1974).

tapering conduit. Obstruction of the ice tunnel by blocks of ice became so severe that on 16 June 1818 the whole downstream sector of the cone and adjacent morainal ridges collapsed. Within half an hour some $10 \times 10^6 m^3$ of water rushed through the widening gap. As in 1595, the ice flood burst through the narrow canyon at Mauvoisin and spread a thick layer of rocky debris over a small alluvial plain below. Along the steep reaches of the valley the flood wave changed into a large debris flow that swept away houses, bridges, crops, and trees. The loss of life (50 people) was relatively light compared to what it might have been, because every settlement downstream had been alerted to the imminent collapse of the ice dam via a series of signal stations placed along the sides of the valley (Pictet, 1818; Beattie, 1836, v. 1, p. 43-48).

To prevent another potentially catastrophic buildup of ice at Mauvoisin, Venetz continued his work. Between 1822 and 1824 he channeled runoff from nearby ravines onto the cone, a measure that helped reduce the buildup of ice. He also widened the channel of the Dranse with wooden sills to prevent individual blocks of ice from accumulating and thus blocking the flow of the river (Richter, 1889b). In 1842, when Forbes (1845, p. 264) visited the valley these works were still being maintained by the cantonal engineering service. A permanent tunnel through bedrock along the endangered section of the Mauvoisin gorge was contemplated, but not constructed because of its cost.

Piuro — Plurs (A35)

Location: Val Bregaglia, Chiavenna, Lombardia, Italy (E2)
Date(s): 4 September 1618 (also 1675, 1760, 1858)

The Val Bregaglia, east of the old city of Chiavenna (333 m), is a narrow, steep-walled valley cut into gneisses and amphibolites of the Pennine core complex. The Mera River flows westerly, parallel to the steeply dipping foliation of the bedrock. High-gradient torrents and ravines drain mountain ranges up to 2800 m on both sides of the river (Fig. 108). Sagging bedrock slopes and debris cones locally convert the Mera River into a potentially dangerous torrent. According to tradition, a Roman town located upstream from Chiavenna was destroyed by a debris flood of the Mera River in the 8th (?) century.

Despite the rugged setting, the small medieval town of Piuro (400 m), approximately 4 km east of Chiavenna, became one of the most prominent cities in the region. Its economic dominance stemmed from a monopoly in the manufacture and trade of heat-resistant pots, pans, and pipes made of the local Lavezzi stone (also known as Ofenstein, lapis olaris, and pietra olare). This rock, a grey-green metamorphic serpentine, was mined in several adits on the north slope below Prato del Conte, south of Piuro. At the time of the great disaster of Piuro a broad debris-mantled bedrock terrace above the mine adits accommodated several

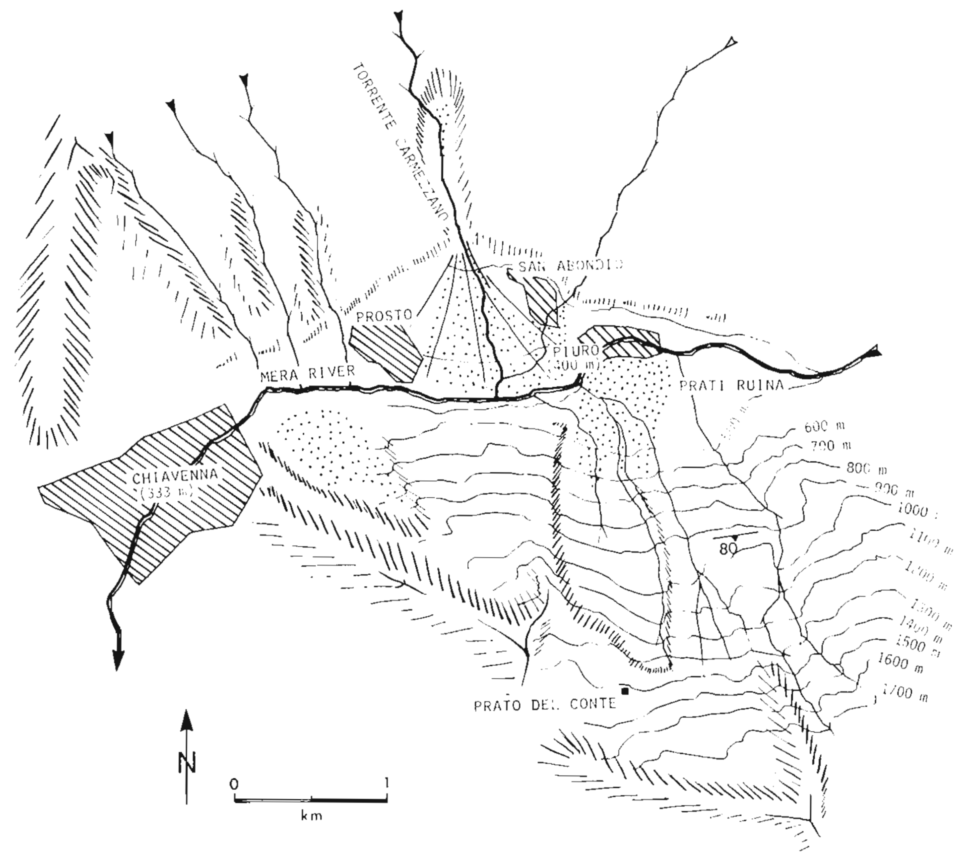


Figure 108: Sketch map of the steep scarp slope south of Piuro, the source of the catastrophic debris-rock slide of 1618. The debris lobe of 1618 and the debris cone of the Carmezzano Torrent have severely limited the development of the narrow and flood-prone Mera valley above Chiavenna.

homesteads. Piuro itself consisted of merchants' stores, markets, exclusive residences, villas and palaces and during the late summer months was filled with visitors from all over Europe.

At least ten years prior to the disaster that was to strike Piuro, mining activity and clearing of the uplands initiated a process of accelerated creep on the slope above the adits. Surface cracks appeared along the terrace of Prato del Conte at this time. However, the main failure of a thick layer of bedrock and colluvium above the adits was triggered by an intense rainstorm between 25 August and 3 September 1618. After a week of almost uninterrupted precipitation the clouds lifted on the morning of 4 September, and the prosperous city bathed in sunshine. Along the eastern sector of the mining zone a small rock-debris avalanche buried a few houses, but this did not incite any great panic, because such rockfalls had been experienced before. However, the peasants working on the Prato del Conte terrace 500 m above the town witnessed some unfamiliar events: the ground underneath their feet vibrated with increasing intensity; there were snapping and rumbling noises from within the mountain; the bees left their hives and cows became restive. One of the men, while cutting a tree, saw a huge crack propogate across the slope. Immediately he scrambled into Piuro to spread the news. The citizens, reluctant to abandon their cherished surroundings, reacted downright hostile and the messenger of doom was lucky to escape physical abuse! Nevertheless, by nightfall a large apprehensive crowd assembled at the cathedral. At this moment a deep breakaway scarp opened and 3 to 4 $\times$ $10^6 m^3$ of rock and surficial debris failed along a composite rupture surface dipping approximately 30° towards the town. Within seconds, the rock-debris avalanche buried some 200 buildings and killed approximately 1200 people (Fig. 109, 110). The Mera River was blocked for more than an hour and the citizens of Chiavenna fled to the mountainsides, anticipating a flood. This flood, fortunately, did not materialize when the river overflowed the slide mass (Heim, 1932, p. 140-141; Montandon, 1933, p. 300). The catastrophe marked the end of Piuro. The memory of its fall was kept alive by legends and a famous illustration by A. Scheuchzer.

Immediately to the west of Piuro, is the bouldery cone of the Carmezzano Torrent with the communities of San Abondio and Prosto on its flanks (Fig. 108). This cone experienced debris flows in 1675 when Prosto was buried 'to the chimneys', in 1760 when San Abondio was destroyed, and in 1858 when a hamlet by the name of Santa Maria disappeared (Beattie, 1836, v. I, p. 120; Montandon, 1933, p. 308 and 320). An isolated church tower, surrounded by boulders of granitic gneiss, bears witness to these disasters (Fig. 111).

For many years, the surface of the Prati Ruina at Piuro has hosted vineyards and a few service chalets (Fig. 110). Recently, some of the ancient walls and buried artifacts of Piuro have been excavated; the retrieved objects are now exhibited in a small museum at San Abondio. New housing (Burgo Nuovo) extends onto formerly uninhabited parts of the Prati Ruina and Carmezzano cones. The bouldery channel of the Carmezzano Torrent is flanked by a large stone dyke. Lavezi (or Laveggi) stone is now refined into decorative panels and pots in a small factory located on the periphery of the debris cone east of Prosto.

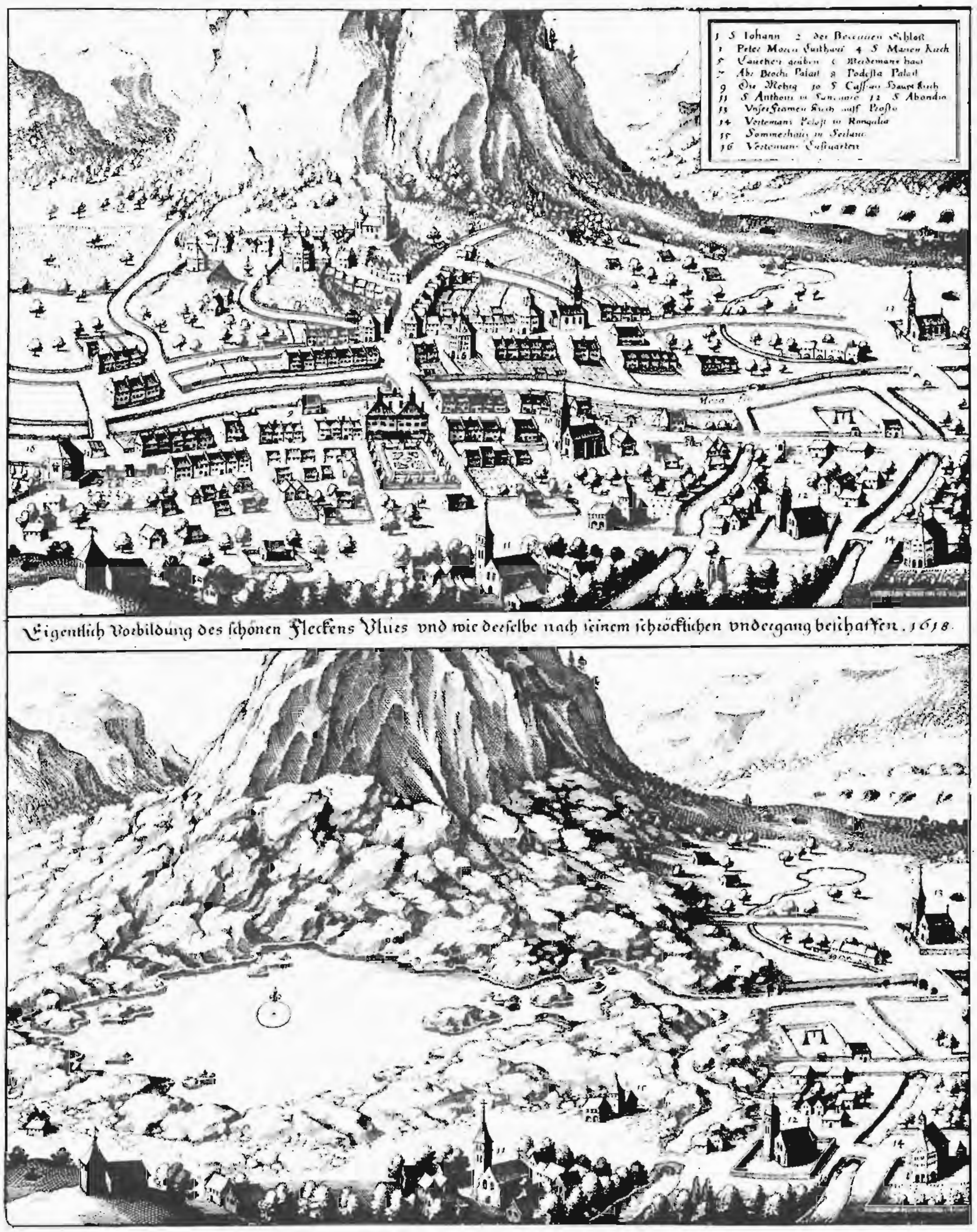


Figure 109: The town of Piuro-Plurs before and after the 1618 catastrophe, as depicted by A. Scheuchzer in 1746.

Figure 110: Southward view onto the slightly conical surface of Prati Ruina and the detachment zone of the 1618 debris-bedrock slide (on the right). The town of Piuro is at the bottom. (GSC 204166-R)

Figure 111: Church tower projecting from lobes of gneissic rubble on the cone of the Carmezzano Torrent. (GSC 204166-S)

Tête Noire (A36)

Location: Giffre Valley, Cirque du Fer à Cheval, Haute Savoie, France (B2)
Date(s): 21 February 1602

Cirque du Fer à Cheval is an imposing amphitheatre of flat-lying carbonate and shale formations in the Helvetic cover complex near the head of the Giffre Valley (Fig. 112). From the meadows of Plan du Lac (950 m) stepped walls of limestone, separated by ledges of recessive shale, rise to the glaciated crest of Le Cheval Blanc (2830 m). Intensely fractured and semi-detached spurs of limestone mark the lower border of perennial snow and ice fields of the summit ridge (Fig. 113). Tête Noire (2167 m), the most conspicuous of these bedrock promontories, is partly veiled by the mist of cascading glacial torrents.

On 21 February 1602, a portion of the Tête Noire face collapsed along a slightly concave and almost vertical fracture surface. Several million cubic metres of disintegrating carbonate rock first hurtled into a funnel-shaped ravine and then fanned across the flats along the Giffre River. The rock avalanche annihilated the village of Entre-Deux-Nants and damaged several nearby hamlets, killing 29 people. Precursory rockfalls may have forewarned some of the inhabitants of the upper valley who apparently had abandoned their homes prior to the catastrophe. Carbonate debris also choked the flow of the Giffre River, and a subsequent debris flood devastated much cultivated land downstream from Plan du Lac (Montandon, 1933, p. 299-300).

Today carbonate slabs still protrude through the partly forest-covered surface of Plan du Lac which hosts a major restaurant-hotel complex and attracts many visitors during the summer months. Downstream from the rock avalanche

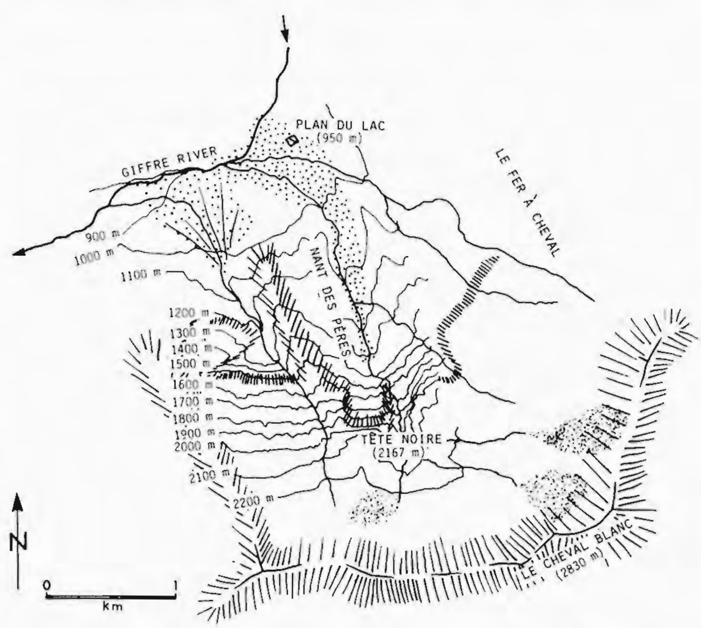


Figure 112: Sketch map of the Cirque du Fer à Cheval and the historical rock avalanche lobe below Tête Noire.

deposits, a large debris cone below the bedrock cliffs west of Tête Noire also adds debris to the Giffre River, still characterized by an extremely coarse bedload. Several precarious pinnacles along the Tête Noire cliff maintain their threatening pose above the stunning backwall of the valley.

Figure 113: View of the vertical detachment wall of Tête Noire from Plan du Lac. (GSC 204166-T)

Salvan (A37)

Location: Trient Valley, Valais (Wallis), Switzerland (B2)
Date(s): 31 January 1635

The communities of Salvan and Marecottes (1200 m) huddle on a south-facing bedrock terrace (934 m) above the precipitous gorge of the Trient Torrent, a western tributary of the Rhone River. The terrace is framed by the summit ridge of Le Luisin (2785 m) which consists of intensely fractured schist and gneiss of the Helvetic basement.

On 31 January 1635, a rock avalanche (mixed with snow?) broke away from this ridge and demolished part of the village below (Montandon, 1933, p. 302).

Today the rock terrace of the Salvan-Marecottes is almost completely built over and except for some highly irregular topography in the ski village of Marecottes there are few hints of the rock avalanche of 1635. Numerous snow bridges in the broken bedrock terrain below the summit ridge now protect the community of Salvan snow avalanches.

Chablais (A38)

Location: Haute Savoie, France (B2)
Date(s): 11 April 1635 (also 29 July 1715, 12 March 1943)

The Chablais includes a major part of the foothills and front ranges of the western Alps known as Préalpes Médianes (Fig. 114a). From the south shore of Lac Léman (430 m) this pleasant agricultural district rises southward to ridges and peaks with elevations up to about 2000 m. Meadows, farms, and forests interlace along gently sloping mountainsides controlled by the northeasterly trending bedrock structures of allochthonous panels of Pennine and Helvetic shale-carbonate successions. Dip slopes or scarp faces of carbonate ridges pierce the more recessive shale formations, particularly along transverse fault zones. The open valleys are commonly floored by late Pleistocene surficial deposits. Many of the shale slopes show signs of incipient instability or creep.

Sometime during the early Middle Ages, a medieval village at the foot of Pointe d'Autigny was buried when some $7 \times 10^6 m^3$ of massive limestone near the hinge zone of an anticlinal ridge along the Dranse River collapsed (Montandon, 1933, p. 285).

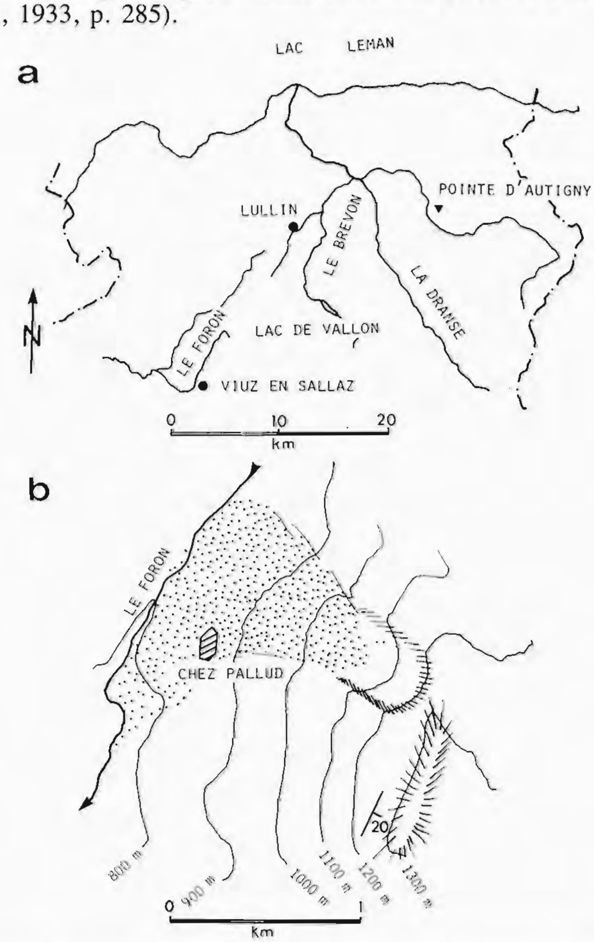


Figure 114: a) Index map of the Chablais foothills region south of Lac Léman (Lake Geneva). b) Sketch map of the slide mass of the 'Déluge de Viuz'.

On 11 April 1635, triggered by an earthquake (?), a segment of a dip slope composed of calcareous shale and limestone failed and swept into the village of Lullin, obliterating 20 houses and killing 64 people (Mougin, 1914, p. 252-253). Today, there is no evidence of the slide mass which possibly underlies the eastern outskirts of the present community of Lullin. The valley below Lullin is also known for its chronic slumps of surficial deposits which have required control work along torrents, surface drainage, and revegetation of colluvial terraces.

On 29 July 1715, a relatively gentle scarp slope, composed of southeast-dipping calcareous shale, evaporite, and clays, failed along a 350 m wide semicircular crown crack at an elevation of 1200 m east of the Foron Valley (Fig. 114b). After initial failure the slide mass of $2.5 \times 10^6 m^3$ accelerated and entered the Foron Valley (800 m), where it buried three hamlets consisting of 20 houses under 7 m of debris (Montandon, 1933, p. 305-306). A bowl-shaped depression marks the source of this 'Deluge de Viuz'. Several hamlets (e.g. Chez Pallud) have developed on or near the slide mass which is now mantled by a dense protective forest, in contrast to the open fields and pastures of the surrounding countryside.

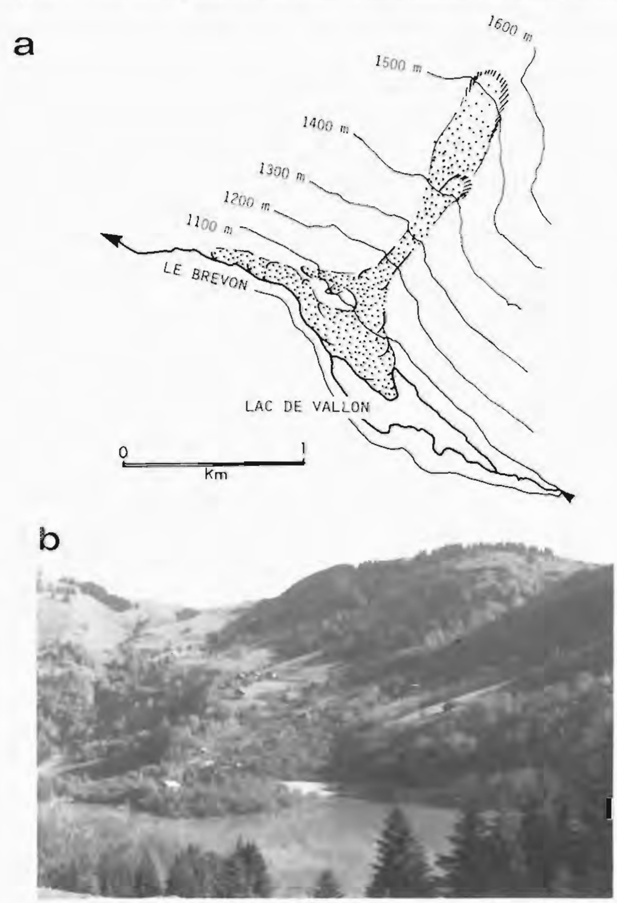


Figure 115: a) Sketch map of the deposits of the slow slump-debris flow that stemmed the flow of the Brévon Torrent in 1943. b) View of Lac de Vallon and the 1943 debris lobe; note new residential buildings on the slide deposits (GSC 204166-U).

On 12 March 1943, a cultivated slope in the upper Brévon basin showed signs of outward bulging; ground cracks soon defined a distinct debris lobe descending at a rate of 200 m/day. Set in motion by snowmelt infiltration the failure involved $2 \times 10^6 m^3$ of intensely deformed limestone and shale. The frontal part of the lobe soon impounded the west-flowing Brévon Torrent (Fig. 115a) resulting in a lake that eventually extended 1 km upstream (Lac de Vallon). The lake overflowed without doing much damage (Moret, 1943), and has not changed significantly since. The slide mass came to rest and now hosts new residential buildings (Fig. 115b).

Bourg-St. Maurice (A39)

Location: Isère Valley (Tarentaise), Savoie, France (B3)
Date(s): May 1636 (also 163 A.D., 1363, 8 September 1579, May 1636, 30 September 1732, 3 June and 10 August 1868, 31 March 1981)

The town of Bourg-St. Maurice (800 m), located at the western approaches of the Petit St. Bernard Pass, has been the centre of the upper Tarentaise since Roman times. The town and surrounding villages cluster near a major bend of the Isère River and at the confluence of four major tributary torrents from the north: the Arbonne (and Nantet), the Charbonnet, the Versoyen, and the Reclus (Fig. 116). The upland ridges in the vicinity rise to almost 3000 m and are underlain by intensely deformed calcareous shale, evaporites, flysch, and slate of the Helvetic (Dauphinois) and Penning (Briançonnais) cover complexes. Southeast-facing dip slopes are particularly prone to sagging and slumping. The Arbonne and Reclus torrents have accumulated huge debris cones attesting to voluminous mass movements in the past. The most serious historical failures in the area have occurred in slate-evaporite terrain underlying the lower Arbonne basin.

An old inscription, unearthed in the vicinity of Bourg-St. Maurice, states that in 164 A.D. the Roman Emperor Verus issued an order to '. . . rebuild, out of his own money, the roads washed away by the powerful torrents, the river control works, bridges, temples, and baths of Bergintrum. . . '. Bergintrum was located close to the modern site of Bourg-St. Maurice and it is probable that the Arbonne was one of the torrents ravaging the Roman installations (Montandon, 1933, p. 279; Mougin, 1914, p. 746).

The medieval town of Bourg-St. Maurice expanded from a cluster of buildings at the axis of the Arbonne cone. Commercial exploitation of salt occurrences in the upland basin and a flourishing saltern near the town soon led to extensive deforestation of upland forests. Ensuing unbridled runoff, enhanced infiltration, and deep erosion along tributary torrent channels profoundly changed the stability of the bedrock slopes in the Arbonne basin. In 1356 the Nant Blanc, a northern branch of the Arbonne, suddenly abandoned its regular channel and drained through a subterranean conduit in gypsiferous limestone terrain. After a period of apprehension, the residents of Bourg-St. Maurice soon forgot this peculiar incident. Then, in 1363, the Nant Blanc forcefully emerged in its old channel. Massive ground subsidence,

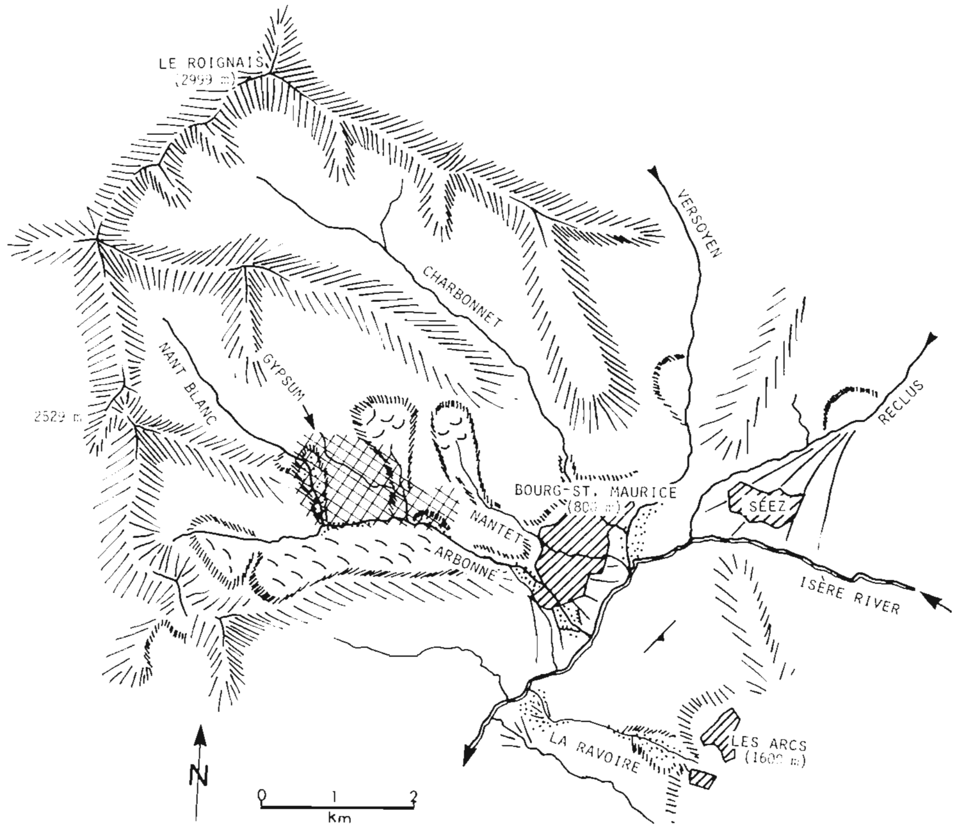


Figure 116: Sketch map of the environs of Bourg-St. Maurice showing large incipient bedrock slumps in shale-evaporite terrain in the Arbonne basin. Criss-cross pattern indicates outcrop of gypsum. Note the location of the large ski resort of Les Arcs on the high bedrock terrace south of the Isère River and the track of the 1981 debris flow along the Ravoire.

slumping and temporary blockage of the lower gorge generated huge pulses of debris debouching onto the cone of the Arbonne and laying to waste much of Bourg-St. Maurice. The town was reconstructed 200 m north of the debris-strewn area, but still on the Arbonne cone. On 8 September 1579, debris flows from the Arbonne and flooding of the Isère River endangered the lower parts of the rebuilt town.

The most destructive cycle of debris flows from the Arbonne basin occurred between 1630 and 1636, when major slope movements in the uplands blocked several tributary branches of the torrent. At first the primitive dykes flanking the torrent channel contained the flows of water and debris. Then, in May 1636, massive debris flows burst through the protective works and demolished 52 houses in Bourg-St. Maurice (Fig. 117a). Reconstruction again shifted away from the central sector of the cone (Montandon, 1933, p. 301-302).

On 30 September 1732, debris flows, triggered by intense rain and unseasonally early snowmelt, overwhelmed the westernmost section of the town and blocked the Isère River. When this barrier failed a floodwave swept downvalley, destroying bridges and mills. Effects of this floodwave were felt as far away as Grenoble (Mougin, 1914, p. 751).

The next destructive cycle of the Arbonne was initiated by rapid delayed snowmelt in the spring of 1868. On 3 June 1868, slumps along the Nantet Torrent developed into a massive debris flow, and on 10 August 1868, a rainstorm triggered a few more pulses of debris.

Towards the end of the 19th century erosion in the Arbonne basin was reduced by reforestation and construction of check dams in the gorge (Mougin, 1931, p. 531-539). In the 20th century Bourg-St. Maurice has not been victimized by major debris flows. Recently, the town again has spread towards the channel of the Arbonne. However, the torrent now is flanked by high stone dykes and two strips of protective forest (Fig. 117b). Incipient instabilities on the mountainsides near Bourg-St. Maurice tend to change into flows during periods of heavy rain and rapid snowmelt, as for example in April 1970 (Jail and Vivian, 1971, p. 493-494).

A map depicting slope hazards in the vicinity of Bourg-St. Maurice was published recently following the construction of large ski facilities at Les Arcs (1600 m) across the Isère Valley (Pachoud, 1979). Hotel complexes, logged ski runs, and paved roads now cover a subalpine terrace between 1600 and 1800 m south of the town. Inconspicuous ravines below the ski station have cut deeply into the slaty bedrock and, during an abrupt snowmelt on 31 March 1981, some $0.3 \times 10^6 m^3$ of debris was propelled down the Ravoire gorge and fanned across the Isère Valley. Expensive control works are being built to counteract this new hazard, which again has been brought on partly by human activity.

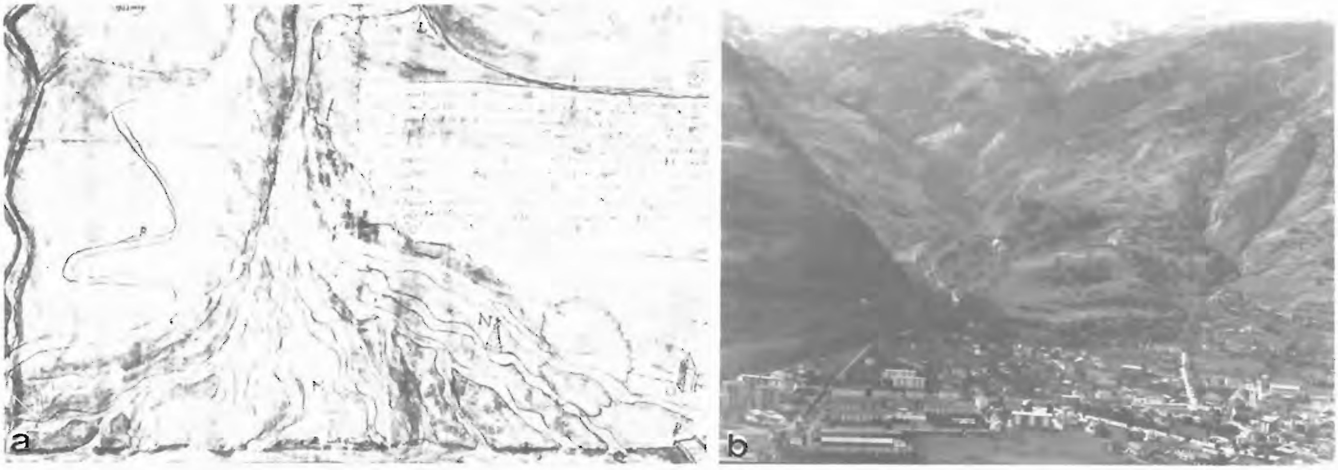


Figure 117: a) Drawing of the Arbonne cone made after the debris flow disaster of 1636 (from Palluel-Guillard, 1978). b) View of the Arbonne cone and Bourg-St. Maurice in 1981; the torrent channel on the cone is flanked by a protective strip of forest (lower left) (GSC 204166-V).

Randa (A40)

Location: Matter Valley, Wallis (Valais), Switzerland (C2)
Date(s): 13 January 1636 (also April 1737, 1786, 27 December 1819, 31 January 1848, 1865, 25 August 1973)

The Matter Valley is a narrow U-shaped trough carved into metamorphic rocks of the Pennine core of the western Alps (Fig. 118). Rugged glacier-clad mountains loom high over small settlements and well known tourist centres (e.g. Zermatt). High-gradient glacial torrents, hanging glaciers, rockfall lobes, sagging slopes, and persistent snow avalanches limit the extent to which the Matter Valley can serve human needs.

One of the largest debris cones flanking the torrential Matter Vispa is the Dorfbach cone. Crowded against the upstream flank of this debris cone, the community of Randa (1429 m) faces the bulging ice front of the Bis Glacier. This glacier originates in a bowl-shaped accumulation zone below the ice-covered pyramids of the Weisshorn (4505 m) and Bishorn (4134 m) on the western crest of the Matter Valley.

During the Little Ice Age Randa survived several severe ice avalanche disasters which originated by collapse of overhanging ice above the main glacier. The most serious catastrophe struck on 13 January 1636, when a large mass of ice collapsed on the Weisshorn face and swept down the Bis Glacier demolishing most buildings in Randa, and killing 36 people. Two smaller ice avalanches in April 1737 and 1786, inflicted only minor damage on the settlement which by then huddled against the eastern valley wall, 80 m above the Vispa Torrent.

The big ice avalanche of 27 December 1819, was documented by I. Venetz, whose written account was pulished by Agassiz (1840, p. 158-161). More than $10 \times 10^6 m^3$ (?) of ice apparently became unhinged along a deep transverse crevasse on the face of the Weisshorn. The stream of broken ice and snow first descended the Bis Glacier, struck the valley floor, and then climbed to the apex of the Dorfbach cone. Although most of the village escaped the direct impact of the avalanche, a powerful air blast preceded the stream of ice and dislodged millstones, uprooted trees, hurled blocks high above the valley floor, and blew the roofs off several buildings. In addition, Venetz noted that

'. . . strangely several barns located on the bank opposite (Randa), below the glacier, although practically covered with debris, have not suffered any damage because they were protected from the force of the wind. It is even more remarkable that only two persons were killed, although several families were blown away with their houses and buried under snow. . .' (Agassiz, 1840, p. 160).

Smaller nondestructive ice avalanches were registered on 31 January 1848, and in 1865. With incredible tenacity the community of Randa perservered on its plot, daring avalanches of snow and ice, and avoiding the ravages of nearby glacial torrents.

Recently, the Bis Glacier and its tributary ice fields began to grow again after having shrunk for more than a century. Residential construction at Randa meanwhile has invaded the path of the historical ice avalanches (Fig. 119). On 25 August 1973, a slab of ice with a volume of close to

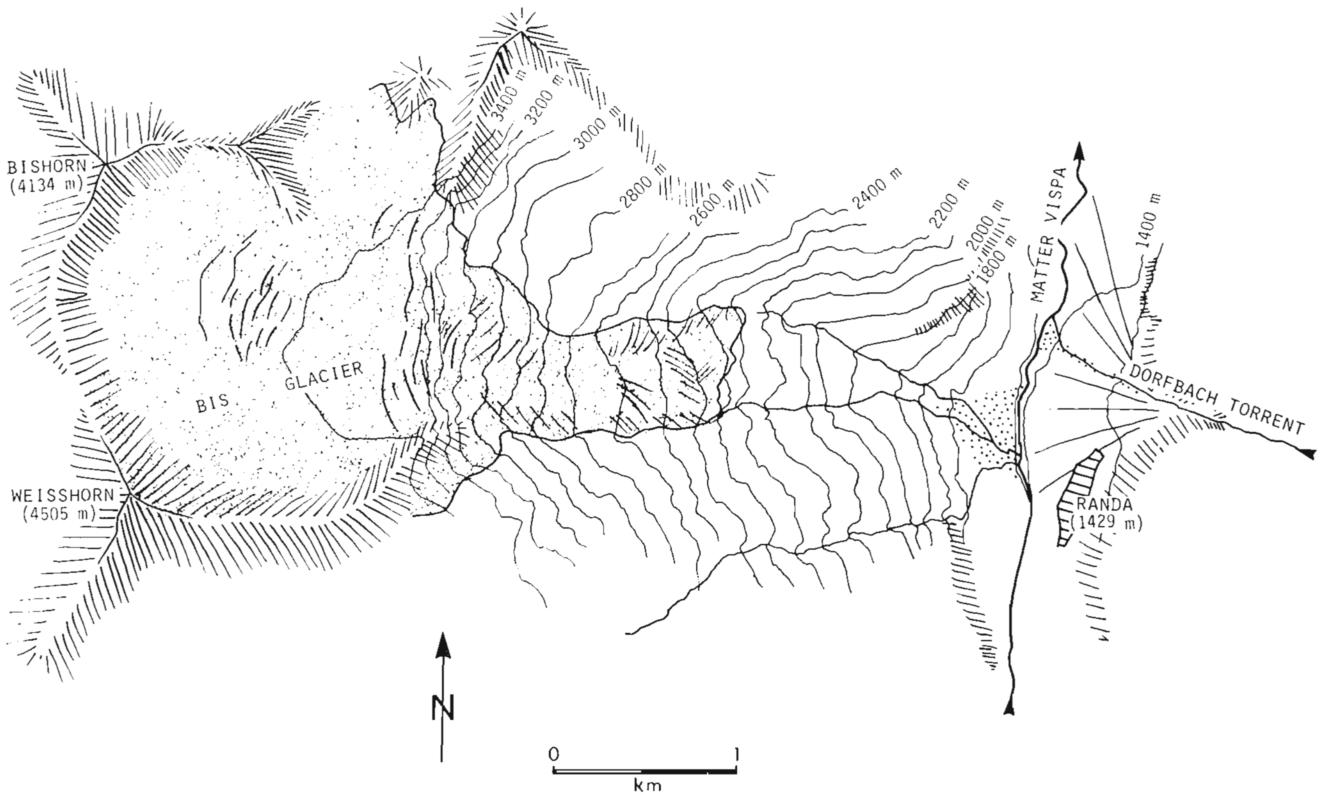


Figure 118: Sketch map of the Bis Glacier and the village of Randa in the Matter Valley. Ice avalanches from the cliffs of the Weisshorn have repeatedly swept over the cone of the Dorfbach Torrent.

$0.5 \times 10^6 m^3$ began to slip away from the steep flank of the Weisshorn. Fortunately, it disintegrated into three harmless icefalls, which were watched with some apprehension from the village (Röthlisberger, 1979, p. 136-141).

Figure 119: View of the Dorfbach cone and the community of Randa from the snout of the Bis Glacier; note several residential buildings near the axis of the cone in the runout zone of potential ice avalanches (GSC 204166-W).

Antronapiana (A41)

Location: Valle d'Antrona, Piemonte, Italy (D2)
Date(s): 27 July 1642

The town of Antronapiana (908 m) adorns the head of the Valle d'Antrona, a tributary valley west of Valle d'Ossola. An impressive circle of mountains surrounds the head of the valley (Fig. 120). Three kilometres northwest of Antronapiana the sheer east wall of Cime di Pozzuoli (2602 m) exposes a tectonic contact between a lower succession of amphibolite-greenstone and an upper unit of gneisses of the Pennine basement complex. The tectonic contact and the metamorphic foliation dip approximately 40° west-northwest and into the mountainside. Most of the upper rock face is massive (Fig. 121a). It is possible that selective erosion of the greenstone units at the base of the mountain over-steepened the wall of gneisses above.

On 27 July 1642, a major section of the Pozzuoli cliff collapsed along a steep concave fracture zone and a stream of gneissic blocks, totalling $12 \times 10^6 m^3$ in volume, cascaded onto the bottom of the valley. The front of this rock avalanche annihilated parts of Antronapiana and killed 93 people (Montandon, 1933, p. 303).

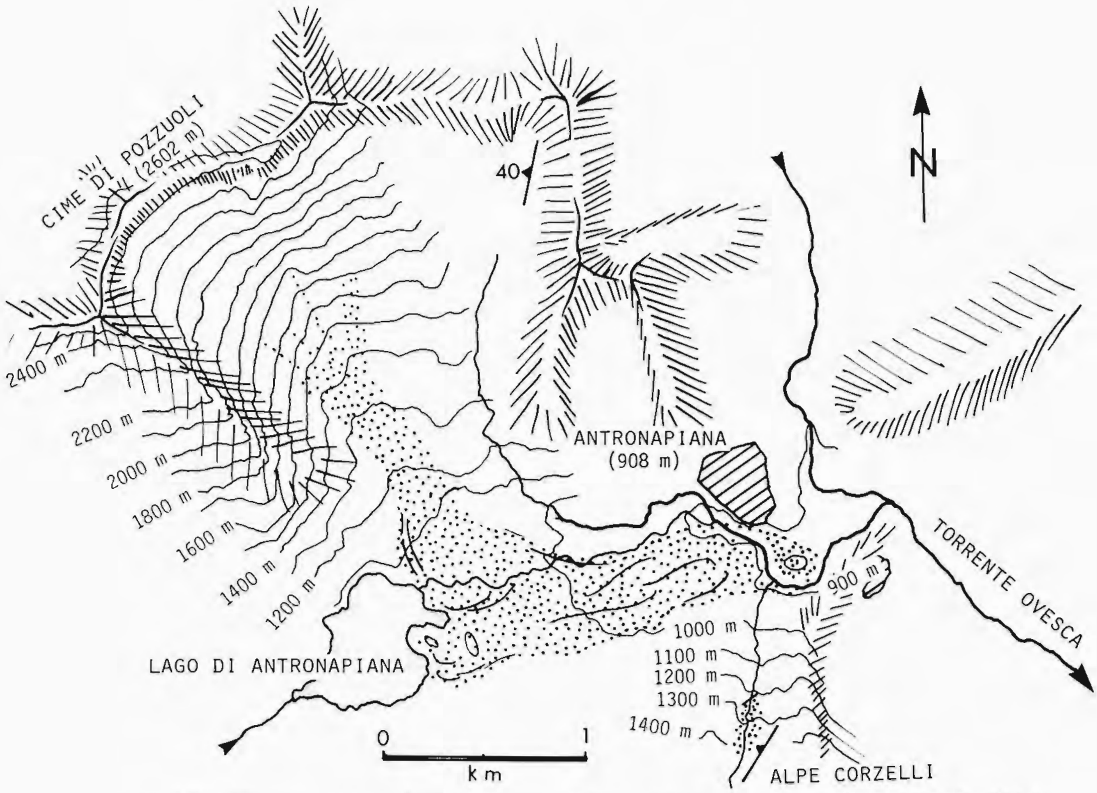


Figure 120: Sketch map of the Antronapiana rock avalanche lobe of 1642 which originated by failure of the gneissic scarp face of Cime di Pozzuoli.

Figure 121: a) View of the scarp face of Cime di Pozzuoli and forest-covered rock avalanche lobe below (GSC 204166-X). b) The town of Antronapiana as seen from the unstable bedrock ledge below Alpe Corzelli (GSC 204166-Y). c) Concrete berm-retaining wall and surface drainage ditch at the toe of the incipient bedrock failure at Alpe Corzelli (GSC 204166-Z).

The slide mass blocked the channel of the east-flowing Torrente Ovesca creating the deep blue Lago di Antronapiana. Overflow of the lake along the northern margin of the blocky lobe has reworked some of the rock fragments on the slide mass but has only slightly modified its original surface which today is mantled by a mixed spruce-larch forest (Fig. 121b). Recently several vacation homes have been built on the rugged surface of the debris stream.

As in several other tributary valleys of Valle d'Ossola precipitous ravines and snow-avalanche chutes sporadically release bouldery debris from instabilities in bedrock and colluvium high on the mountainsides. The Antrona Valley was severely struck by debris flows during the disastrous rainstorm of autumn 1951. One of the ravines, directly across the valley from Antronapiana, displays in its upper funnel-shaped catchment zone a bedrock-and-debris slump of approximately $1 \times 10^6 m^3$ (Alpe Corzelli). As this slide mass sits 500 m above the town it represents an obvious threat. A deeply anchored concrete berm has been placed against the toe of the mass and paved drainage ditches guide seasonal runoff across the incipient slide (Fig. 121c).

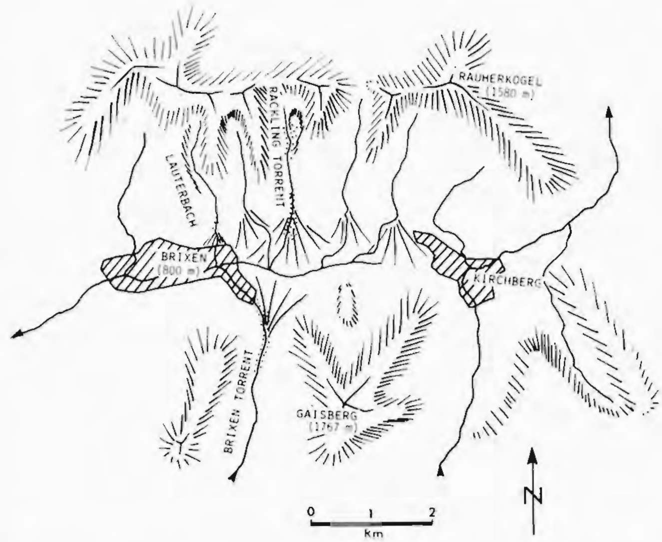


Figure 122: Index map of the Brixen Valley showing the location of the notorious Rackling Torrent and the expanding communities of Brixen and Kirchberg.

Rackling (A42)

Location: Brixen Valley, Tirol, Austria (H1)
Date(s): 1648 (also July 1946)

Most of the rolling upland ridges in the famous ski region of the Kitzbüheler Alpen are underlain by phyllite, calcareous schist, and greenstone of the Austroalpine basement complex. Notoriously unstable bedrock embankments and discontinuous Pleistocene surficial deposits occur in small torrent basins along the sides of major valleys.

One of the main valleys in this region is the Brixen Valley (800 m), which follows the westerly trend of phyllitic bedrock formations (Fig. 122). In the past most of the valley bottom was swampy, thus settlements clustered on the flanks of large debris cones (Fig. 123). During the Middle Ages the cone of the Rackling Torrent hosted the settlements of Rackling and Steinhäring. However, beginning in 1638 debris flows mobilized from the toes of huge bedrock slumps in the upland basin menaced these communities, and within ten years both had disappeared underneath blankets of debris. In the centuries since then the surface of the Rackling cone has remained essentially free of buildings and only recently has the cone been used for agriculture (Strele, 1936, p. 129).

Figure 123: View of the uninhabited Rackling cone and its partly forested catchment area which is characterized by sagging slopes of phyllite; note farm located in the zone between the torrents where forests have a critical protective function. (GSC 204167-A)

Some of the torrents in the Brixen Valley carried destructive mass flows during torrential rains in July 1946. Debris from the Lauterbach and the Brixen torrents engulfed most of the village of Brixen. Farther down the valley temporary blockage of the Brixen Torrent and a subsequent burst of debris claimed five lives.

In recent years open land on the debris cones of the Brixen Valley and nearby areas has been under tremendous development pressure. Most of the torrents with known histories of debris flows have been provided with protective dams, debris retention areas, or arrays of check dams in the uplands. Nevertheless, incipient deep-seated bedrock slumps have been exceedingly difficult to control by these conventional techniques.

Brentonico (A43)

Location: Val Lagarina, Rovereto, Trentino, Italy (F3)
Date(s): 30 December 1648 (also 8 May 1885)

The community of Brentonico (700 m) sits on a gently sloping bench of relict colluvium high above the west bank of the Adige River (150 m). A small torrent incised into unstable shale-carbonate terrain of the eastern flank of Monte Campo (1600 m) crosses this bench just south of Brentonico on a debris fan marked as 'Masere' (= debris) on topographic maps.

On 30 December 1648, a violent rainstorm and accompanying smowmelt reactivated a bedrock slide lodged above the hamlet of Fano, which was located on the debris fan. The debris lobe moved slowly enough for all inhabitants to abandon the threatened buildings before they disappeared underneath a blanket of rubbly debris (Strele, 1936, p. 129). On 8 May 1885, debris movements again claimed five buildings in Brentonico.

In recent years improved access and new recreational development in the area have led to increased construction activity at Brentonico. A few large vacation apartment blocks have sprung up on the eastern flank of the Masere, an area traditionally restricted to agricultural use.

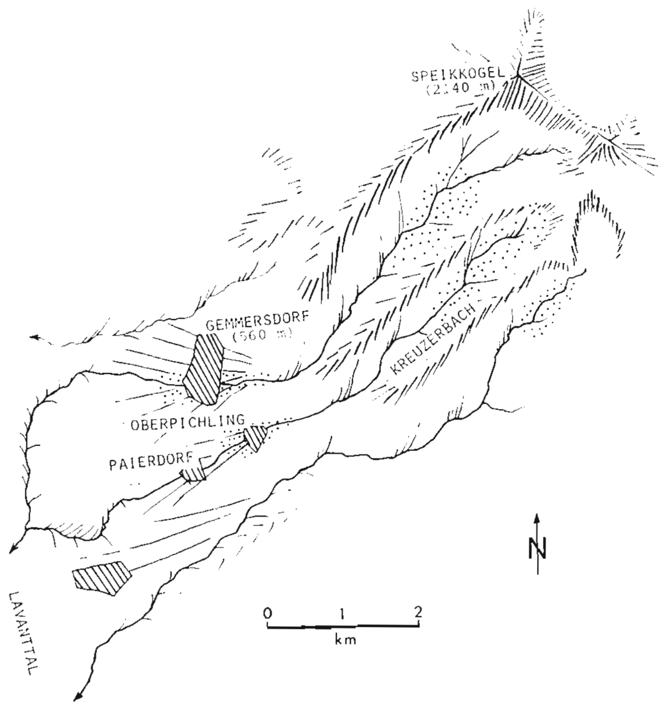


Figure 124: Sketch map of the relict debris cones at the foot of the Speikkogel Massif. Only the uppermost segments of the cones are not incised by the torrents.

Gemmersdorf (A44)

Location: Lavant Valley, Kärnten (Carinthia), Austria (J2)
Date(s): 7 June 1660 (also 7 September 1916)

The village of Gemmersdorf (560 m) is the largest of several agricultural settlements on the gently sloping terrace between the Lavant Valley and the Speikkogel Massif (Fig. 124). The Speikkogel Massif (2140 m) is underlain by metamorphic rocks of the Austroalpine basement complex. In Pleistocene time the cirques below the summit ridge hosted small glaciers which supplied debris to fans along the side of the south-draining Lavant Valley. Near Gemmersdorf the lower segments of these relict Pleistocene debris cones have been dissected by creeks which, under normal conditions, are harmless. In contrast, the upper segments of the cones are not incised and are mantled by convex lobes of coarse debris derived by erosion from 'debris bowls' below the Speikkogel.

On 7 June 1660, heavy rains mobilized large masses of debris in the basin above Gemmersdorf. Blockage of the channel probably set the stage for an outburst of a debris flow which demolished several buildings and killed 29 people in Gemmersdorf. A remarkable contemporary pencil sketch of the disaster, accompanied by descriptive notes (Fig. 125), has been found in a regional archive by Fresacher (1965). For centuries after the catastrophe local legends attributed the sudden burst of debris to an infuriated dragon ('Lindwurm') which was thought to have inhabited the uplands.

On 7 September 1916, a similar debris flow overwhelmed the hamlets of Oberpichling and Paierdorf. The disaster occurred after three weeks of steady rain and thunderstorms. At the head of the Kreuzerbach basin a slope failure, involving possibly as much as $0.5 \times 10^6 m^3$ of surficial material blocked runoff from several tributary torrents. When the debris barrier failed a rapid flow of rocks, trees and mud descended the torrent and spread across the cultivated land of the upper cone. At least 22 people lost their lives (Wittmann, 1952).

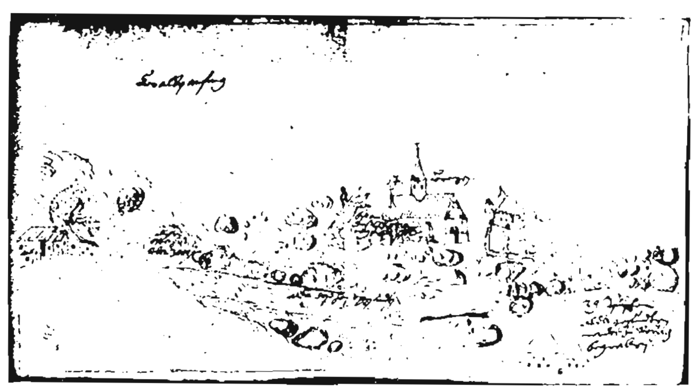


Figure 125: Contemporary illustration of the destructive bouldery debris flow which overwhelmed the village of Gemmersdorf in 1660 (from Fresacher, 1965).

In recent years, the composite cones at the foot of the Speikkogel Massif have been invaded by residential development. In many of these areas blocks of past debris flows protrude through cultivated meadows and fields. The creek beds themselves bear little evidence of the debris potential in the uplands.

Lake Traun (A45)

Location: Salzkammergut, Oberösterreich, Austria (I1)
Date(s): June 1664, (1737, 27 to 31 July 1897, 10 to 13 September 1899, 18 July 1910, 23 July 1955)

The Salzkammergut is one of the most scenic sections of the eastern front ranges of the Alps. Sparkling lakes fill drowned valleys and morainal ampitheatres framed by rugged carbonate ranges of the Austroalpine cover complex and rolling foothills underlain by recessive formations of the Helvetic and Molasse zones. The lakes are rimmed by benches of late Pleistocene surficial deposits. Torrents and rivers flowing into these lakes generally follow west-trending bedrock structures and have built substantial fan deltas.

Throughout most of medieval time the region was a centre of salt mining, refining and trading. Extensive clearcuts in the upland basins provided timber and charcoal necessary for the mines and salterns. Storms and persistent seasonal rainfall resulting from the blockage of moist air masses against the front ranges of the Alps ('Schnürlregen' = rains on a string) have accentuated the erosional scars created by careless logging practices of the past.

Lake Traun (440 m) in the eastern Salzkammergut (Fig. 126) is flanked by precipitous carbonate slopes on the south and by gentle shale-sandstone terrain in the north. The contact between the two rock types is a south-dipping thrust fault. The Gschliefgraben (Gschlief = slide) on the north side of this fault zone is an elongate depression extending from the eastern shore of the lake to the foot of the calcareous cliffs of the Traunstein Massif (1691 m). It follows the structural trend of the recessive shales or marls and its tributary ravines branch into a headland (900 m) composed of well-cemented flat-lying breccia derived from the adjacent limestone cliff. Most of the Gschliefgraben is filled with a slowly moving lobe of debris totalling 2 to 3 $\times$ $10^6 m^3$ in volume. This composite earth-debris flow originates from bowl-shaped slumps along a head scarp where blocks of carbonate breccia become immersed in a matrix of remolded shale. From several tributary branches the flow develops into a central tongue characterized by sharply defined lateral levees and having a high water content. Towards the lakeshore the tongue develops into a bulging lobe whose front falls at about 7° towards the water and is dissected by two torrents which discharge into the lake. Zones of discrete differential shear are delimited by numerous pressure ridges which suggest that the shore front itself partakes in the movement of the superincumbent debris. In the upper section of the flow average rates of movement of approximately 6 m/year have been documented from long term observations (Baumgartner, 1980).

In June 1664, following three years of heavy rains and floods (1661 to 1663), the Gschliefgraben debris lobe apparently accelerated and pushed an entire country estate from the adjacent delta cone into the lake. In 1737 rapid debris flows, mobilized by the torrents along the slow-moving lobe, overloaded the cone and caused a failure of the shore front including buildings located on it. A large section of the debris front and underlying lacustrine silts slumped into the lake leaving behind a concave scar. Almost two centuries later, during rainstorms between 10 and 13 September 1899, on 18 July 1910, and on 23 July 1955 pulses of debris again damaged lakeshore properties.

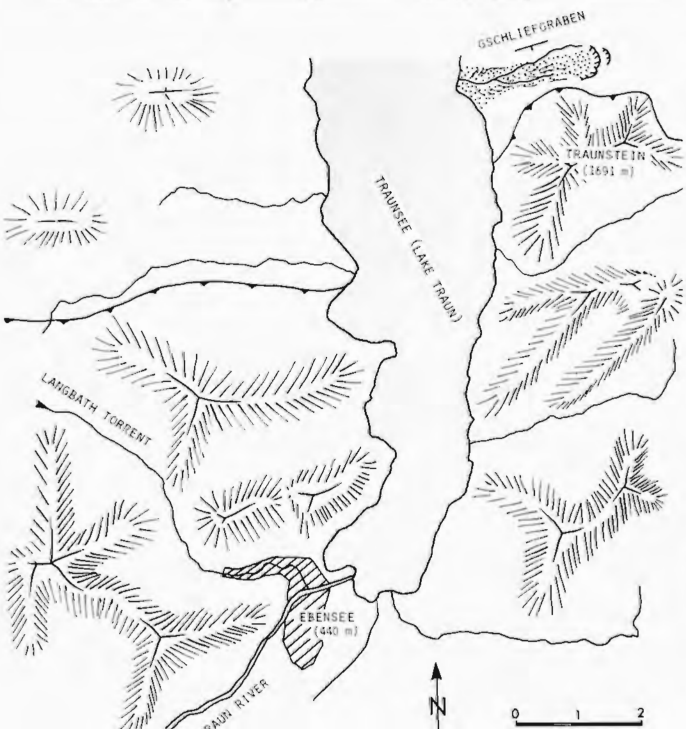


Figure 126: Index map of Lake Traun showing the location of the Gschliefgraben debris lobe and the Langbath basin in the carbonate terrain south of the boundary thrust fault (barbs on the upthrown side).

Figure 127: View across the toe of the Gschliefgraben debris lobe; note embayment marking the head scarp of the historical subaqueous slump and densely reforested zone above vacation bungalows. (GSC 204167-B)

The attractive shore front along the periphery of the Gschliefgraben cone hosts a variety of permanent dwellings, vacation bungalows, and a major hotel. Recently (1975) much of the area was zoned as being too dangerous for any kind of new residential development. Check dams have been installed along the lower reaches of the two torrents draining the debris lobe to avoid undercutting of unstable embankments. A dense protective forest of conifers has been planted above the threatened buildings (Fig. 127). In the uplands only marginally effective drainage ditches have been excavated and movement of the main contributing branches of the slow earth-debris flow is being monitored by regular surveys.

An entirely different problem once threatened the community of Ebensee on the composite delta of the Langbath Torrent and Traun River. The Langbath Torrent drains a sizeable carbonate basin; its wide braided channel is flanked by benches of relict Pleistocene colluvium. In the heyday of medieval salt mining the torrent was used extensively to drift logs from the uplands to the salterns of Ebensee. For this purpose strong timber linings were installed along the channel embankments as early as the 14th century. When timber drifting was abandoned during the 19th century unstable colluvium again slumped freely into the torrent. Between 27 and 31 July 1897, a major rainstorm dumped a total of 380 mm of rain onto the Langbath basin. Debris avalanches along the embankments blocked the swollen torrent creating massive flows of uprooted trees and carbonate rubble; in Ebensee the debris flows covered buildings and shifted the channel of the Traun River. Restoration work in the area had barely begun when, between 10 and 13 September 1899, another storm brough 500 mm of rain to the basin. This time debris flows buried the houses along the channel of the torrent to the roof tops (Wühl, 1980).

An energetic restoration campaign, involving at times as many as 1000 workers, was launched with the aid of a personal grant by emperor Franz Josef who happened to own an exlusive hunting preserve in the Langbath basin. This work, supplemented in recent years by diligent forest practices and modern control structures, has largely neutralized the potentially destructive effect of severe rainstorms.

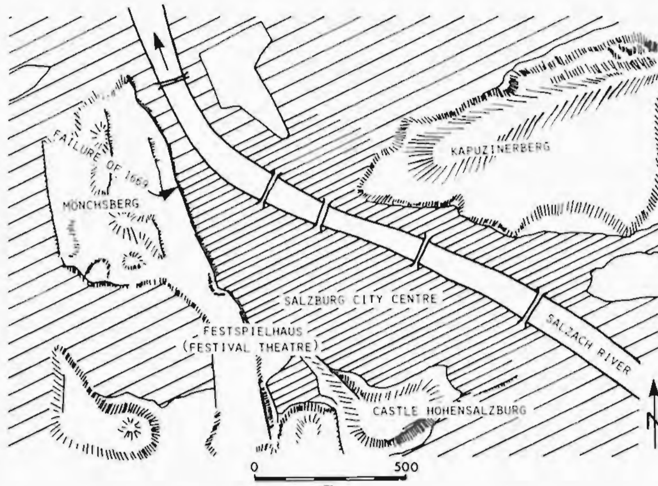


Figure 128: Sketch map of the city centre of Salzburg; north-northwesterly trending fractures control the orientation of the Mönchsberg wall which is composed of Pleistocene conglomerate.

Figure 129: Southward view along the vertical Mönchsberg wall in the city centre of Salzburg. St. Marcus church in the foreground; festival theatre and castle in the background; note apartment buildings at the base of the cliff. (GSC 204167-C)

Salzburg (A46)

Location: Salzburg, Austria (11)
Date(s): 15 July 1669 (also 1493, 1614, 1665)

The capital city of Salzburg (420 m) spreads along the banks of the north-flowing Salzach River (Fig. 128) which here crosses the carbonate mountains of the Austroalpine front ranges and flows through more subdued terrain underlain by marl and sandstone of the Alpine foothills. In Salzburg several steep-sided bedrock ridges of carbonate, flanked by terraces of flat-lying Pleistocene conglomerate ('Nagelfluh'), project up to 200 m above the streets of the city. The orientation of vertical cliffs along the Mönchsberg, Kapuzinerberg, and the Castle of Hohensalzburg are controlled mainly by north-northwesterly trending fractures. During the early growth of the city lack of space along the river bank forced rows of residential buildings against the east-facing Mönchsberg conglomerate cliff.

In 1493, 1614 and 1665 small rockfalls from the Mönchsberg wall crashed into the dwellings below and claimed the lives of several people. However, the most serious failure along the cliff occurred at 2:00 a.m. on 15 July 1669, when a slab of conglomerate with a volume of approximately 10 000 m^3 rotated outward from the crown of the conglomerate terrace, toppled onto a row of houses along the base of the Mönchsberg, and killed at least 220 people. It also demolished the old St. Marcus church. When a large crowd converged onto the scene of the accident, another smaller slab broke away above the street and killed another 30 persons (Strele, 1936, p. 128-130).

Despite the undiminished threat of rockfalls from the Mönchsberg the church and the houses were rebuilt in the same place soon after 1669 (Fig. 129). However, since the catastrophe the rock wall has been regularly inspected and scaled by teams of 'Steinputzer' (= stone cleaners). This practice, still performed today, has become an annual Salzburg folk tradition. Even the focus of Salzburg's modern cultural life, the world famous Festspielhaus (festival theatre), has been placed against the Nagelfluh wall. Numerous rock anchors, monitoring devices, and other works have substantially reduced the danger of rockfalls. In recent years a huge underground cavern has been excavated inside the Mönchsberg to alleviate the city's chronic parking problem!

Other steep bedrock slopes within the city, including the precipitous cliff below the castle have failed in rockfalls or have threatened to fail in the past. It is thus not surprising that considerable pioneering work on rock mechanics was done in this city (Müller, 1963)

Ötz Valley Ice Floods (A47)

Location: Ötztal Alps, Tirol and Südtirol (Alto Adige), Austria and Italy (F2)
Date(s): 17 July 1678 (also 25 July 1600, 15 June 1680, 30 June 1717, 18 June 1725, 28 July 1737, 17 July 1772, 14 June 1845, summer 1848, 1 August 1862, 5 August 1874, and 7 August 1890)

The impressive mountain ranges of the Ötztal Alps, flanked by the wide valley of the Inn and Etsch (Adige) rivers, dominate the border region between western Austria and northern Italy (Fig. 130). Peaks and high ridges, jutting above icefields and valley glaciers reach elevations of more than 3000 m, and are composed of high grade metamorphic rocks of the Austroalpine basement complex. Runoff from the glaciated upland basin flows in major torrents draining either to the Inn River on the north or to the Etsch River on the south.

During the Little Ice Age some of the glaciers advanced over steep bedrock lips into narrow valleys and thus impounded glacial lakes. The most significant of these ice-dammed lakes were the Rofen Eissee (Richter, 1892), those along the margin of the Gurgl Glacier, and the proglacial and englacial lakes of the Wütenkar Glacier and the Stubai, Kauner, Ridnaun, Matscher, and Langtauferer valleys.

The events related to the Rofen Eissee (= ice lake) have been chronicled by Walcher (1773), Stotter (1846), Richter (1892), Finsterwalder (1897), and Hoinkes (1969). The Rofen Eissee developed whenever the snout of the Vernagtferner (Ferner = glacier) advanced rapidly to the bottom of the Ventertal (Vent Valley), blocking the east-flowing Rofenache Torrent, a tributary of the Ötztaler Ache. On 25 July 1600, the first documented burst of the impounded Rofen Eissee through the snout of the Vernagtferner resulted in a debris flood which damaged large sections of the upper Ötz Valley. One year later, on 12 July 1601, an even larger ice lake overflowed; the population was on alert, anticipating another violent burst which, however, did not materialize. The worst ice flood occurred on 17 July 1678, less than a year after the glacier again had reached the bottom of the valley. The lake first overflowed the ice dam harmlessly. Then, with abruptly rising temperatures in late June, the crevassed mass of ice could no longer withstand the water pressure exerted by the lake. On 17 July, the dam failed during a cloudburst and a deluge of water, ice, and floating debris swept over fields and roadworks of the upper Ötztal. Numerous buildings were knocked off their foundations. Damage was compounded by simultaneous debris flows that blocked the Ötztaler Ache near Längenfeld thus converting the valley bottom into a lake. Fortunately, the people of the Ötz Valley had been forewarned by the roaring sound of the approaching flood and only two lives were lost. In the bitter aftermath of the catastrophe a travelling journeyman was burned at the stake in Meran, after he had been accused and found guilty of conspiring with the Devil to bring this disaster onto the inhabitants of the valley!

This drastic action, however, did little to prevent another blockage of the Rofenache Torrent soon thereafter, and on 15 June 1680, another overflow from the ice lake took out bridges and damaged buldings. At the time J. Kuen, a community leader in Längenfeld hiked to the ice barrier and with twelve helpers cut a drainage ditch across the regenerating ice dam. This channel was kept open by natural erosion and eventually the lake disappeared.

A fourth surge of the Vernagtferner closed the gorge between 1770 and 1774, but this time the Eissee emptied without doing much damage. However, apprehension rose among the people of the valley, and a government commission, led by J. Walcher, a professor of mechanics at the University of Vienna, visited the Rofen Eissee. In a publication Walcher (1773) discussed the mechanics of glacial advances, formation of the ice lake, its drainage, and commented on possible remedial measures. He favoured a consensus arrived at in consultation with the local population:

> ' . . . that the torrent channels be diligently cleared; that the objects that might be carried away by the waters be removed; that low bridges and crossings be raised; that the channels be maintained at a reasonable depth and possibly be straightened; that, finally, strong "Archen" (= stone levees) be built to protect low-lying properties . . .' (Walcher, 1773, p. 39).

Walcher indicated that the people would have liked to see the construction of a flood retention dam below the ice lake, but that the financial burdens were considered too great.

A final advance of the Vernagtferner in the 1840s culminated in an ice-debris flood on 14 June 1845. The burst, monitored by a scientific team from the University of In-

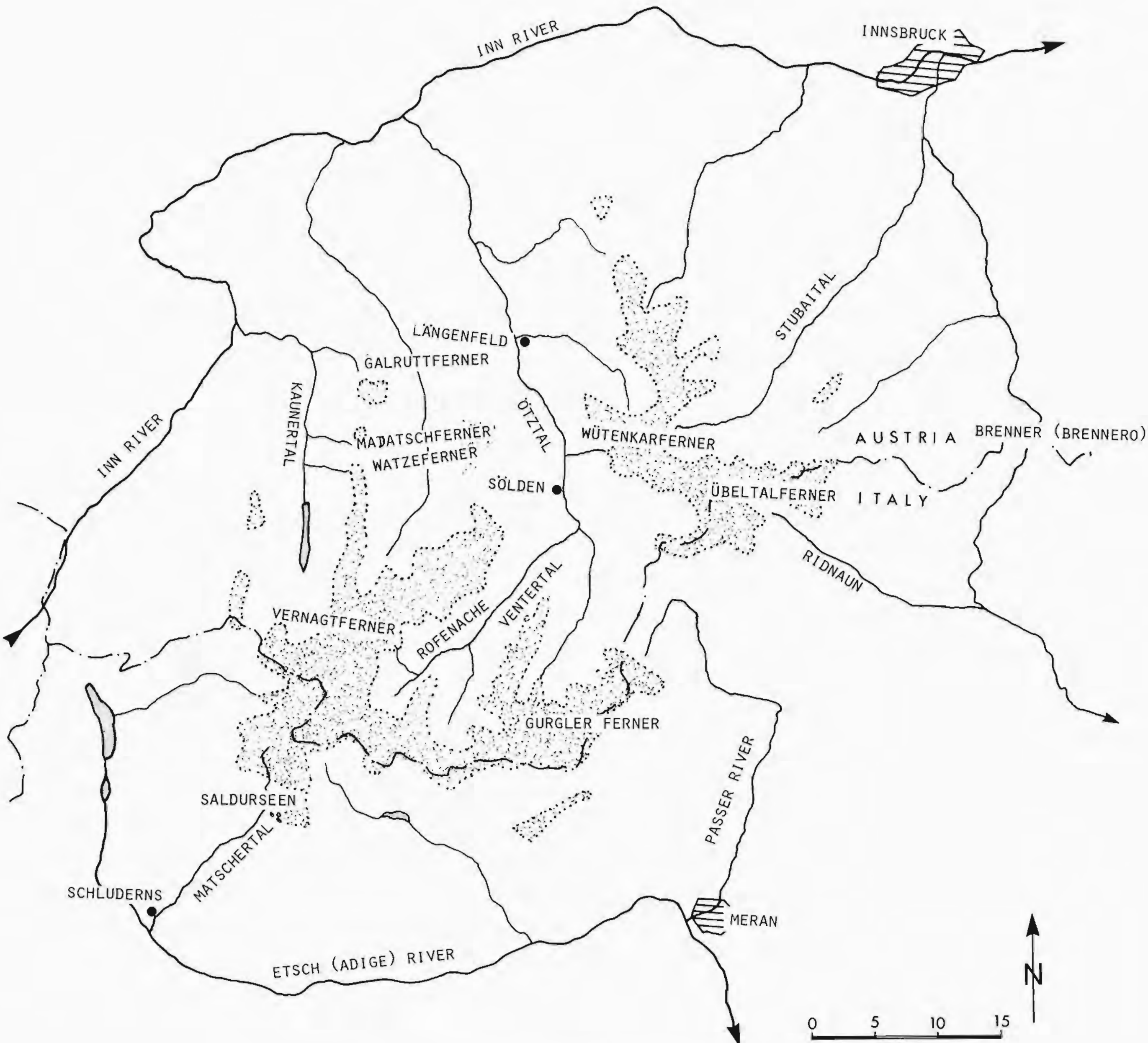


Figure 130: Index map of the Ötztal Alps region of the eastern Alps.

nsbruck (Fig 131), involved a volume of approximately 1 × $10^6 m^3$ of water creating a frontal wave 10 m high at a point 5 km below the dam. The ice flood took out numerous bridges, damaged buildings, and devastated fields in the upper Ötztal (Stotter, 1846). In the century that followed the snout of the Vernagtferner withdrew into its cirque basin.

The Gurgl Eissee, a shallow ice-margin lake at the confluence of the Gurgler Ferner and Langtal Ferner, experienced several historical midsummer bursts, which, however, resulted only in minor damage to human works in the upper Ötztal. The largest of these floods occurred on 30 June 1717, and in the summer of 1848.

On 18 June 1725, after a period of warm and rainy weather, a small body of water near the front of the Wütenkarferner near Sölden, Ötz Valley, suddenly drained into the straight ravine below the glacier, mobilized morainal debris and generated a massive debris flow which crossed the steep debris cone at the bottom of the valley. The debris lobe pushed the swollen Ötztaler Ache against a rock wall along the west bank of the river and blocked its flow until a lake,

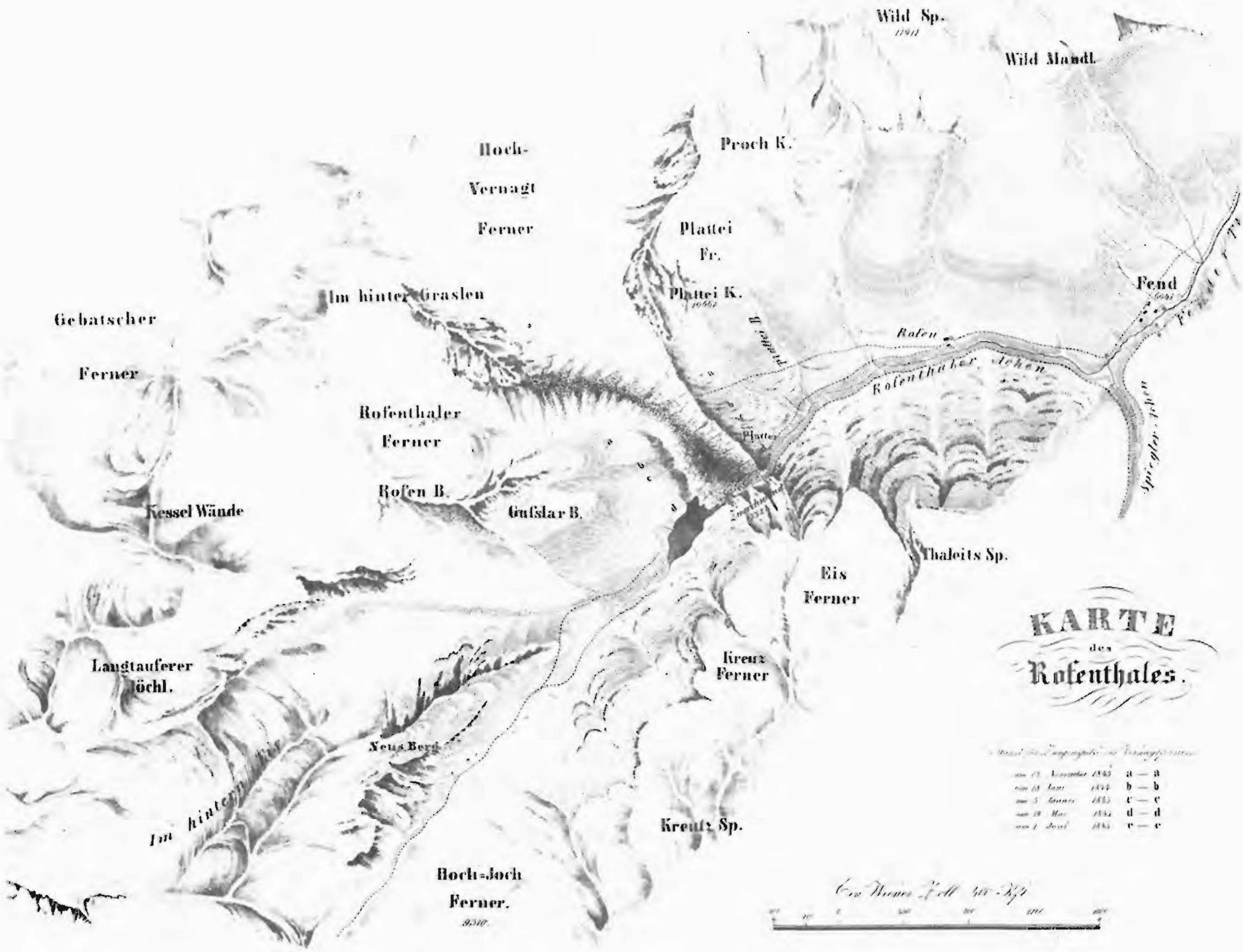


Figure 131: Map of the Rofen ice lake produced by blockage of the Rofenache Torrent (here called Rofenthaler Achen) by the Vernagtferner (after Stotter, 1846). The position of the glacier snout at various times during its advance between November 1843 and June 1845 is shown.

600 m long, had formed upstream. The debris barrier failed, triggering a flood that picked up more debris farther downstream. The raging river demolished most bridges and transported rubble that covered large tracts of cultivated land (Walcher, 1773, p. 60-61).

On 17 July 1772, following a period of very hot weather, an ice avalanche cascaded into a small glacial lake in the upper Stubaital, and the overtopping waters advanced as a debris flood along this valley, causing considerable damage (Schaubach, 1865). During the Little Ice Age the Ridnaun Valley also experienced repeated ice-debris floods, resulting from the blockage of the Übeltalferner snout (Übeltal = evil valley). Sporadic and rapid subglacial drainage of the ice lake prompted the construction of a stone retention dam in the valley below the glacier as early as 1745. The early structure, the Agglsbodensperre, was substantially improved in 1880, but by this time the glacier had begun to retreat into its firn basin. This stone-masonry dam provided for temporary storage of floodwaters and their gradual release through openings in the crown of the dam (Richter, 1889b).

Ice floods also threatened the Matscher Valley between 1737 and 1901. Most of the floods originated in small proglacial lakes, the Saldur lakes near the head of the valley. The most serious debris flood of probable glacial origin occurred on 28 July 1737, during a period of fine weather, when debris carried out by a swollen torrent blocked the gorge above the town of Schluderns; the subsequent massive flow of debris onto the community killed eleven people and covered buildings up to several metres (Stacul, 1979, p. 51).

Figure 132: a) Steep debris cone of a glacial torrent (locality Am See) in the Kaunertal (GSC 204167-D). b) Warning sign along the recently active torrent channel of the cone shown above (GSC 204167-E).

Ice-debris floods from small bodies of water have been experienced in the Kaunertal, a valley flanked by bedrock walls and steep debris cones. On 1 August 1862, an englacial (?) lake burst forth from the Watzeferner and devastated a hamlet. On 5 August 1874, a proglacial lake that had formed behind the looping Little Ice Age terminal moraine of the Madatschferner broke through its unstable morainal dam. A massive debris flow, generated from eroded moraine, cascaded into the valley destroying the valuable fishstocks of the Kaunertal. Similarly, on 7 August 1890, a piece of ice, 2000 m^3 in volume, collapsed ('calved') from the snout of the retreating Galruttferner into a proglacial moraine-dammed lake; a wave overlapped the moraine and developed into a debris flow with a volume of $0.15 \times 10^6 m^3$, which deluged a section of the Kaunertal (Koch, 1875, 1892).

Downwasting and retreat of the Ötztal glaciers, which continued until the 1960s, greatly reduced the hazard of ice-debris floods in the region. Several of the high basins are now part of complex hydroelectric reservoir-and-diversion schemes and some of the glaciers have been developed into year-round ski areas; as a result most of the communities have grown and prospered. The readvance of the glaciers in recent years may initiate a new cycle of more frequent ice floods (Fig. 132).

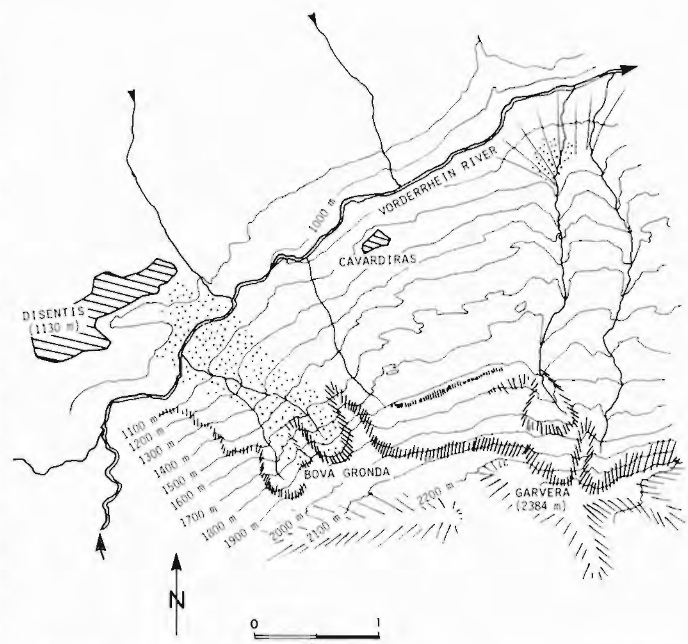


Figure 133: Sketch map of the north-facing sagging slope of the Garvera Massif near Disentis and slide mass of 1683 (stippled) below the Bova Gronda scar.

Disentis-Muster (A48)

Location: Vorderrhein Valley (Rein Anteriur), Graubünden (Grisons), Switzerland (D2)
Date(s): 29 June 1683

In the area of Disentis (Muster) the Vorderrhein Valley (1000 m) follows a fracture zone that parallels south-dipping gneisses of the Helvetic basement complex. On the south side of the river the Garvera Massif (2384 m) rises in a scarp face which is the zone of detachment for a gigantic sagging slope involving approximately $1000 \times 10^6 m^3$ of gneissic bedrock (Fig. 133). In historical times the sagging slope and the scarp face of the Garvera Massif have been the locus of repeated mass movements.

The most dramatic slope failure occurred on 29 June 1683, when a rock slab of 10 to $20 \times 10^6 m^3$ failed along the westernmost part of the Garvera ridge (Bova Gronda). From

Figure 134: View of the Bova Gronda scar and slide deposits of 1683; note recent rock-debris avalanche chute below Bova Gronda, farm buildings on the slide mass, and the serrated skyline of the sagging Garvera slope. (GSC 204167-F)

an elevation of 1900 m the intensely fractured rock mass streamed to the bottom of the valley (Fig. 134), its front climbed 120 m up to the terrace of Disentis (1130 m) and temporarily halted the waters of the Vorderrhein River. The rock avalanche killed 22 people (Heim, 1921, p. 948).

Today, the settlements of the valley are perched on terraces between torrent cones on the north side of the river. Several small hamlets also dot the surface of the sagging Garvera slope. A large campground has been opened recently along the bouldery river bench just below Disentis.

La Valle Agordina (A49)

Location: Agordo, Veneto, Italy (H2)
Date(s): April 1701 (also 18 August (?) 1748, 23 to 27 April 1888)

The twin community of La Valle Agordina (850 m) spreads over the head of a composite relict debris fan, which rises on the northeast side of the Cordevole Valley and is incised by two torrents, the Torrente Missiaga on the north and the Torrente Bordina on the south (Fig. 135). Tributary branches of both torrents originate below the rugged carbonate walls of Monte Castello (2499 m). Both torrents parallel the Linea della Sugana, a zone of complex, southeast-directed thrust faults cutting bedrock of the South Alpine cover complex. The carbonate cliffs of Monte Castello are composed of flat-lying massive Dolomia Principale Formation and rest on recessive redbed-shale units. Subsidence of recessive shales favours sporadic collapse of the superincumbent carbonate walls which form the spectacular skyline of the region.

In April 1701, the west-facing wall of the southern spur of Monte Castello failed along a rupture surface dipping approximately 45° to the west. The failure involved a total rock mass of 5 to 10 × $10^6 m^3$. The toe of the rockslide rotated outward over the possible snow-covered and water-saturated terrace of Malga Moschesin (1800 m) and changed into a violent debris stream down the Missiaga Torrent. A wall of rock and mud crashed into the village of La Valle

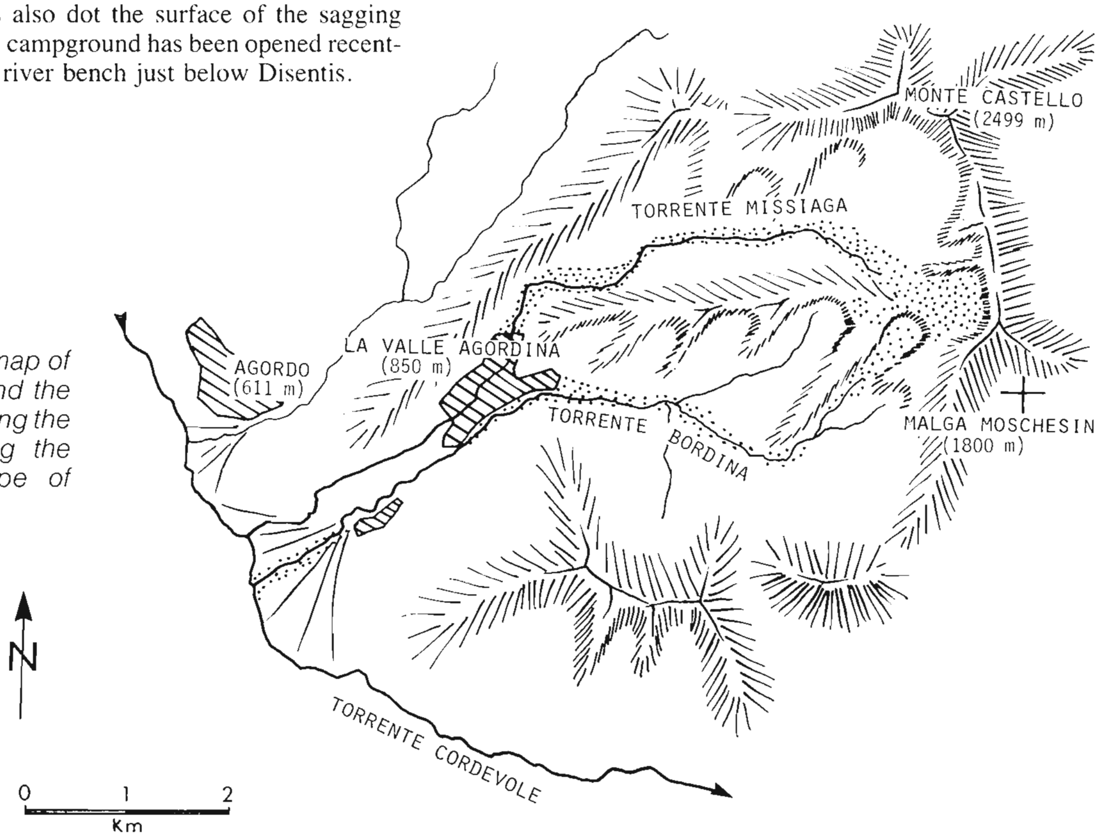


Figure 135: Sketch map of La Valle Agordina and the debris flow tracks along the two torrents draining the unstable scarp slope of Monte Castello.

Agordina, erasing several buildings and leaving 48 inhabitants dead (Montandon, 1933, p. 304). The track of the debris stream can still be followed from the bedrock failure to the undulating meadows above La Valle Agordina, which gradually are being built over (Fig. 136).

The 1701 failure had another effect: the slide mass which overloaded the saturated shale terrace below Monte Castello gave rise to a huge incipient slump, involving most of a small tributary basin north of the Torrente Bordina. In 1748 part of ths slump mass was mobilized into debris flows which damaged the hamlet of Connagia. More than a century later, during the rainy decade between 1880 to 1890 the upland slopes began to move once more. A series of debris flows issuing from the toe of the slump mass were documented by an anonymous member (A.S.) of the Club Alpino Italiano in the Revista Mensile of the club (Volume for 1888, p. 138-140). According to this account the inhabitants of Connagia were startled by bellowing and crashing sounds on the morning of 23 April 1888. A few montanari immediately set off to investigate and soon discovered the source of the noise: a bulging and splitting lobe of rockslide deposits, surficial debris, trees, and water with a total volume of several million cubic metres had begun to force its way through a funnel-shaped northern tributary of the Bordina Torrent and was slowly approaching Connagia. The inhabitants of the village had enough time to remove their belongings from some thirty buildings before they were crushed on the evening of 26 April 1888. The front of the bouldery debris lobe had advanced at an average rate of 50 to 60 m/h.

Today the large concave head scarp and the rockslide deposits are still clearly visible west of Malga Moschesin. During the catastrophic rainstorm of 4 November 1966, the Bordina Torrent, like many others in the Cordevole region, again carried a heavy load of debris. Since then the threat of mass movements has been fully recognized. A heavy stone-masonry dyke and a concrete wall now separate the bouldery track of the Bordina Torrent from fields and newly constructed vacation homes in Connagia.

Figure 136: View down the track of the 1701 debris flow along Torrente Missiaga; La Valle Agordina in the background. Slight undulations in the meadows indicate positions of large carbonate blocks embedded in a shaly matrix (GSC 204167-G).

Diablerets (A50)

Location: Valais (Wallis), Switzerland (C2)
Date(s): 23 June 1714 (also 23 July 1749)

The mountain ridge of Les Diablerets (3209 m) follows a southwest-trending axial culmination in the Helvetic fold-and-thrust belt of western Switzerland (Fig. 137). The summit plateau of the Diablerets is composed of massive Urgonian limestone, capped by an ice field, the Glacier de Diablerets. Below the summit cliffs the southeastern scarp face of the mountain is underlain by sandy calcareous slate. Although the deformed formations generally dip gently to the north, cleavage, fracture and bedding planes dip southeast, parallel to the face of the mountainside.

Beginning in early June 1714 repeated rockfalls and ominous rumbling from Les Diablerets attracted the attention of the people working on the alpine pastures of Derborence (1400 m) below the cliffs. At this time meltwater from the glacier probably was percolating into opening bedrock cracks. Finally, on 23 June 1714, several million cubic metres of rock (and ice) became detached from the face of the mountain and cascaded over the gently inclined terrace below. The rock stream overwhelmed alpine pastures, annihilated 55 buildings and killed 15 people. A freakish circumstance led to the incredible survival of one of the men: a huge block crashed into a chalet, and stopped where it landed, thus protecting the rest of the frail structure from the streaming debris; trapped alive inside the cheese room of this chalet and cut off from the outside by a thick blanket of rubble, but supplied by the cheese he had been preparing and a trickle of water, he set about digging himself out. Eventually, after three months of strenuous and injuring work, the emaciated figure emerged from the chaotic surface of the slide mass looking like a ghost from the netherworlds. His story was later retold in the powerful novel 'Derborence' by C.F. Ramuz.

On 23 July 1749, a recurrent failure of about 30 $\times$ $10^6 m^3$ set off another avalanche of rock and ice which covered the slide deposits of 1714. This time precursory rockfalls and crashing noises were heeded by the mountain people, for most of them had left the uplands by the time the wedge of rock and ice plummeted off the bedrock bench below the main cliff, then glanced off the vertical wall of L'Ecorcha, and streamed 3 km down the gorge of the Lizerne Torrent. Nevertheless, five people perished in the rock stream. The blocky lobe impounded the Derborence Torrent into the Lac de Derborence. The total mass of the two Diablerets debris streams ('Liapey') has been estimated at approximately 50 $\times$ $10^6 m^3$ (Heim, 1932, p. 130-133, p. 187 and 1921, p. 460-461; Becker, 1883, p. 310-316).

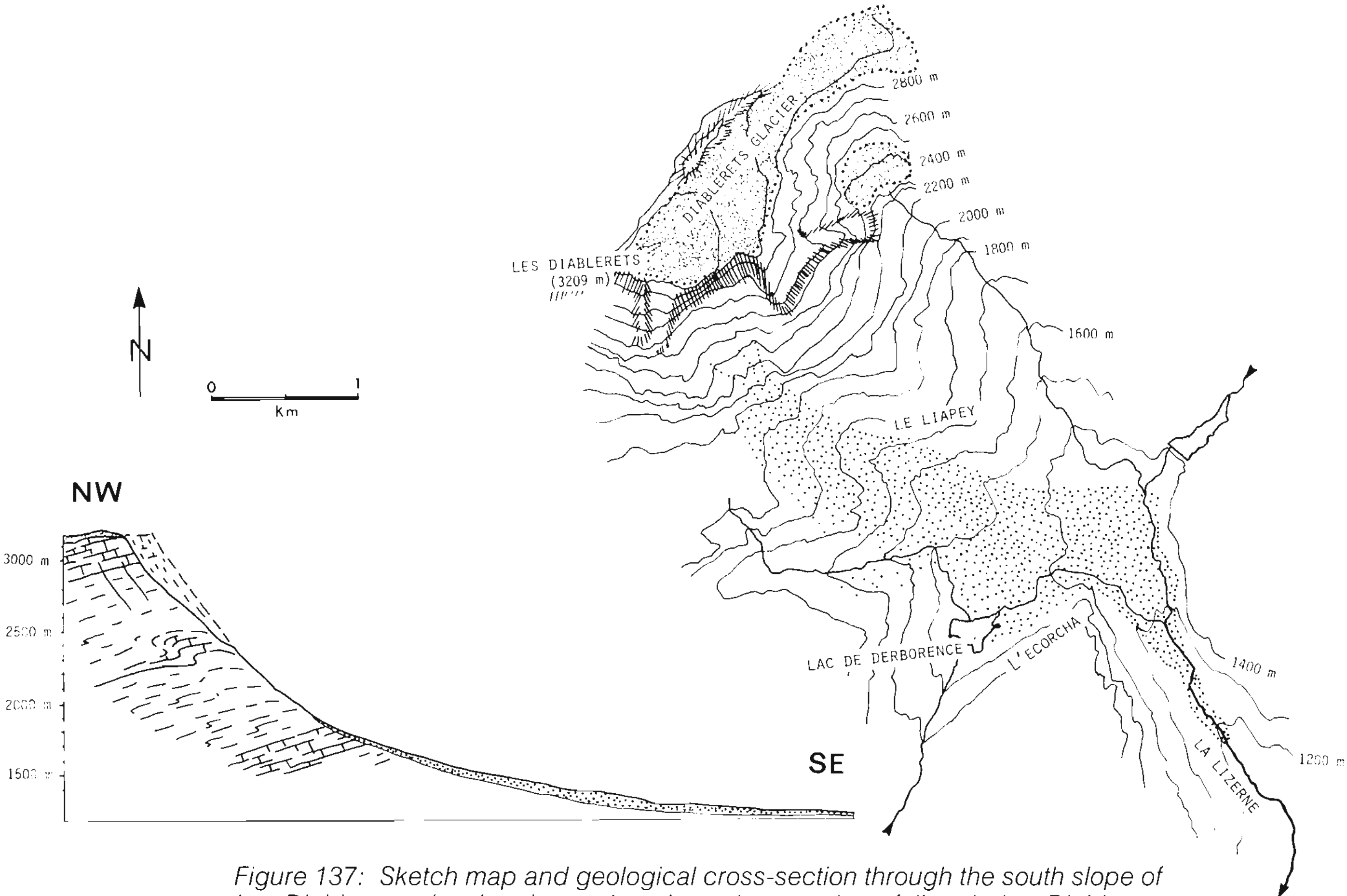


Figure 137: Sketch map and geological cross-section through the south slope of Les Diablerets, showing the wedge-shaped scarp slope failure below Diablerets Glacier.

Since 1749 several smaller rockfalls have built a conspicuous talus cone at the foot of the Diablerets wall (Fig. 138). The largest of these occurred in 1881 and 1944.

Today the basin of the Derborence is accessible via a spectacular road that winds its way along steep cliffs and across the Liapey to Lac de Derborence. Several vacation chalets, a camp ground, and a restaurant dot the margin of the blocky surface which is now partly covered by a struggling pine forest. During the summer months the area is visited by many people.

Figure 138: Northwestward view of the detachment zone of the Diablerets rock avalanches and rockfalls; note chalets on the promontory east of the blocky deposits of Le Liapey. (GSC 204167-H)

Val Ferret — Val Veni (A51)

Location: Valle d'Aosta, Italy (C3)
Date(s): 12 September 1717 (also 15 August 1728, 14 and 19 November 1920)

Val Ferret and Val Veni form a remarkably linear trough along the Italian side of the Mont Blanc Massif (4807 m). The valleys parallel the northeasterly trend of the Helvetic basement crowned by the imposing ice-covered granitic spires of Mont Blanc (Fig. 139). About twenty glaciers cling to cirques and bedrock ledges along the southeast-facing valley wall. During the Little Ice Age (1550-1850) most of these glaciers reached the valley bottom (1300-1800 m). Moraines fronting these ice tongues receive abundant debris from massive rockfalls, ice-rock avalanches, and debris flows (Porter and Orombelli, 1981).

During the early 18th century the Triolet Glacier played a part in catastrophic mass movements in the upper Val Ferret. Published accounts differ in emphasis and detail, but it is probable that two major catastrophes occurred (compare the accounts of Rabot, 1905; Sacco, 1918; and Montandon, 1933, p. 306). On 12 September 1717, a mass of rock, ice and water, possibly set off by the burst of a small englacial water pocket, developed into a bouldery debris flow that overwhelmed pastures below the glacier, killing seven people and much livestock. On 15 August 1728, about 10 to 20 $\times$ $10^6 m^3$ of granitic debris, having broken away from a bedrock spur above the Triolet firn basin (Aiguille de l'Éboulement), developed into a rock-ice avalanche that moved at great speed 2 km beyond the glacier terminus where it annihilated a hamlet.

On 14 and 19 November 1920, a section of the spectacular Peuterey ridge collapsed onto the Brenva Glacier whose snout then was near the bottom of Van Veni. By scouring ice, snow, and morainal debris the rock avalanche lobe swelled to a volume of 4.5 $\times$ $10^6 m^3$; it blocked the flow of the Doire de Veni and created a small lake (Valbusa, 1921). The thick blanket of granitic debris on the lower Brenva Glacier protected the ice from solar radiation and thus triggered a spectacular advance of the glacier snout that lasted until 1940, while all other glaciers in the region experienced dramatic retreat. Between 1940 and the mid-1960s downwasting and retreat of the glacier left behind an imposing complex of moraines (Fig 140). More recently the snout of the Brenva Glacier has advanced almost to the entrance of the Mont Blanc tunnel at Entrèves.

It is possible that during the initial glacier advances of the Little Ice Age an ice avalanche of major proportions occurred in the vicinity of the Brenva Glacier. In 1842, J.D. Forbes was told by one of the locals that

''..on St. Margaret's day, the 15th of July, no one knows in what year, the inhabitants of the village of St. Jean de Pertus, which was then overhung by the Glacier de la Brenva, instead of keeping the fête, pursued their worldly occupations: – the hay is dry, they said: the weather is fine: let us secure it. But the sacrilege was soon punished. Next day the glacier descended in a moment, and swallowed up the village with its inhabitants. . .' (Forbes, 1845, p. 206-207).

Today the Val Veni and Val Ferret are frequented by numerous visitors during the summer tourist season. New residential and hotel buildings have sprung up in the lower parts of the valleys. Although they are generally protected from snow avalanches, much of the hazard from rock and ice falls has been found difficult or uneconomic to eliminate.

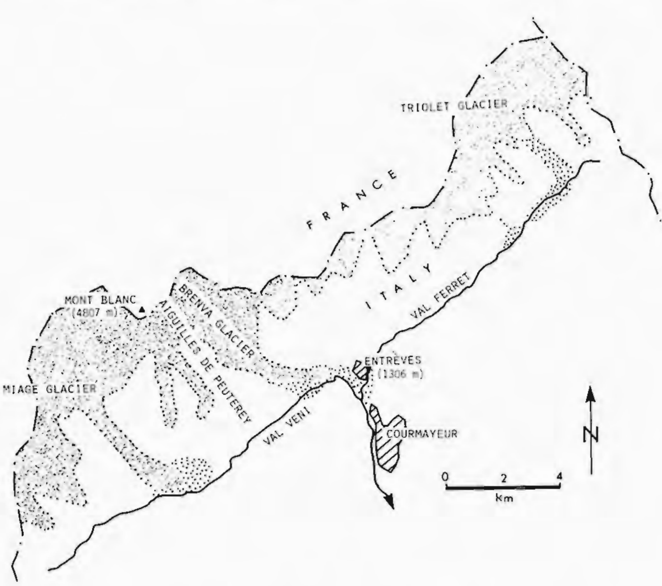


Figure 139: Index map of the hanging glaciers of Val Ferret and Val Veni along the southeastern approaches of the Mont Blanc Massif.

Figure 140: View from the debris-strewn lower part of the Brenva Glacier towards the Aiguilles de Peuterey. (GSC 204167-I)

Schlanders — Silandro (A52)

Location: Vintschgau (Val Venosta), Südtirol (Alto Adige), Italy (F2)
Date(s): 29 May 1731

Schlanders (720 m) is the prosperous commercial and tourist centre of the Vintschgau (Val Venosta), a fertile valley following an east-trending lineament within the Austroalpine basement complex (Fig. 141). The Vintschgau is drained by the east-flowing Etsch (Adige) River which receives tributary torrents from glaciated catchment basins rimmed by peaks with elevations above 3000 m.

Schlanders, like other communities in the valley, developed near the head of a debris fan on the north side of the flood-prone river. The Schladraun Torrent flows from an elongate upland basin through a narrow bedrock gorge onto a gentle fan. The shape of the basin is controlled by steep north-trending fractures cutting steeply south-dipping mica schist. Unstable dip slopes and relict colluvial deposits supply most of the bedload of the torrent.

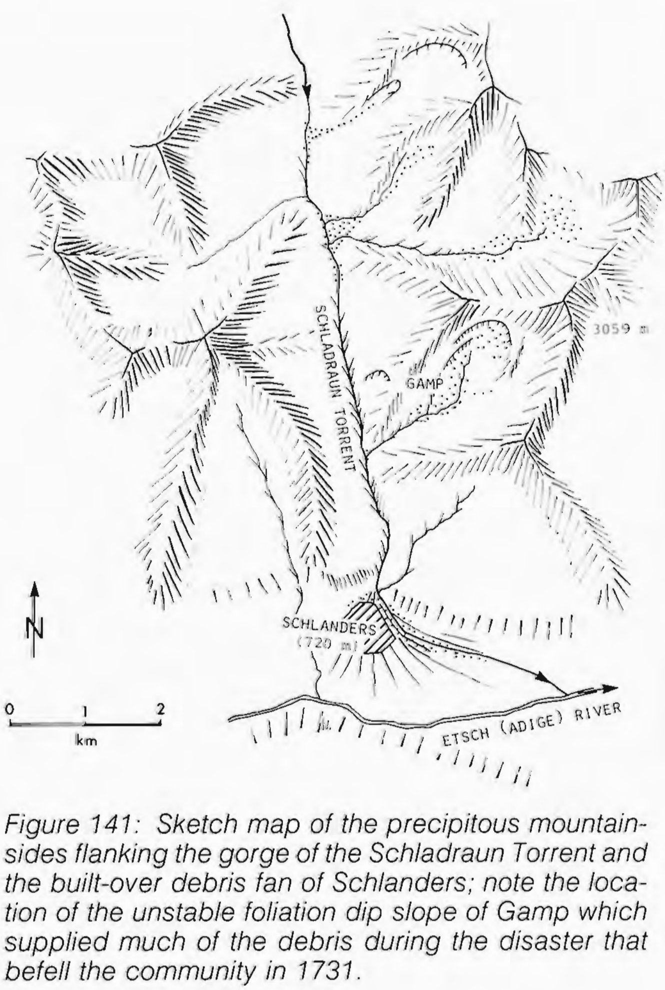


Figure 141: Sketch map of the precipitous mountainsides flanking the gorge of the Schladraun Torrent and the built-over debris fan of Schlanders; note the location of the unstable foliation dip slope of Gamp which supplied much of the debris during the disaster that befell the community in 1731.

Occasional debris flows were experienced during the early history of Schlanders and in 1461 'water walls' were first built across the head of the debris fan. However, aggradation behind the walls raised the channel of the torrent. Then, in the late 17th century, several upland slopes (e.g. Gamp slope) were completely denuded by clearcut logging. Incipient instabilities on foliation dip slopes and along steep colluvial ravines supplied increasing amounts of debris to the torrent.

On 29 May 1731, a devastating series of debris flows burst forth onto the upper section of Schlanders. Waves of debris surged across the now buried protective walls and covered some 30 houses under more than 5 m of mud, sand, and rocks. After the catastrophe most of the buildings were excavated or rebuilt; but even today the entrance to the old church is almost 2 m below street level (Fig. 142). Construction also began on a massive diversion dam of dry stone masonry to confine the channel of the torrent to the eastern fan sector. The dam eventually reached a length of 300 m, a height of 10 m, and a maximum crest width of 5 m (Fig. 143). It is one of the most impressive protective structures in the Alps surviving from this early period (Stacul, 1979, p. 56-57).

Aggradation behind the dam since its construction has raised the eastern sector of the fan by several metres above the built-over western sector, but the dam has served well to the present day. Recently, check dams and transverse steel-beam rakes have been added along the channel. Resilient stands of pine are slowly reconquering the formerly denuded uplands, thus contributing to the stability of steep colluvial slopes. The former off-limits eastern sector of the fan now hosts the outdoor sports facilities of the town.

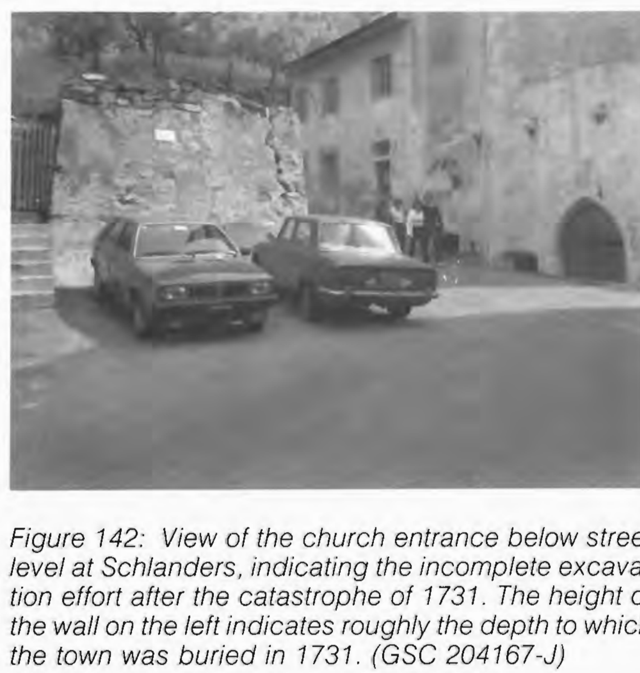


Figure 142: View of the church entrance below street level at Schlanders, indicating the incomplete excavation effort after the catastrophe of 1731. The height of the wall on the left indicates roughly the depth to which the town was buried in 1731. (GSC 204167-J)

Figure 143: Two views of the protective debris deflection dam between the channel of the Schladraun Torrent and the modern buildings of Schlanders. (a. GSC 204167-K, b. GSC 204167-L)

Meiringen (A53)

Location: Aare Valley, Bern, Switzerland (D2)
Date(s): 10 July 1733 (also 5 April 1650, July 1762, 1764, 26 May 1792, 1975)

The town of Meiringen (600 m) takes up the north side of the Aare Valley where the latter transects a high mountain range composed of complexly deformed southwest-trending calcareous rocks of the Helvetic cover complex (Fig. 144). In days past villages and hamlets were forced against precipitous bedrock walls or onto apices of gently inclined debris fans by the unruly Aare River. Meiringen grew around the apex of the Alpbach fan. The Alpbach Torrent reaches the valley floor via a series of waterfalls behind the town.

During the late Middle Ages large amounts of bedload, eroded from excessive clearcuts in the tributary basins, converted the river bed of the Aare into a braided network of shifting channels; the slightest blockage of the river by debris flows from tributary torrents created lakes and swamps on the extremely low-gradient floodplain. Repeated debris flows along the Alpbach Torrent, generally due to embankment failures in the uplands, often caused flooding near Meiringen.

On 10 July 1733 a serious debris flow buried a large part of Meiringen and sparked the construction of a dry masonry deflection wall. In July 1762 and 1764 massive flows overtopped this wall, dumping a debris blanket 5 to 10 m thick on the remnants of several ruined buildings. The present church near the apex of the Alpbach fan is the fifth on the site; its original foundation rests under 7 m of debris. Comprehensive training works along the Aare River, initiated in the mid-19th century, later supplemented by control works and arrays of check dams along some of the tributary torrent branches, have made life in Meiringen more tranquil (Ringgenberg, 1975).

By living close to the bedrock walls defining the valley the people of the region also learned how to cope with slope failures above their settlements. In 1649 herders working on the alpine pastures across the valley from Meiringen noticed that a huge crack had opened behind a limestone cliff above the hamlets of Balm and Falcheren at an elevation of 1400 m. In anticipation of a major collapse inhabitants in these settlements removed their belongings from the zone they considered too dangerous. On 5 April 1650, during springtime snowmelt, a rock spur, possibly in excess of $1 \times 10^6 m^3$, ruptured along the crack noted a year before and a disintegrating blocky lobe came hurtling down the mountainside. Preceded by a powerful airblast the debris lobe demolished several of the abandoned buildings. Today the rock avalanche deposits on the valley floor are overgrown by a protective and protected forest.

Similarly, on 26 May 1792, following a period of rapid snowmelt and intense rainfall, a large mass of calcareous slate and carbonate failed along the dip slope of the Engelhörner Massif. The resulting debris lobe split into two parts. Its larger western portion slid towards the Reichenbach Torrent, demolished several homesteads, killed three people, and impounded the swollen torrent. The eastern lobe descended the Geissholz gorge, spilled onto the debris cone of Geissholz and continued down a torrent channel towards Willigen. Both arms of the slide mass poured into the Aare Valley, inflicting considerable damage.

More recently, there has been concern about rockfalls in Meiringen. Scarcity of suitable development land had necessitated expansion of residential construction activity towards the historical rockfall chute of the Kirchberg cliff, an overhanging limestone ledge 300 m above the town (Fig. 145). As early as 1914 about 3000 m^3 of rock broke away from the cliff and tumbled onto the then unoccupied flat below. Since then, some twenty homes have been built there.

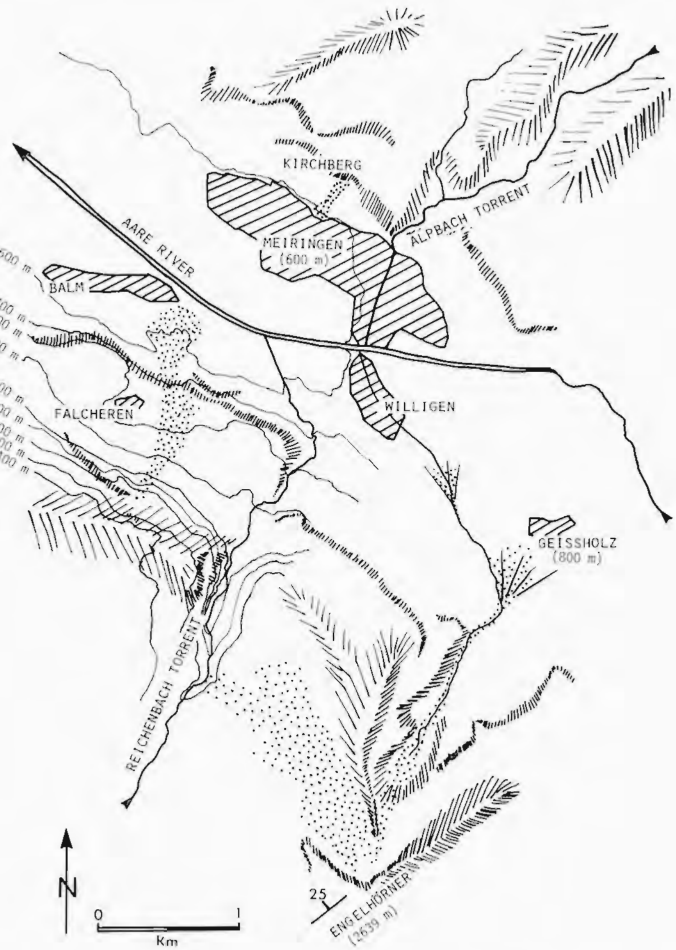


Figure 144: Index map of the vicinity of Meiringen; note the trained channel and floodplain of the Aare River which transects southwest-trending carbonate ranges.

Figure 145: Detachment zone and track of the artificially released Kirchberg rockfall along the outskirts of Meiringen. (GSC 204167-M)

By 1973, continued rockfalls totalling 60 000 m^3 pointed to a serious threat to human lives. After other alternatives had been pondered a decision was reached to blast away the unstable rock mass in three stages while the endangered buildings below were evacuated. The operation was carried out successfully in 1975. Damage inflicted by scattering blocks was minor compared to that which would have resulted if the whole unstable rock mass had failed unexpectedly and struck occupied homes (Neiger, 1978).

Férsina (A54)

Location: Trento, Trentino, Italy (G2)
Date(s): October 1747 (also 13th century, 1686)

The drainage basin of the Férsina Torrent, a major tributary of the Adige River, extends some 30 km into the mountains east of Trento (200 m), the capital city of the Trentino (Fig. 146). The Férsina Torrent drains upland basins underlain by carbonate and metamorphic rocks of the South Alpine basement and cover complexes. The torrent enters the Adige Valley through a narrow carbonate gorge below the conical surface of Pergine (500 m). The city of Trento itself spreads over the gently inclined debris fan of the Férsina on the eastern bank of the Adige River. From excavated Roman ruins within the city it appears that sand and gravel, deposited by debris floods of the Adige and Férsina, have raised the level of the townsite by 4 km during the last 2000 years.

Deforestation in the uplands, combined with rare storms, made life along the Férsina Torrent very difficult from the start. On the Pergine plain two hamlets, Arzenaca and Bracese, disappeared in debris floods as early as the 13th century (Strele, 1936, p. 127-128). Pergine protected itself by constructing heavy stone walls and dykes during the 16th century. Excessive bedload and debris carried by the Férsina frequently blocked the gorge above Trento and then broke forth in catastrophic debris floods sweeping through the old city centre. In 1537 the first wooden dam was erected at Pontalto (Ponté Alta) in the gorge to retain debris brought down by the river during storms. However, this protective structure collapsed in 1542 when the first serious onslaught of debris pounded against the timber abutments. Subsequent masonry dams at Pontalto did not fare much better: one collapsed in 1564, another in 1686; during the dam failure of 1686 Trento suffered considerable damage. After this, attention shifted to a scheme whereby the channel of the torrent was diverted onto the southern flank of the fan through a sparsely populated section of Trento; but, just to make sure, another stone-and-timber dam was also erected at Pontalto. This dam held until 1747 when sections of rotten timber gave way and a deluge of stony debris demolished numerous houses in Trento. In the decades that followed the dam was reinforced repeatedly and reached the impressive height of 40 m. It withstood the otherwise regionally catastrophic period of debris floods in 1882 (Seckendorff, 1884, p. 117-123).

In this century the course of the Férsina has been controlled by a combination of dykes, walls, check dams, and sills. Many potential debris sources in the uplands have been neutralized by reforestation; nevertheless, blocky bedload in much of the upper torrent channel still demonstrates the strong seasonal variations in the competence of the torrent. The waterfall over the dam of Pontalto is now a tourist attraction (Fig. 147); a marker details the history of the long battle which the people of Trento had to wage against the torrent.

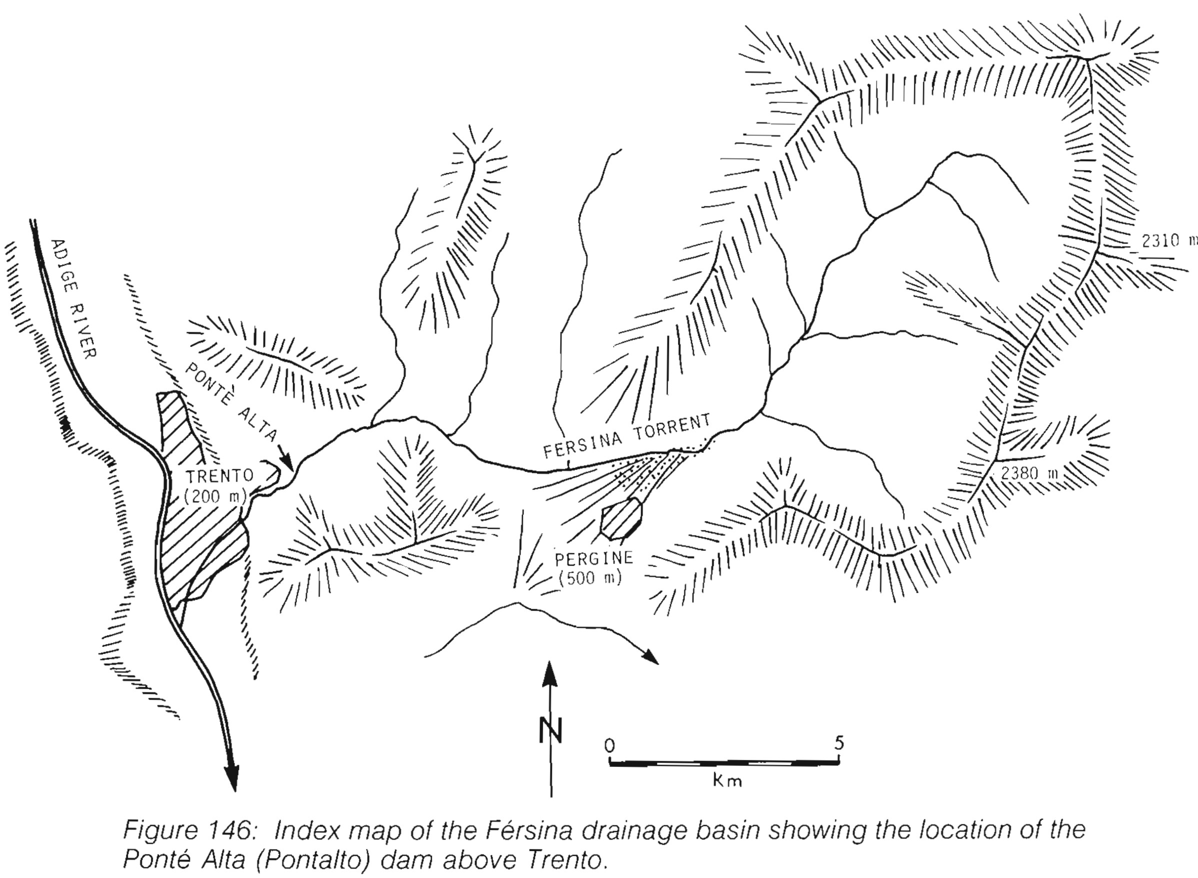


Figure 146: Index map of the Férsina drainage basin showing the location of the Ponté Alta (Pontalto) dam above Trento.

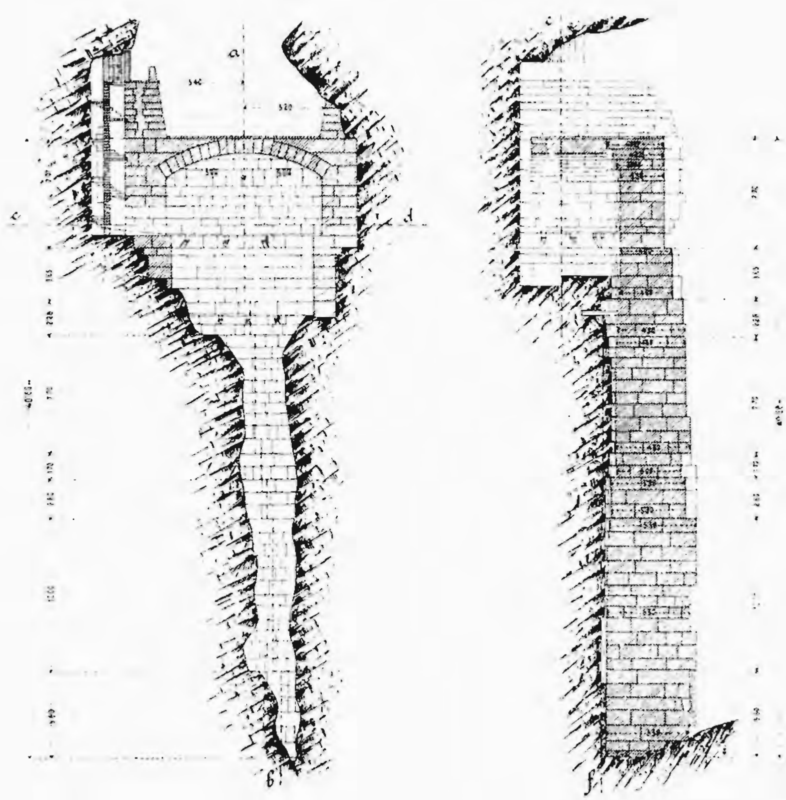


Figure 147: The debris retention dam at Ponté Alta (Pontalto) across the gorge of the Férsina Torrent; drawing from Seckendorff (1884).

Grigno (A55)

Location: Val Sugana, Trentino, Italy (G2)
Date(s): 18 August 1748 (also June 1564, September 1665, 16 September 1882, 4 November 1966)

The south-flowing Grigno Torrent drains a large catchment area underlain by a variety of rock types of the South Alpine basement and cover complexes, which are locally covered by late Pleistocene surficial deposits. The torrent enters the Val Sugana through a narrow carbonate bedrock gorge and crosses a gentle fan on the north side of the Brenta River (Fig. 148). In the past, the community of Grigno (250 m) huddled against the apex of the fan to avoid the regular floodwaters of the Brenta. It was thus exposed to sporadic debris flows caused by blockage of the lower Grigno gorge. Destructive debris flows such as those of June 1564 and September 1665 were generally followed by long periods of deceptive tranquility. By the early 18th century many of the formerly forest-covered colluvial benches along Grigno Torrent had been almost completely denuded in large clearcuts; erosion along steep gullies was intensified during rainstorms. On 18 August 1748, a regional rainstorm triggered numerous debris avalanches in the surficial deposits bordering the torrent. Pulses of debris collected into a massive flow which swept Grigno, killing 19 people. After this event protective forests were established along the most unstable ravines and slopes of the upper Grigno Valley. However, on 16 September 1882, excessive runoff and infiltration of water during one of the greatest storms in the history of the southern Alps triggered 35 major debris avalanches in the Grigno basin sending surges of debris towards the gorge. Bursting forth in waves, the debris accumulated to a depth of 4 to 7 m and destroyed 28 houses in Grigno (Strele, 1936, p. 130 and 134; Seckendorff, 1884, p. 161-162). Beginning in 1883, check dams were erected along the torrent and a set of 3 m high stone walls was erected on the fan to guide the torrent through the village. Immediately above Grigno a confining bedrock nose was blasted away to allow the easier passage of debris through the gorge. In recent years, the destructive impact of mass movements has been confined to a few precarious debris chutes carved into colluvial upland terraces. During the storm of 4 November 1966, with its 200 m of rain in three days, a debris avalanche destroyed several buildings in Molini (Gorfer, 1977, p. 949). At the same time erosion along one of the most notorious debris sources, the Frana di Casa Campestrin, north of Pieve Tesino (871 m), undercut several buildings and threatened to expand towards and thus engulf the whole village. Approximately 3 $\times$ $10^6 m^3$ of unstable debris is still present in this area (Largaiolli and Siro, 1977). Since 1975 check dams and terracing, combined with revegetation have been used to counteract erosion in tributary gullies of the torrent, including the Frana di Casa Campestrin.

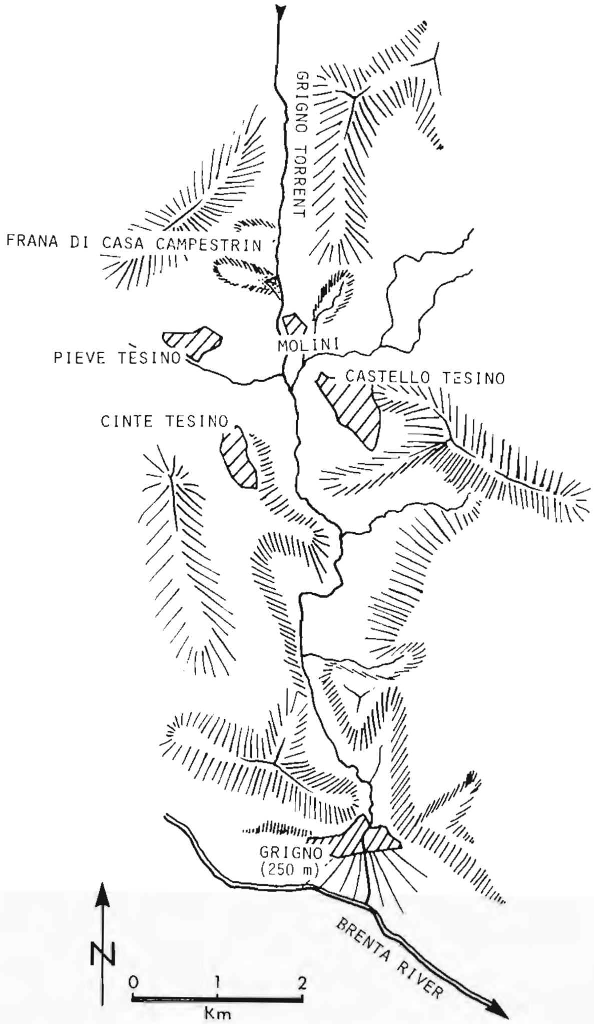


Figure 148: Sketch map of the lower Grigno basin showing the location of major debris sources near Pieve Tesino and the Grigno fan.

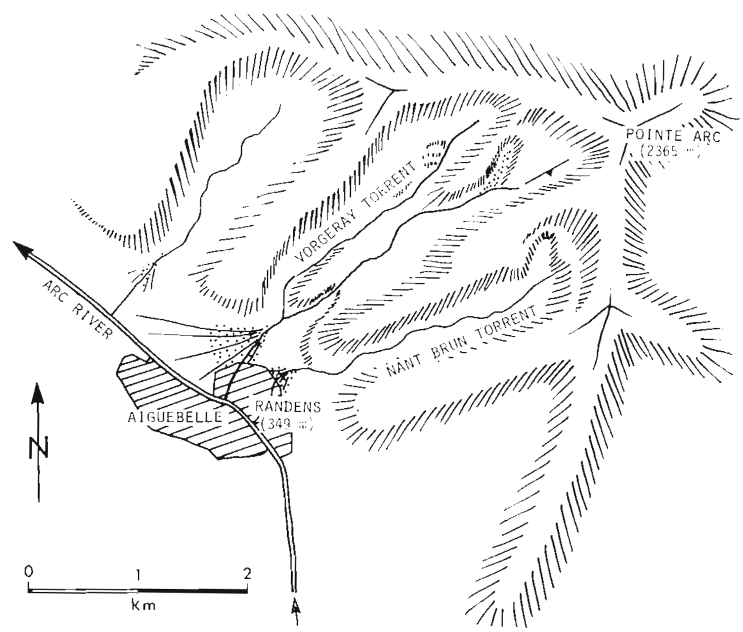


Figure 149: Sketch map of the composite debris cone of Randens and high-level source areas of historical debris flows down the Vorgeray and Nant Brun torrents.

Randens (A56)

Location: Aiguebelle, Arc Valley, Savoie, France (B3)
Date(s): 12 June 1751 (also 12 June 1760, 7 June 1806)

Randens (349 m) is located at the periphery of two coalescing debris cones along the eastern bank of the Arc River opposite the town of Aiguebelle (Fig. 149). The debris cones extend from the river to the steep gorges of the Vorgeray and Nant Brun torrents which are carved into mica schist terrain of the Helvetic basement complex. Scarps along steep dip slopes below Pointe Arc (2365 m) indicate sagging and incipient slumping of the intensely deformed metamorphic bedrock in the uplands of both torrents. In the early 18th century large sections of the uplands lost their forest cover due to clearcut logging.

Between 1748 and 1751 heavy rains and rapid spring snowmelts set off massive slides along tributary ravines of the Vorgeray. These slides developed into debris flows which reached the upper part of Randens, then located on the apex of the Vorgeray cone. The most serious flow in this series occurred on 12 June 1751; it almost demolished the church. A thunderstorm on 12 June 1760, mobilized an even greater volume of unstable debris along the upper Vorgeray

Figure 150: Recently excavated and renovated building foundations at the apex of the Nant Brun Torrent cone; note two generations of walls on the far side of the building. (GSC 204167-N)

gorge, and a fast moving lobe containing huge slabs of mica schist engulfed stables and livestock in the village. Another cycle of debris flows seriously threatened Randens between 1806 and 1812. On 7 June 1806, one flow smashed into two buildings and killed 4 people. (Mougin, 1914, p. 1165-1169).

Since 1807 stone masonry dykes along the axis of the Vorgeray cone have successfully contained minor debris flows. In the uplands reforestation has contributed to stabilizing embankments of debris chutes and ravines. In recent years residential construction has extended onto the bouldery debris of the upper cones. Here, formerly buried building walls locally serve as pedestals for new homes (Fig. 150).

Massif de Platé (A57)

Location: Servoz, Arve Valley, Haute Savoie, France (B3)
Date(s): 4 August 1751 (aslo 'early in the Christian era', February 1471, 17 September 1852, 16 April 1970)

The Massif de Platé (2553 m) overlooks the upper valley of the Arve River and the glacier-covered northern buttress of Mont Blanc, the highest mountain of Europe. Near Servoz (870 m) the Arve River leaves the granitic basement terrain, crosses an ancient lake bottom (Le Lac), and drops west-northwesterly into a narrow gorge carved into north-dipping shale-carbonate units of the Helvetic cover complex underlying the Massif de Platé (Fig. 151). North of the Arve gorge the land rises smoothly to the terrace of Plaine Joux (1300 m) and then in a series of steps to a serrated summit ridge known as Rochers des Fiz (2200 m). The flat-topped ridge is crowned by massive Urgonian limestone. Bedding dips 10 to 25° to the north-northwest. The orientation of the carbonate scarp face of Rochers des Fiz is controlled by a system of east-northeasterly trending fractures; fractures also penetrate the karst plateau behind the summit ridge (Goguel and Pachoud, 1978).

In the past the slopes below the Platé Massif have been affected by bedrock slumps. Most of the area is covered by a blanket of angular limestone debris resulting from rockfalls from the Urgonian cliff. Torrent gorges cross this unstable terrain.

Sometime during the early Christian era a large slide mass from the sagging toe zone of the Platé slope apparently blocked the gorge of the Arve, impounding a lake that flooded the plains of Servoz. The lake waters eventually overflowed, not into the original gorge but into the parallel gorge of Le Châtelard, located 0.5 km to the south. The lake seems to have persisted until the 13th century when a failure of the debris dam cleared the old channel, re-establishing the old course of the river. The catastrophic debris flood that resulted from this burst wiped out a locality named St. Denis (Dyonisia) in the Arve Valley below (Mougin, 1914, p. 268).

Towards the end of February 1471 a major segment of the sagging terrain failed again, raising the river channel to a height of 150 m above its normal level. Once more, rising waters threatened to inundate the community of Servoz. However, before this could happen the inhabitants of the area excavated a ditch across the slide mass, thus draining the lake without further damage to property above or below the gorge (Mougin, 1914, p. 268).

On 4 August 1751, a major section of the summit ridge of the Rochers des Fiz began to subside on a large scale at the Col du Derochoir. Overhanging slabs of limestone crashed onto the terrace of Plaine Joux and a major rock avalanche obliterated six houses and killed six people. Subsidence and outward flow of the argillaceous strata below the Urgonian cliff continued for more than a month; snapping noises and dust clouds soon attracted the attention of the people in the surrounding villages and rumour reached the royal court in Torino that a volcano had erupted in the region. To check on the truth of these alarming reports the scientist Donati was sent out to ivestigate. After completing his field assignment Donati could report:

'. . . that after travelling for four days and two nights without stopping I found myself in the face of a mountain completely enveloped in smoke, from which broke away continuously, day and night, masses of rock, with an astounding noise stronger than thunder or the battery of a cannon. All the peasants had left the neighbourhood, and only dared to watch these collapsing cliffs from a distance of more than two miles. The surrounding fields were covered with dust, much resembling ashes; in some places the dust had been carried by the winds to distances of five miles. The peasants said that they had seen at times a red smoke during the day and that during the night it was accompanied by flames. These observations created the general belief that a new volcano would be found; I examined the so-called ashes and found that the dust was nothing but pounded marble. I observed the smoke attentively but I saw no flames and received no odour of sulphur, and neither the creeks nor the springs presented signs of sulphurous matter. Thus I was convinced that there were no sulphurous emanations. I passed through the smoke, and, although alone and without escort, I advanced to the edge of the defile. I saw an immense crag that was crumbling into the abyss, and I observed that the smoke was nothing but the dust arising from the fall of the stones. I searched for and later found the reason for the collapse of these rocks; because I saw that a great part of the ground below the disintegrating mountain consisted of earth and stones that were not ordered in layers or beds, but were superimposed without order. From this I could see that at this very mountain similar collapses had occurred before and had left the crag that fell this year without support and a considerable overhang. That cliff consisted of of horizontal beds, of which the two lowest were made up of shale or a laminated fractured rock without much consistency; the two upper beds resembled the marble of Porto Venere, but they were split by many transverse fissures. The fifth layer again consisted of shale completely broken in vertical slabs and forming the upper surface of the fallen mountain. On this upper surface there were three lakes whose water penetrated the openings of the rock layers that

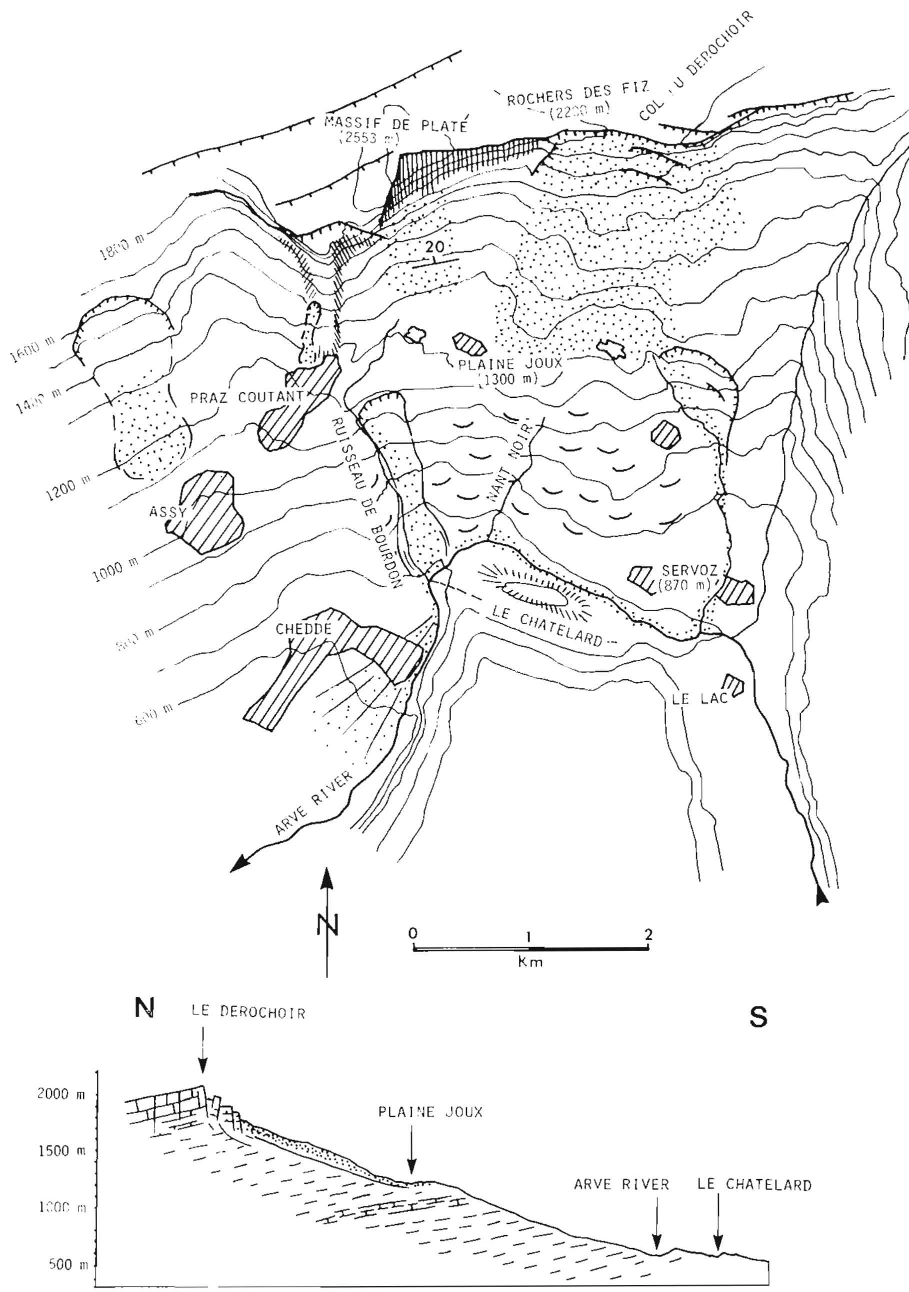


Figure 151: Sketch map and geological cross-section of the Massif de Platé (after Pachoud, 1976).

were thus separated from each other and undermined their foundations. The snow that had fallen this year in the Savoie higher than ever in memory, increased this force to such an extent that it caused the fall of 3 million cubic spans (= approximately 20 $\times$ $10^6 m^3$) of rock, enough to form a mountain of its own. . . .' (de Saussure, 1781, v. II, paragraph 493).

Donati's remarkable report concluded with the prediction that the slope movements would come to rest, which eventually they did. The slightly concave re-entrant of the Col du Derochoir has since remained fringed by huge leaning slabs of limestone (Fig. 151). It is flanked by two deeply fractured bedrock spurs.

On 17 September 1852, an intense rainstorm set off large bedrock-debris slumps at the headwaters of the Nant Noir and Ruisseau de Bourdon torrents below the terrace of Plaine Joux. For almost a year masses of debris descended to the bed of the Arve River threatening repeatedly to stem its flow. Debris flow activity along these two torrents continued to be substantial until about 1856 (Goguel and Pachoud, 1978).

Figure 152: View of the sanatorium of Praz Coutant; note the snow-debris avalanche track behind the building. (GSC 204167-O)

During the winter of 1969-1970 most of the Savoie experienced exceptional snowpacks and snow-avalanche activity. The spring snowmelt was greatly delayed by late snowfalls. On 5 April 1970, a small wet-snow avalanche slightly damaged a sanatorium on the thinly forested bench of Praz Coutant. On 12 and 13 April 1970, about 50 mm of rain fell on the area and infiltrated the slope above Praz Coutant, which is inclined 30-40° and is underlain by slaty marl capped by about 2 m of colluvium. On 16 April 1970, a blanket of colluvial debris several tens of metres wide cracked open along a head scarp 300 m above the sanatorium of Praz Coutant. The mass of soil and trees, about 30 000-40 000 m^3 in volume, followed the path of the snow avalanche and crashed in to the western wing of the building (Fig. 152). The walls that had resisted the impact of the snow avalanche collapsed under the impact of the debris, and 72 people, mostly children, lost their lives (Jail and Vivian, 1971, p. 475-482).

Today several sanatoriums, hotels, and private homes dot the 'sun slope' of the Platé. An elaborate network of roads provides access for hikers and excursionists who come to enjoy the magnificent view of Mont Blanc. It is unlikely that past catastrophes will greatly inhibit use of this mountainside in the future. The site at Praz Coutant which suffered from the debris avalanche of 1970 has been protected by a high deflection dam. Several active slumps along the terrace rim of Plaine Joux are being drained by open ditches. The possibility of a major collapse of the Rochers des Echines above Praz Coutant has been recognized: it would not only endanger Praz Coutant but also the town of Chedde in the Arve Valley (Fig. 151). The importance of monitoring incipient movement along the cliff has been emphasized by Pachoud (1976) and Goguel and Pachoud (1978).

Grächen (A58)

Location: Matter Valley, Wallis (Valais), Switzerland (C2)
Date(s): 9 December 1755 (also 12 September 1855)

Grächen (1600 m) is a large village spreading over a bedrock terrace 600 m above the northeast-trending Matter Valley and commanding a spectacular view of the surrounding mountain ranges. The terrace is a remnant of an old valley bottom carved into gently west-dipping metamorphic successions of the Pennine core zone (Fig. 153). Rising above the Grächenwald forest the mountain spur of the Durlochhorn (2723 m) consists of semidetached rock towers and sagging bedrock terrain mantled by rock avalanche deposits which extend to the fringes of Grächen. The area surrounding Grächen is one of the most seismic regions in the central Swiss Alps (Pavoni, 1977).

On 9 December 1755, probably as a consequence of an earthquake whose Mercalli intensity in Visp has been estimated at VIII, a part of the Durlochhorn failed and a rock avalanche buried one third of Grächen. Similarly, a series of earthquakes in July 1855 probably set the stage for serious rockfalls on 12 September 1855, which damaged several buildings in the village (Montandon, 1933, p. 307-308).

In recent years Grächen has developed into an impressive tourist centre and building activity has approached the rockfall lobes of past centuries (Fig. 154). The Grächenwald forest above the community will provide protection only against minor rockfalls.

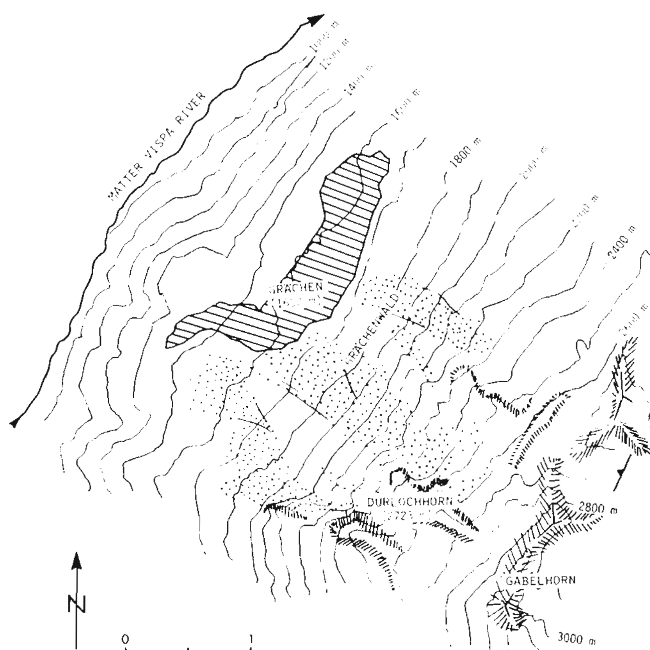


Figure 153: Sketch map of the bedrock terrace of Grächen above the valley of the Matter Vispa; note blocky rockfall debris below the Durlochhorn.

Figure 154: View of the lower slopes of the Durlochhorn and the conical rockfall deposits behind Grächen. (GSC 204167-P)

Umhausen (A59)

Location: Ötz Valley, Tirol, Austria (F1)
Date(s): 9 to 11 July 1762 (also summer 1807, 2 August 1851, 18 June 1855, 29 June 1965)

The Ötztal (Ötz Valley) is a deep north-trending glacially-sculptured trough that crosses gneiss, amphibolite, and mica schist terrain of the Austroalpine basement complex. Mountains bordering the valley attain elevations of more than 3000 m (Fig. 155). The course of the Ötztaler Ache (river) is confined by rock avalanche deposits, rockfall lobes and debris cones. The most conspicuous of these obstacles, the gigantic Taufererberg rockslide, has a volume of $2100 \times 10^6 m^3$ and originated by failure of a gneissic ridge on the west side of the valley above Umhausen (1036 m) about 8700 years ago (Erismann et al., 1977).

The prehistoric Taufererberg slide mass originally blocked the Ötztaler Ache and its eastern tributary, the Horlachbach Torrent. Erosion of the fractured rock mass by the Horlachbach Torrent has created the large debris cone of Umhausen. The aggradational flat of Längenfeld south of Taufererberg also receives several high-gradient torrents.

The people of the valley, here as in other parts of the Alps, preferred the sporadic threat of debris flows to the regular floods of the river (*see* Ötz Valley Ice Floods). However, at times the debris flow hazard has been substantial. The name Umhausen (= resettlement) suggests that the community had a predecessor elsewhere on the cone. Between 1760 and 1770 Umhausen again was in serious difficulties and during the same decade debris flows forced complete evacuation of the nearby hamlet of Östen. Between 9 and 11 July 1762, an intense rainstorm filled the gullies and ravines above Umhausen with debris which broke forth simultaneously demolishing 70 buildings and killing 9 persons. Reconstruction of Umhausen shifted partly to Neudorf (= new village) a locality farther down the debris cone (Strele, 1936, p. 130-131; Leys, 1977, p. 68-73).

In the summer of 1807 a large debris flow buried a hamlet located between Umhausen and Längenfeld. At the same time Östen was demolished once more. On 2 August 1851 and on 18 June 1855 debris flows again caused damage in the central Ötz Valley.

Since the mid-18th century the most notorious debris cones have been left undeveloped with broad strips of protective forest along their axes (Fig. 156). However, the recent development boom has brought some residential buildings close to the dangerous torrents. Sporadic debris flows, exceeding the design magnitude of protective dams, walls, and linings, demonstrate the limits of the 'achievable' in this kind of environment. For example, on 29 June 1965, an intense regional rainstorm caused severe erosion in the uplands of the Fischbach Torrent; Längenfeld suffered heavily when a debris flow and flood carrying some $0.15 \times 10^6 m^3$ of bouldery debris broke across dykes and invaded the town at the beginning of the tourist season (Heuberger, 1975; Leys, 1980).

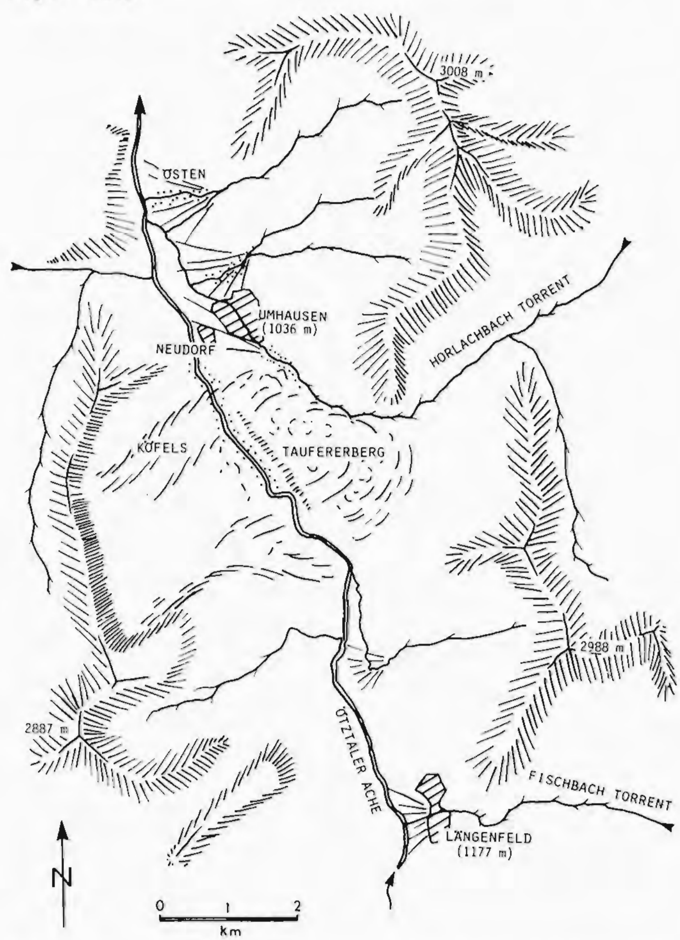


Figure 155: Index map of the central Ötztal area; note the gigantic prehistorical Taufererberg rockslide and the inhabited debris cones at the mouths of tributary torrents.

Figure 156: View of the hamlet of Österreich; note dense protective forest along the axis of the cone (background), the lone church (a remnant of a former settlement), and the farm buildings on the flank of the cone. (GSC 204167-Q)

Neumarkt-Egna (A60)

Location: Etsch (Adige) Valley, Südtirol (Alto Adige), Italy (G2)
Date(s): 18 November 1767

The towns of Neumarkt (Egna) and Vill (226 m) are situated on the large debris fan of the Truden Torrent which drains parts of a rhyolite plateau (1200 m) of the South Alpine basement complex on the east side of the Etsch Valley (Fig. 157). Cutting through extensive late Pleistocene surficial deposits in the upper Truden basin, this torrent has built a substantial fan out into the Etsch Valley. Many other fans and cones in this valley are of similar origin and all of them have been traditional sites of settlement, occupied prior to the first attempts to control the floods of the unruly river.

In the 13th century a predecessor of the village of Neumarkt apparently was located along the outer periphery of the Truden fan; about 1221 (?) it was swept away by a flood of the Etsch River. Subsequent growth of two communities, Neumarkt (=new market) and Vill, therefore shifted up the debris cone. There sporadic flows of debris threatened both villages, necessitating the construction of primitive protective walls.

On 18 November 1767, a rainstorm deluged the eastern Etsch basin (Sonklar, 1883, p. 86). One of the many torrents that were choked by the debris eroded from tributary ravines was the Truden Torrent. As a result, a massive flow swept into Neumarkt, obliterating several buildings and killing 20 people (Strele, 1936, p. 131). After this catastrophe two huge stone-masonry dykes, 800 m long and 6 m high, were built along the channel of the torrent to guide debris flows straight towards the river. Check dams constructed since 1884 have contributed substantially to the stabilization of embankments in the upper Truden basin. Today the cone hosts highly productive vineyards and the two expanding communities.

Gaishorn (A61)

Location: Palten Valley, Steiermark, Austria (J1)
Date(s): 1768

The Palten Valley is a wide trough-shaped depression that follows a northwest-trending fault zone in phyllitic-calcareous bedrock formations of the Austroalpine basement complex. The low-gradient Palten River receives torrents via debris cones from both sides of the valley (Fig 158). One of

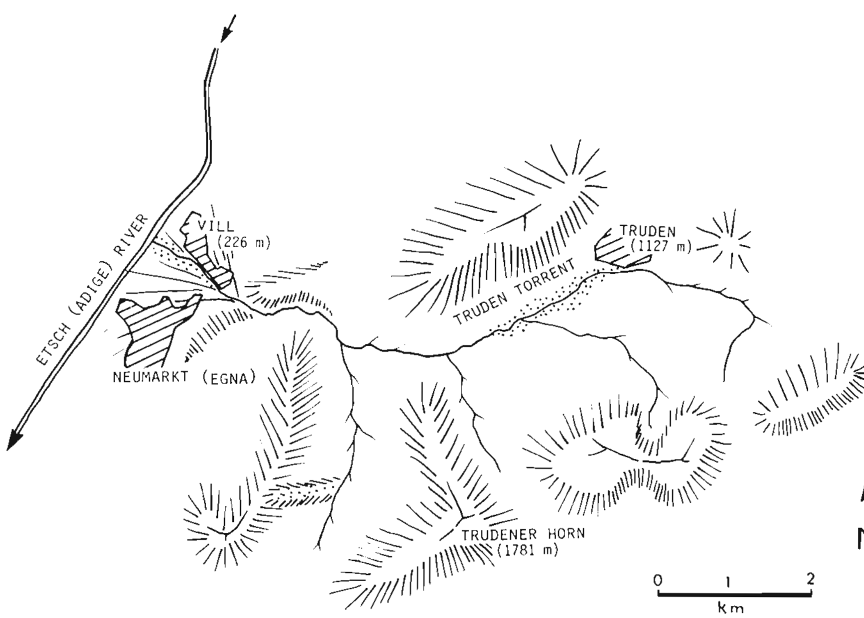


Figure 157: Sketch map of the basin of the Truden Torrent and the neighbouring communities of Vill and Neumarkt (Egna).

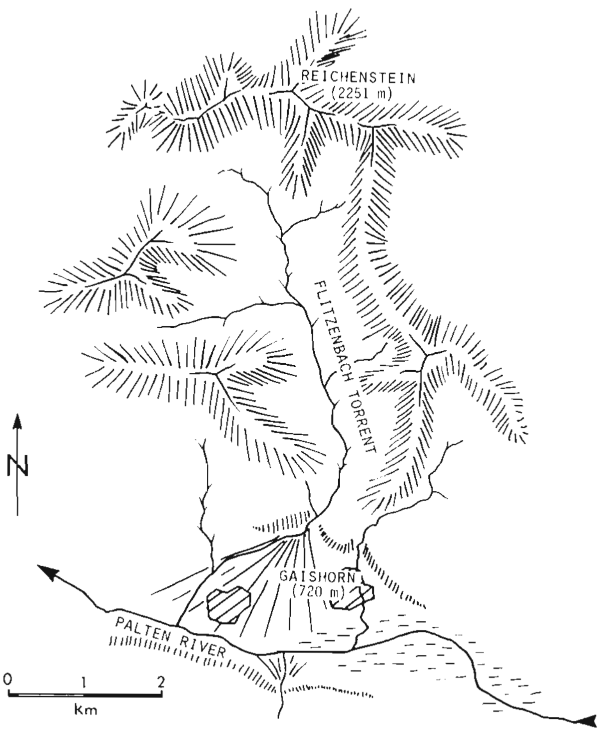


Figure 158: Sketch map of the Flitzenbach basin, the Gaishorn cone, and the swampy terrain of former Lake Gaishorn.

the largest debris cones is that built by the Flitzenbach Torrent, whose tributary branches collect runoff from bare calcareous cliffs of the Reichenstein Massif (2251 m) north of Gaishorn (720 m) and whose lower part is deeply incised into unstable phyllitic bedrock which tends to slump or creep across the channel.

Beginning in 1746 serious blockages of the Flitzenbach channel by embankment failures set off debris flows which repeatedly engulfed the community of Gaishorn. A massive flow during a regional rainstorm in 1768 not only ravaged the village but also blocked the Palten River (Stini, 1938). As a result, a lake formed and flooded productive agricultural land upstream from the cone. Continued debris flow activity from the Flitzenbach basin not only restricted use of the cone to its easternmost sector, but also maintained the Gaishornsee (Lake Gaishorn) for more than 150 years. In 1925 it was decided to drain the channel of the Palten River and thus regain the flooded land. At the same time check dams and other control works were built in the Flitzenbach basin to diminish the input of fresh debris by the torrent. Nevertheless, debris movements continued with occasional blockage of the Palten River; the river channel had to be dredged repeatedly. At the same time swampy conditions prevailed along the former lake bottom. In recent years there have been plans to re-establish the lake as an economically and environmentally more appealing alternative to the costly dredging operations along the periphery of the Gaishorn cone. A recent housing development also has invaded the central sector of the debris cone.

Monbiel (A62)

Location: Klosters, Graubünden (Grisons), Switzerland (E2)
Date(s): 17 June 1770

The hamlet of Monbiel (1291 m) crowds the western flank of a large debris cone at the foot of the south-facing Schildflue Massif (2887 m). The scarp face of the mountain above Monbiel is sculptured by deep ravines and snow avalanche chutes (Fig. 159). Bedrock consists of Pennine phyllite and dolostone, overlain in tectonic contact by high grade metamorphics of the Austroalpine basement complex. Foliation, bedding, and the tectonic contact dip approximately 20° to the northeast; numerous fractures parallel and perpendicular to the valley fragment the rock faces.

During the winter of 1769-1770 heavy snowfalls, occasionally mixed with rain, created a thick snowpack throughout western Austria and eastern Switzerland. Spring of 1770 was delayed and snow began to melt in May. The delayed but sudden snowmelt was accompanied by several rainstorms during which slope failures and debris flows occurred widely in this part of the Alps.

On 17 June 1770, intense rainfall and snowmelt triggered the collapse of approximately 70 000 m^3 of fractured gneissic and carbonate bedrock along a steep valleyward-dipping fracture zone above Monbiel at an elevation of 1700 m. Channeled by the steep embankments of a bedrock ravine, the crumbling rock mass fanned over the cone of Monbiel, erasing 13 houses and killing 17 people (Blumenthal, 1925; Heim, 1932, p. 122).

After the catastrophe most of the community has remained in its somewhat protected niche astride the cone and away from the regular snow avalanches from the Schildflue slope (Fig. 160).

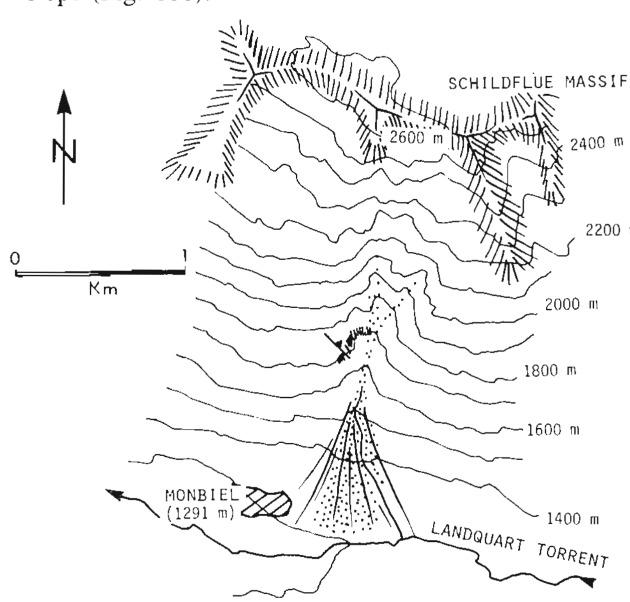


Figure 159: Sketch map of the debris cone east of Monbiel.

Figure 160: View of the rock avalanche lobe of 1770 on the cone of Monbiel. Farm buildings on the lower cone are located in the fall line of rare far-reaching snow avalanches from the Schildflue slopes. (GSC 204167-R)

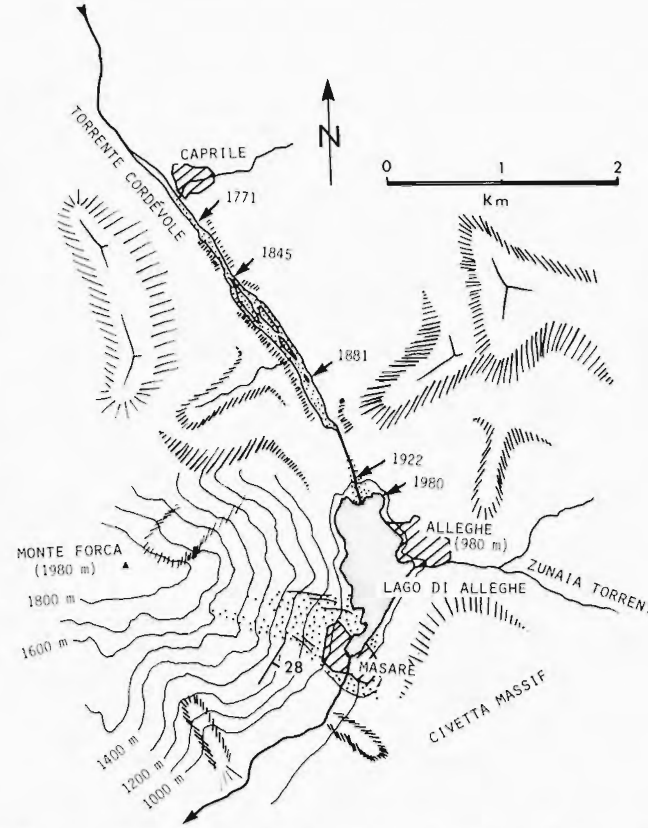


Figure 161: Sketch map of the environs of Lago di Alleghe and the carbonate dip slope of Monte Forca; note the historical change in the position of the Cordèvole delta front (indicated by years).

Lago di Alleghe (A63)

Location: Cordèvole Valley, Veneto, Italy (G2)
Date(s): 11 January 1771 (also 1 May 1771)

Lago di Alleghe has one of the most enchanting natural settings in the Southern Alps. The scenic village of Alleghe, the dark forests behind it, and the white scarp face of the Civetta Massif looming in the background have been illustrated in many travel books. The lake owes its origin to a catastrophic rockslide that broke away from the (never depicted!) west side of the Cordèvole Valley (Fig. 161).

The valley walls flanking the Torrente Cordèvole near Alleghe (980 m) are part of an east-dipping panel of carbonate and shale of the South Alpine cover complex. The dip slope of the western valley wall is composed of limestone and marl rising to the spur of Monte Forca (1980 m). Bedding dips at an average angle of 28° to the southeast.

On 11 January 1771, a slab of $20 \times 10^6 m^3$ of thick-bedded carbonate rocks failed above a horizon of thin-bedded argillaceous limestone below Monte Forca and crashed to the bottom of the valley. Prior to the collapse of the rock cliff the peasants working on the high pastures of Monte Forca had noticed opening of cracks which to them appeared as if 'a plough had just passed through'. The slide erased three hamlets and killed their 49 inhabitants. The front of the rock avalanche crossed the Cordèvole Valley and climbed more than 50 m up its east wall, blocking the flow of the Cordèvole Torrent. One by one, five hamlets on the upstream side of the slide mass disappeared beneath the waters of a rising lake which soon extended 4 km upstream (Fig. 162). In February 1771 an engineering mission from Belluno arrived at the site of the new lake and contemplated the possibility of draining it. It was quickly recognized that this was impossible. However, an unanticipated second failure on the Monte Forca dip slope was to prove more disastrous than the feared overflow of the lake.

On 1 May 1771, a remnant of the overhanging Monte Forca cliff broke away from its precarious pedestal and the disintegrating mass plunged into the newly formed lake, setting up a wave that surged over the opposite shoreline and up the channel of the Zunaia Torrent. In the village of Alleghe the wave damaged the church and killed three people (Casal, 1898; Angelini, 1976, p. 45-57). The rockslide barrier at the lower end of the lake withstood the impact of the wave train. Later, stone-masonry and concrete linings were installed to stabilize the overflow channel on the blocky Masarè (=debris).

Since 1771 numerous buildings, including the romantic Villa Paganini, have turned the formerly forbidding Masarè into a sizable community of vacation homes. However, the catastrophe of 1771 did not remove all unstable rock from the failure surface above the Cordèvole Valley (Fig. 163). The southern section of the Monte Forca is still a downward tapering wedge of dolomite similar in cross-section to the slab that failed in 1771. Several buildings have recently sprung up at its base.

During every major rainstorm the Cordèvole Torrent adds an enormous load of sand and silt to Lago di Alleghe (Trener, 1957). The original north arm of the lake has changed into a wide braided reach and the depth of the lake has decreased from 50 to approximately 15 m.

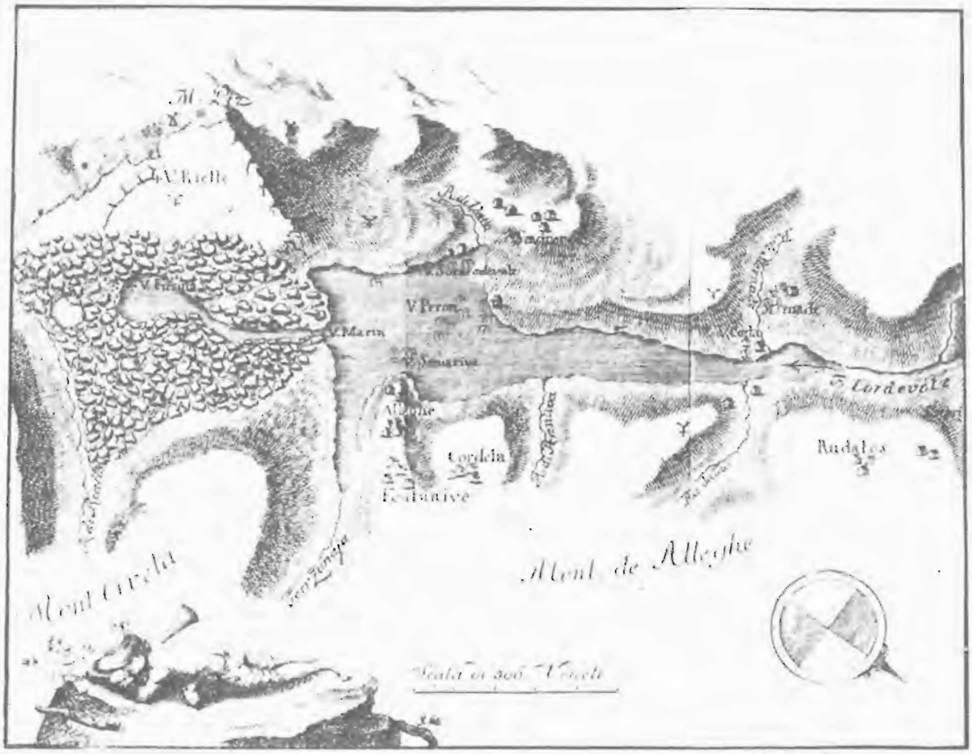


Figure 162: Map (made in 1785) of the impounded Cordèvole Torrent and the Lago di Alleghe; drowned hamlets have been indicated by the artist (from Angelini, 1976).

a

Figure 163: a) View of the Monte Forca detachment zone and rock avalanche deposits of 1771; note recently constructed vacation homes on the blocky Masarè (photo taken in 1980) (GSC 204167-S). b) View along the detachment surface of the rockslide to Lago di Alleghe; village of Alleghe on the left (GSC 204167-T).

Fucine (A64)

Location: Val di Sole, Trentino, Italy (F2)
Date(s): 17 September 1772 (also 1 September 1757, 10 October 1789, 15 August 1846, 25 September 1885)

Fucine rests on a small cone at the confluence of the Vermigliana Torrent and the Noce Torrent. From the head of the cone a large basin opens westward to Passo del Tonale. The Vermigliana Torrent receives a number of steep tributaries from the north and south which drain basins underlain by gneiss, phyllite and granite of the South Alpine basement complex. These tributaries join the main valley across short but well defined cones covered with blocky debris. Because of the large area of the Vermigliana drainage basin, destructive debris flows at Fucine generally coincide with major regional floods of the Adige and Inn rivers, and are due to blockage of bedload in the narrow bedrock passage above the apex of the Fucine cone.

On 1 September 1757, following a series of thunderstorms, pulses of debris were deposited among the buildings of Fucine. On 17 September 1772, three days of rain accompanied by warm south winds triggered great floods along the Inn and Adige rivers and set off numerous slides in the basin of the Torrent Vermigliana. In Fucine 22 buildings were destroyed by a raging debris flood and half of the nearby town of Ossana disappeared under a mass of rubble that had broken away above a steep ravine. Damaging flows also entered Fucine after the warm rains on 10 October 1789, 15 August 1846, and 25 September 1885 (Strele, 1936, p. 131; Gorfer, 1975, p. 846).

Today the undulating grassy cone of Fucine is still pierced by occasional blocks of granite. Parts of the Vermigliana Torrent are controlled by old stone revetments and sills; however, upstream from Fucine the braided reaches of the torrent are uncontrolled.

Kalkkögel (A65)

Location: Innsbruck, Tirol, Austria (G1)
Date(s): 1 July 1782 (also 1575, 24 July 1750, 8 August 1807, 25 July 1846, 2 June 1908)

The Kalkkögel Massif (2600 m), southwest of Innsbruck, is a deeply dissected complex of flat-lying carbonate formations resting on metamorphic rocks of the Austroalpine basement complex. The upland basins of the main torrents are directly underlain by carbonate rocks; the lower reaches of the torrents are carved into locally unstable schistose bedrock and late Pleistocene colluvium. Debris cones at the mouths of the three torrents draining the massif accommodate the flourishing communities of Fulpmes, Götzens, and Axams (Fig. 164). Clearcutting and overgrazing in the uplands during the Middle Ages enhanced the debris flow hazard which in this area is related to the frequent summer thunderstorms.

Götzens (868 m) suffered debris flow damage at least as early as 1575. The main source of the debris is a deep erosional scar (Grosse Blaike) in a bench of relict colluvium at 1500 m on the west side of the Geroldsbach Torrent. On 24 July 1750, rapid runoff during a thunderstorm generated a massive debris flow that invaded the church and damaged 20

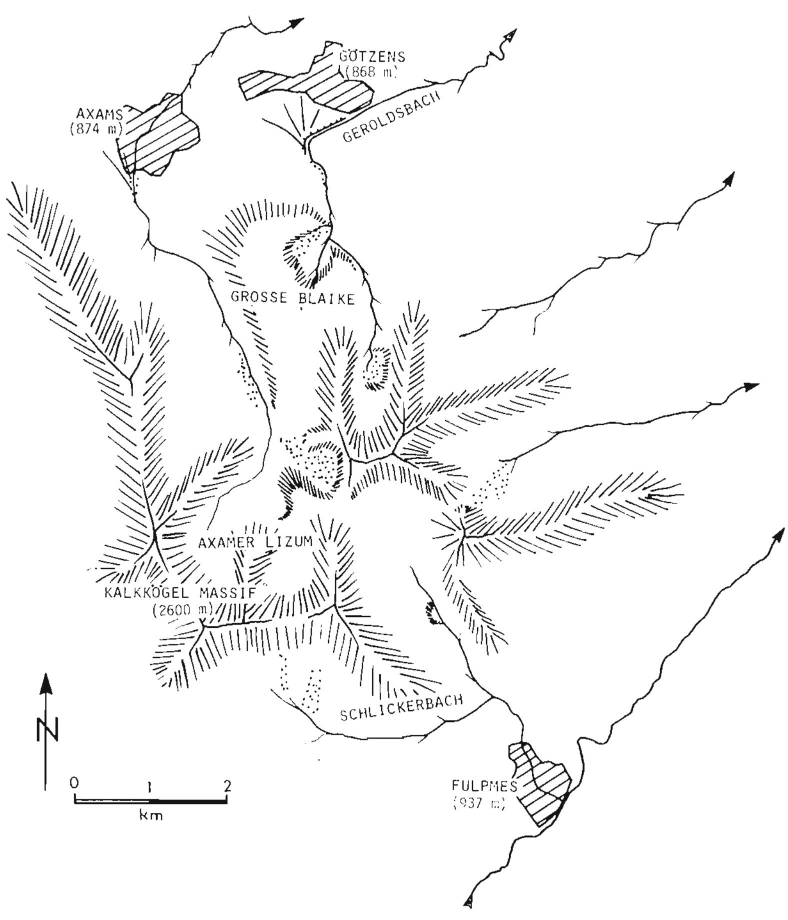


Figure 164: Index map of the Kalkkögel Massif and the major communities on its debris fans; note the forced deflection of the torrent above Götzens.

Figure 165: Protective dam under construction above the apex of the Axams cone; note vertical slits in the dam which allow selective discharge of normal bed-load and minor debris flows. (GSC 204167-U)

houses. On 1 July 1782, another terrible cloudburst-hailstorm mobilized masses of bouldery debris from the Grosse Blaike; this debris flared across the cone of Götzens demolishing 22 houses and killing four people. After this catastrophe the community set to work on a stone-masonry deflection dam to divert debris in the Geroldsbach gorge east away from the village. The cost of maintaining this structure proved to be considerable because debris from minor flows soon filled the space behind the dam. On 25 July 1846, during another thunderstorm, debris completely filled the bend of the deflection dam and a 2 m thick lobe overtopped the wall covering fields near Götzens. The protective structure was then raised to a total height of 10 m and reinforced to a thickness of 4 m. However, the scarred source area responded with new pulses of debris during every major storm. On the evening of 2 June 1908, during a cloudburst-thunderstorm debris with boulders reaching 2 m in diameter again breached the protective dam and in Götzens 22 houses disappeared in the stony lobe (P. Frey, personal communication, 1980).

The torrent systems of the Kalkkögel area also are active during regional rainstorms (e.g. 1807, 1855, and 1882). During such a storm on 8 August 1807, Fulpmes was almost entirely destroyed by massive debris movements along the Schlickerbach Torrent.

In recent years the Kalkkögel has seen an enormous expansion of tourism on account of two Olympic Games staged partly in the Axamer Lizum. Systematic reforestation and terracing of areas such as the Grosse Blaike, carried out since the 1950s, construction of check dams and modern protective dams (Fig. 165), channel linings, and stone-masonry revetments have substantially reduced the hazard on the cones. New clearing for ski runs in the uplands and development pressure on the cones have justified considerable expense to assure safety for inhabitants and visitors.

Surlej (A66)

Location: Lake Silvaplana, St. Moritz, Graubünden (Grisons), Switzerland (E2)
Date(s): 1793

Lake Silvaplana is one of a string of sparkling lakes in the high Engadin Valley, a tourist region blessed with stunning vistas and a pleasant winter climate. The communities of Silvaplana and Surlej have developed on two delta cones on opposite sides of the lake (Fig. 166).

Upland basins in the area are devoid of trees and torrents flow from rugged slopes underlain by Austroalpine metamorphic and sedimentary rocks, in places covered by thick morainal deposits.

Surlej (1800 m) used to be a small agricultural hamlet on the gently northwest-sloping delta cone of a small creek draining the slopes of the Rosatsch Massif (3123 m). In 1793 a huge debris slump on the slopes behind the hamlet changed into a massive flow that completely devastated the settlement (Montandon, 1933, p. 310). The settlement was later rebuilt along the fringes of the debris which is still known as Crap da Sass (Crap = debris).

In recent years the entire upper cone of Surlej has been built over with vacation homes and apartments, part of the overall development of the St. Moritz region (Fig. 167). Only the lowermost part of the debris fan is still in agricultural use.

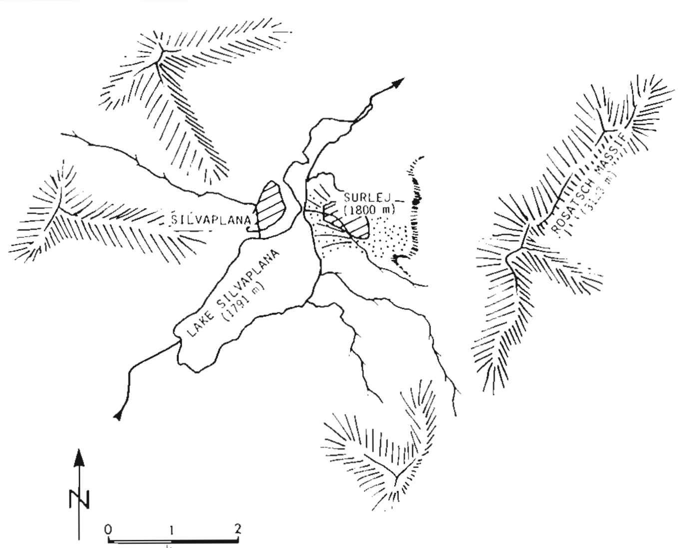


Figure 166: Sketch map of Lake Silvaplana and the two fan deltas on which Surlej and Silvaplana are located.

Figure 167: View towards the recently built-over segments of the Surlej fan delta. (GSC 204167-V)

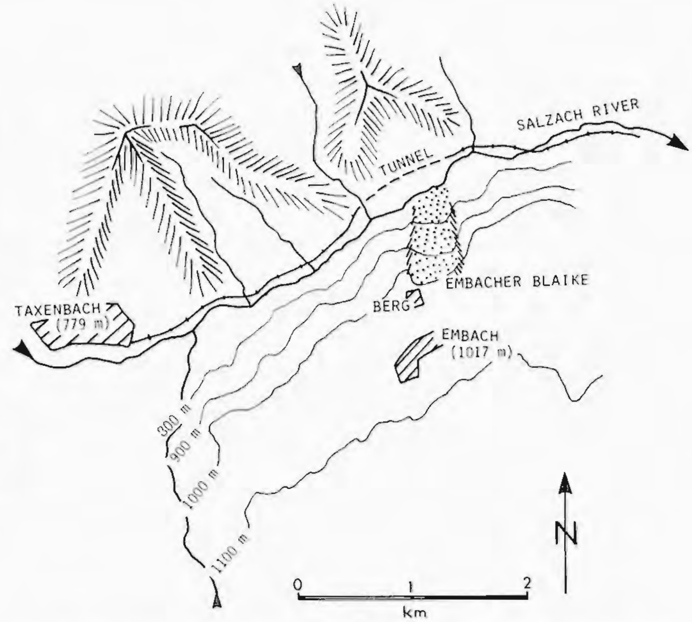


Figure 168: Sketch map of the Salzach River gorge east of Taxenbach, the gravel terrace of Embach, and the slide-prone phyllite slope north of the river. Rail and road beds skirt unstable terrain along the bottom of the gorge.

Figure 169: View of the Embacher Blaike scar; note rubbly toe zone and head scarp of the slowly moving debris slide which is covered by several generations of planted trees. (GSC 204167-W)

Embach (A67)

Location: Salzach Valley, Salzburg, Austria (11)
Date(s): Spring 1794 (also 5 August 1598, June 1874)

Near the community of Embach the Salzach River winds easterly through a narrow gorge cut into intensely deformed phyllitic bedrock of the Austroalpine basement complex and a terrace of Pleistocene surficial deposits (Fig. 168). Tributary basins of the Salzach River along this stretch of the valley are susceptible to sporadic slumping and debris flows. One of the largest areas of chronic slumping is the crown of the terrace of Embach (1017 m), which is approximately 350 m above the river. The Embacher Blaike (blaike = scar) is about 500 m wide and at times has involved as much as $10 \times 10^6 m^3$ of debris moving away from the concave headwall of the sand-gravel terrace.

On 5 August 1598, acceleration of debris movements caused retrogression of the head scarp; one of the opening cracks 'swallowed' a homestead and a family of seven. In the spring of 1794 another section of the head scarp failed, destroying fields and buildings in the hamlet of Berg which is perched on the edge of the terrace. This time a debris lobe, composed of gravel, clay, moraine, sand and phyllitic bedrock blocked the river and created a lake which extended 3 km upstream, as far as the village of Taxenbach. The lake persisted for several years before the river carved a new channel across the toe of the slide mass (Stini, 1938, p. 15 and 20).

Because of the large size of the area susceptible to sliding and the abundance of subsurface water, the unstable terrace at Embach has remained a serious problem to this day. Drainage by surface ditches and attempts at reforestation have only been partly successful. From time to time (e.g. 1935, 1948) these works are distorted by accelerated movements of the whole slope and have to be renewed. The road and railway along the busy transportation corridor were forced onto the unstable bedrock slope along the north side of the river (Fig. 169). There, as early as June 1874, the roof and wall of a railroad tunnel collapsed into the river (Wolf, 1875). The design of the modern highway has required extensive rock engineering to overcome the problems of the sagging terrain.

Niedernsill (A68)

Location: Salzach (Pinzgau) Valley, Salzburg, Austria (H1)
Date(s): 5 August 1798 (also 7 August 1970, 28 July 1971)

Niedernsill (796 m) is one of several communities that have grown on debris cones in the flood-prone and swampy Salzach Valley (Fig. 170). Tributary torrents of the Salzach River enter the valley from catchment basins carved into east-trending low grade metamorphic rocks of the Austroalpine basement complex (north of the river) and the Pennine Tauern Window (south of the river). The Mühlbach Torrent which has built the cone of Niedernsill receives most of its debris from unstable embankments of Pleistocene colluvium. The debris cone has pushed the channel of the Salzach River onto the north side of the valley and created a swampy floodplain on its upstream side. The surface of the cone hosts several settlements, including Niedernsill, Jesdorf, and Mühlbach.

On 5 August 1798, a cloudburst over the upland ridges in this area caused enormous runoff and subsequent erosion along the banks of the Mühlbach Torrent and mobilized an incipient debris slump at the mouth of the Bombach ravine (Fig. 170). This slide mass swerved across the swollen waters of the Mühlbach Torrent creating a lake 20 to 30 m deep.

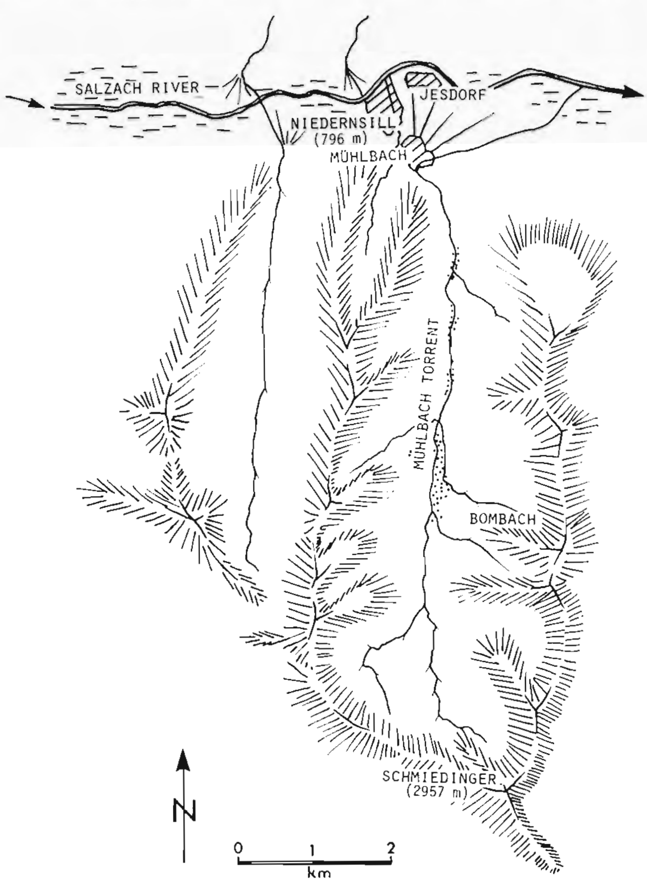


Figure 170: Sketch map of the Mühlbach basin south of the Salzach Valley; note low-gradient swampy floodplain between debris cones of tributary torrents.

Figure 171: View of Mühlbach (foreground) and Niedernsill (background); note the sharply curved protective debris diversion dam and continuing construction activity on the upper fan. (GSC 204167-X)

The oversaturated plug of debris failed and a massive flow completely ravaged the village of Mühlbach at the apex of the cone and invaded sections of Jesdorf and Niedernsill; six persons lost their lives. The debris stemmed the flow of the Salzach River and left a braided blanket of gravel on formerly productive agricultural land downstream from the cone. During this catastrophe more than $6 \times 10^6 m^3$ of debris may have been moved across the cone.

On 7 August 1970, a short and intense thunderstorm deluged the Mühlbach basin; more than 60 mm of rain fell in only 40 minutes. Reactivation of long-dormant erosional scars and local blockage of the torrent channel by log jams produced four debris flows. The wings of pre-existing check dams were torn off as the flows tried to establish a deeper and wider discharge section. A volume of $0.5 \times 10^6 m^3$ of blocky debris flared over the eastern sector of the cone. The frontal wall of debris contained blocks several metres in diameter and locally attained speeds of 50 to 60 km/h. Houses in Mühlbach were severely damaged and three people lost their lives. One building, picked up by a pulse of debris, was carried 500 m down the cone, where its occupant walked away unhurt! This catastrophe spurred the rapid construction of a masonry deflection dam at the apex of the cone.

Almost exactly one year later, on 28 July 1971, thunderstorms again struck the upland basin of the Mühlbach Torrent. This time there was 40 mm of rain in 90 minutes. Debris totalling $0.3 \times 10^6 m^3$ swept across the fan, causing heavy material damage. After this disaster the crest of the deflection dam which had guided only part of the flow was raised to a height of 10 m (Fig. 171). Control works (check dams) in the source area and a large-capacity retention basin at the apex of the fan have been added since then (Hoffmann, 1972).

Lake Lucerne — Vierwaldstättersee (A69)

Location: Uri-Schwyz-Unterwalden, Switzerland (D2)
Date(s): 14 May 1801 (also 1659, 23 July 1764, 8 December 1769, 15 July 1795, December 1879, 8 August 1964)

Lake Lucerne (435 m) occupies a deep fracture-controlled glacial basin carved into the front ranges of the Swiss Alps (Fig. 172). On the north side it is bordered by the conglomeratic Rigi Massif (1801 m) of the Molasse Zone; its southern arm (Urnersee) is walled in by near-vertical carbonate cliffs of the Helvetic cover complex that attain heights of more than 2000 m. Bedding surfaces and major thrust faults dip approximately 20° to the southeast. Communities along the lake are located on small fan deltas and narrow terraces. In the past human works have suffered both directly and indirectly from debris flows and rockfalls into the lake.

In 1659 a massive failure along the conglomeratic cliffs of the Rigi generated a blocky stream that destroyed the baths of Lützelau (Heim, 1932, p. 82; Montandon, 1933, p. 303). On 23 July 1764, a cloudburst deluged the summit ridges west of Lake Lucerne; water and debris collected on the deforested and deeply scarred limestone cliffs, and descended into a ravine that opens onto the steep debris cone with the settlement of Buochs. The rush of water and debris knocked down a dozen buildings and killed 11 people (Boffo, 1977). On 8 December 1769, a wedge of carbonate splintered away near Seelisberg and plunged directly into the lake, setting up a wave train that demolished shoreline installations nearby (Heim, 1932, p. 69).

In the spring of 1795 a series of cracks opened in a debris-covered slope above Weggis at the foot of the Rigi Massif. After a period of intense and sustained rainfall in July a slide mass, 1 km wide began to skid towards the lake, carrying the village of Weggis with it. Its overall movement was so slow that the inhabitants of the community still could salvage their belongings from the doomed homes. The motion of the slide was accompanied by a low rumble and continued for more than two weeks until the whole blocky lobe including the village disappeared into the lake. A new village soon was built near the site that had vanished into the rising waters of Lake Lucerne (Heim, 1932, p. 17; Montandon, 1933, p. 310).

On 14 May 1801, a mass of limestone, resting on calcareous slate and dipping steeply towards the lake below the Buggisgrat on the east side of Urnersee, became detached from its pedestal and plunged into the lake. The coherent rock mass set up a large wave train which rolled over shoreline installations, smashed several buildings and drowned 14 people (Montandon, 1933, p. 312).

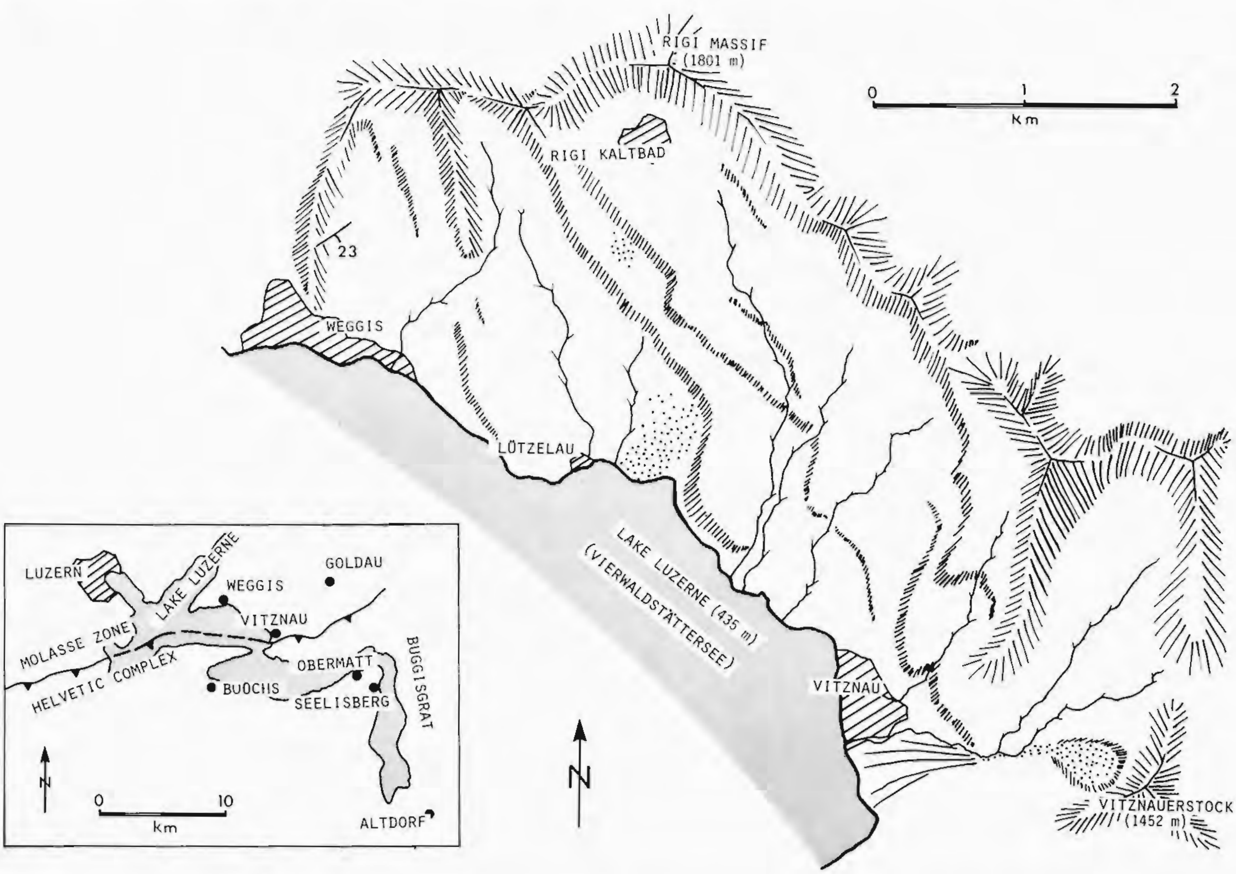


Figure 172: Index map of Lake Lucerne (Luzerne) and sketch map of the debris-covered conglomeratic dip slope of the Rigi Massif.

Late in December 1879 an unstable ledge on the Vitznauerstock (1452 m) failed; the blocky debris of the disintegrating slab temporarily halted on a terrace of calcareous flysch below the summit ridge. However, the terrace with its load of carbonate blocks began to move slowly into the headwaters of the Vitznauer Torrent; there the debris accelerated into a flow which eventually swept across the southern sector of the cone of Vitznau. Fortunately only the northern sector was inhabited and there was no loss of life when the overloaded periphery of the delta cone slumped into the lake (Heim, 1921, p. 436).

On 8 August 1964, a carbonate slab of 40 000 m^3 volume came loose along a quarry face near Obermatt. It set off a wave train in the lake that caused minor damage.

In recent years the communities along Lake Lucerne have continued to thrive. A major through road to the Gotthard Pass which formerly clung to the rockfall-prone eastern shore of the Urnersee has been placed entirely in a tunnel.

Schesatobel (A70)

Location: Bludenz, Vorarlberg, Austria (E1)
Date(s): 1804 (also 1867, 1907, August 1966, May 1967)

The Schesatobel (1300 m), an amphitheatre-shaped upland bowl 700 m across, is probably one of the largest erosional features in the Alps produced by human carelessness (Fig. 173). It rises above the headwaters of the Schesa Torrent, a southern tributary of the Ill River (530 m) and is carved into a 200 m high late Pleistocene ice-margin terrace. The poorly sorted and unconsolidated bouldery material of this terrace rests on deformed evaporitic carbonates of the Austroalpine cover complex.

Prior to 1770 thick forests mantled the Schesa basin, and the torrent was an inconspicuous creek. In 1796 major clearcutting operations completely denuded the basin. The creekbed, which served as a haulage route during logging, and its tributary ravines, became an ever-growing source of debris. During a series of rainy years backward erosion and widening of tributary gullies undercut numerous unstable embankments along the terrace rim and in 1804 the first massive flows of bouldery debris reached the Ill River. In wet years mobilization of debris from the Schesatobel continued to be a major problem: by 1867 the cone of the torrent had pushed the channel of the Ill River against the low terrace of Nüziders, threatening the collapse of several buildings in that settlement.

Towards the end of the 19th century, the Schesa Torrent and Ill River became the target of systematic control works. However, the Schesatobel amphitheatre by now had all aspects of a badland: precariously unstable pinnacles of bouldery deposits alternated with deeply incised ravines. Control works below the Schesatobel rim proved to be more difficult to execute than expected. In order to raise the torrent channel and thus prevent bank scour the abutments of check dams had to be provided with especially wide wings (Fig. 174). Construction activity in the bowl itself was hazardous.

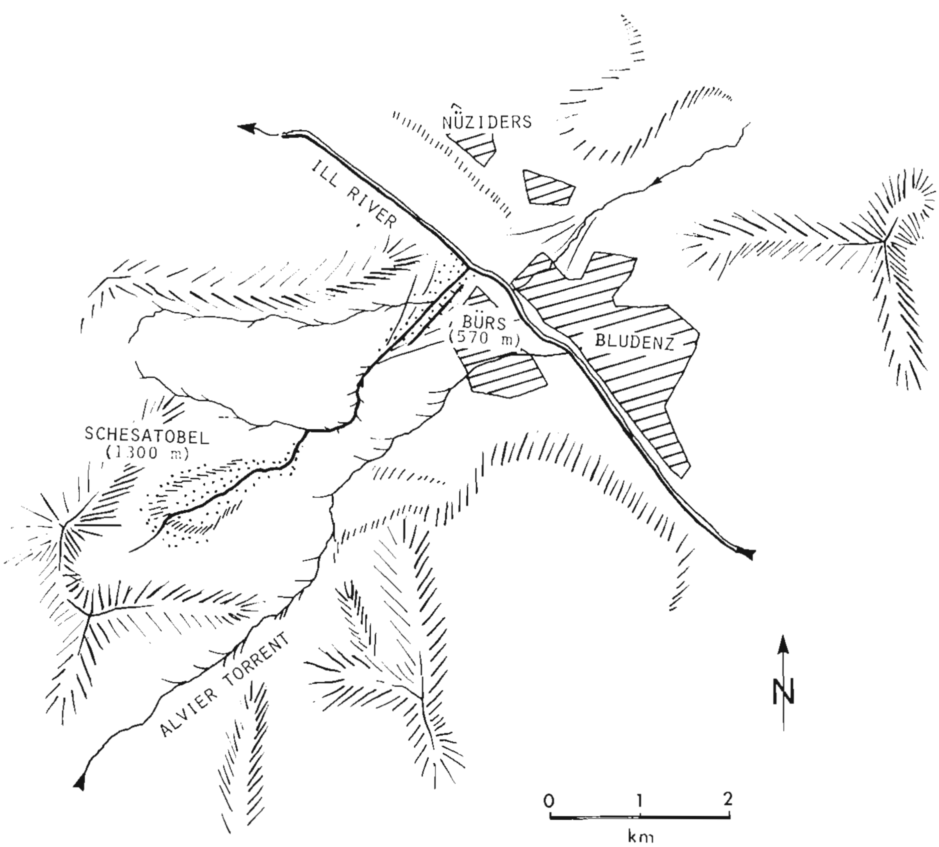


Figure 173: Sketch map of Schesa Torrent and the completely trained Ill River in the densely populated district of Bludenz.

Figure 174: Two views of the large check dams along the Schesa Torrent below the erosional bowl of the Schesatobel: a) check dams confining the braided channel of the Schesa Torrent to the centre of the valley thus preventing lateral scour (GSC 204167-Y); b) steel-concrete check dam with an armoured carapace which minimizes damage to the structure from minor debris flows (GSC 204167-Z).

In 1907 a wall of bouldery debris with a volume of $0.2 \times 10^6 m^3$ collapsed, almost killing a work gang. Only the foresight of a supervisor averted a major disaster. Gradually control of bank erosion and reforestation of bare slopes have reduced the amount of material carried to the Ill River. Nevertheless, local mockery referred to the erosional feature as 'Millionenloch' (=the million hole) in reference to the money poured into it.

Schesatobel continues to be a problem. A rainstorm in August 1966 and another in May 1967 mobilized debris with a total volume of $0.8 \times 10^6 m^3$ and forced a complete modernization of the check dams (Fiebiger, 1974). Existing and planned housing developments in Bürs, on the Schesa cone, have justified construction of a protective debris deflection dam and expenditures for continued control work (Fig. 174).

In the 180 years of its existence the Schesatobel has produced about $40 \times 10^6 m^3$ of debris. Check dams erected during the last 80 years have raised the bed of the torrent by approximately 40 m, allowing alders to invade some of the scarred embankments.

Goldau (A71)

Location: Schwyz, Switzerland (D1)
Date(s): 2 September 1806 (also 1354, August 1874)

The community of Goldau (500 m) is situated on a large apron of angular debris which envelops the southern flank of the Rossberg or Rüfiberg Massif (1570 m). The Rossberg and its famous counterpart across the valley, the Rigi, are part of the Tertiary Molasse Zone of the Swiss Alps and are underlain by panels of resistant conglomerate and recessive shale, dipping to the south-southeast (Fig. 175). Near the toe of the Rossberg dip slope bedding dips 15°, near the top of the mountain the dip increases locally to 30°. Transverse joint sets disrupt the Rossberg slope into large tilted slabs. A large part of the Rossberg south slope is covered by such dislodged blocks; some of them are remnants of prehistorical rockslides, others the result of minor historical rockfalls; however, most of them are remnants of the great 'Goldauer Bergsturz' of 1806.

The earliest known historical failure of the Rossberg dip slope occurred about 1354. This slide probably obliterated the hamlet of Röten, then located in a zone that was to fail again during the great catastrophe of 1806.

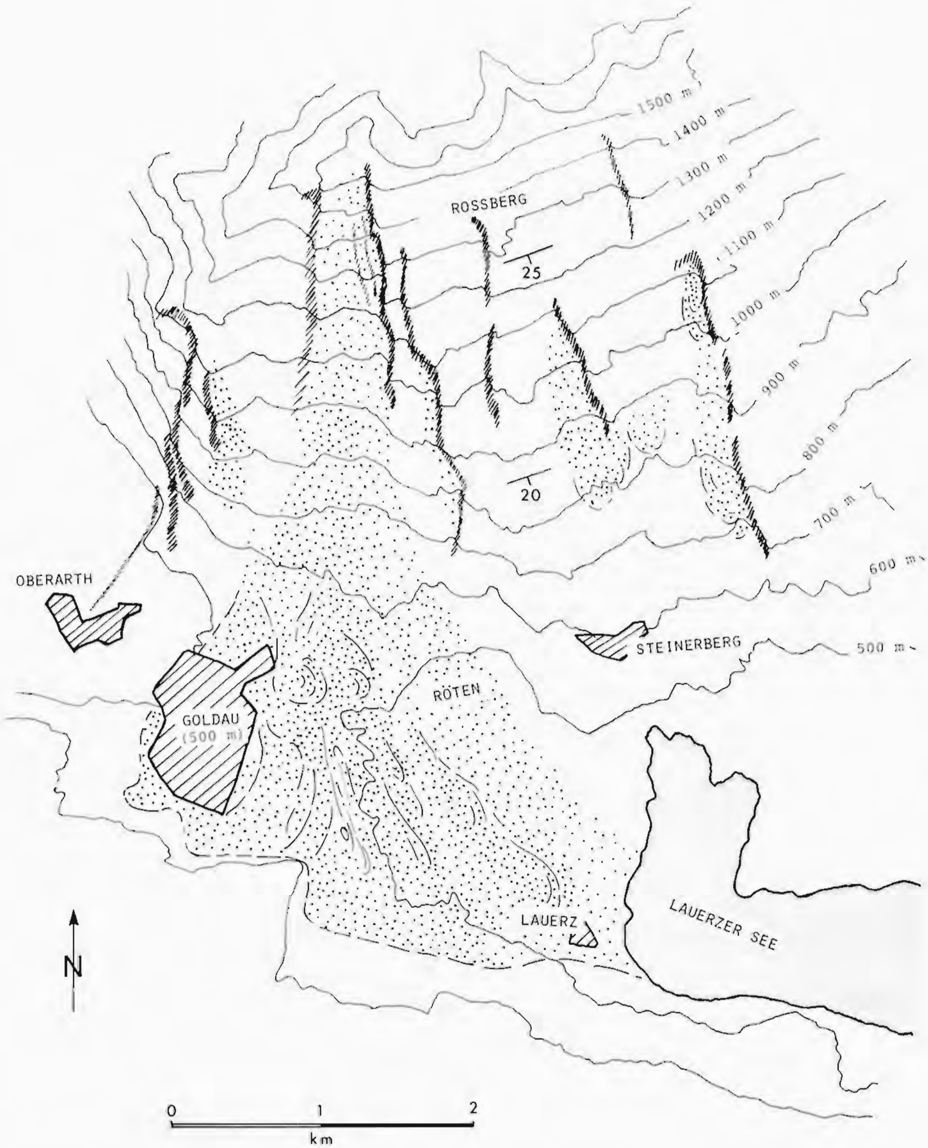


Figure 175: Sketch map of the Goldau rockslide; note minor incipient debris slumps along ledges of the conglomerate dip slope.

The events leading up to the destruction of Goldau in 1806 were chronicled by Zay (1807) in an extremely important monograph; supplemental studies were conducted later by Heim (1921, 1932), Kopp (1936), Lehmann (1942) and Zehnder (1974). Incipient instabilities near the crown of the Rossberg Massif had raised some apprehension at least thirty years prior to the 1806 failure. Movement of part of the slope became obvious during the extremely rainy years of 1799, 1804, and 1805. In 1805 debris flows and bedrock slumps occurred widely throughout eastern Switzerland. Then the winter of 1805-1806 brought an exceptional snowpack and runoff was delayed by a cold spring. The period of delayed, but rapid snowmelt was followed by heavy precipitation in July and August. At this time the upper part of the Rossberg Massif conglomerate walls collapsed into adjacent ravines and trees toppled into open fissures. Vertical fractures, now filled with water, cracked open emitting sounds akin to the discharge of cannons. Some of the people living on the Rossberg packed up and left. Others, including the inhabitants of Goldau, stayed, fascinated by the drama that unfolded above their heads.

On the afternoon of 2 September 1806, the main head scarp of the slide opened along a single fracture and a rock mass of approximately 10 to 20 $\times$ $10^6 m^3$ began to move, first slowly then explosively fast, along a composite bedding plane contact between conglomerate and calcareous shale. Earth tremors were felt throughout the incipient slide area and the rising toe of the disintegrating rock mass instantaneously ploughed into the Lauerzer See where it set off a wave of mud, trees, and water, 20 m high. This wave surged over the lakeside villages of Busingen and Lauerz, and debris filled one seventh of the lake. Destruction was even more violent in Goldau along the western arm of the slide mass; an air blast preceding the frontal wall of the rock avalanche lifted a few lucky souls to safety, but the main body of debris annihilated about 300 buildings and claimed

Figure 176: Reproduction of a watercolor dramatizing the destructive impact of the Goldau rock avalanche on the communities in the valley (original watercolor by D.A. Schmid, sold as a poster at the Bergsturz Museum in Goldau).

457 lives. The terrible roar of the avalanche was followed by a deathly silence and a thick cloud of dust drifted across the valley (Fig. 176).

In spite of the generally turbulent times in Europe, the catastrophe attracted wide attention, and the site of the disaster was visited by many travellers, artists and writers. Lord Byron, who came to Goldau ten years later wrote about the Rossberg slide. . .

'Mountains have fallen
Leaving a cap in the clouds and with the shock
rocking their Alpine brethren, filling up
The ripe green valleys with destruction's splinters,
Damming the rivers with a sudden dash,
Which crushed the waters into mist, and made
their fountains find another channel — thus,
Thus, in its old age, did Mount Rosenberg —'. . . .

The 1806 failure of Rossberg was followed by smaller recurrent slides. At the end of August 1874, a small rock-debris avalanche cascaded down the southwest flank of the mountain, causing minor damage at a locality known as Sonnenberg above Oberarth; the failure was triggered by infiltration of water from two springs into open fractures of deeply weathered bedrock and surficial debris (Baltzer, 1875). During the wet summer of 1910, considerable slippage of old slide material was also noticed along the eastern flank of the mountain by Heim (1932, p. 40 and 55).

Reoccupation of the land devastated by the 1806 rockslide and re-establishment of a community were slow at first. However, the arrival of the Rigi and St. Gotthard railways in the late 19th century turned Goldau into an important transfer centre. Nevertheless, the idea of a railroad tunnel beneath the

Figure 177: View northward from the town of Goldau to the detachment zone of the 1806 rockslide; note large transported blocks in the foreground. (GSC 204168-A)

slide masses along the Rossberg toe zone was abandoned because . . .

'those parts of the Rossberg, which according to the views of geologists might cause similar catastrophes in time spans of several hundreds of years, overhang the southern entrance. . .' (Zehnder, 1974, p. 144).

Today Goldau flourishes on account of its industry and tourism. An excellent small 'Bergsturzmuseum' keeps alive the memory of the disaster which so deeply affected the land and people of this beautiful part of Switzerland. The detachment zone of the rockslide has been set aside as a nature conservancy area (Fig. 177). The eastern upland slope of the Rossberg is still dotted by numerous hamlets and farms.

Schwaz (A72)

Location: Inn Valley, Tirol, Austria (G1)
Date(s): 26 July 1807 (also 600 A.D., February 1527, 1539 (?), 1553, 1569, July (?) 1669)

The town of Schwaz (545 m) owes its origin to mining activities dating back to prehistorical times. The main mining boom, based on silver and copper ores found southeast of the city, lasted from about 1500 to 1700. During this period Schwaz grew rapidly, spreading over most of the cone of the Lahnbach Torrent (Lahn = avalanche), a southern tributary of the Inn River (Fig. 178). This torrent drains a bowl-shaped basin on the northwest side of the Kellerjoch (2200 m) which is underlain by gneissic bedrock of the Austroalpine basement complex. Discontinuous pockets of late Pleistocene ice-margin deposits perched against bedrock between 1000 and 1500 m are dissected by tributary ravines. Below the debris sources the torrent passes through a narrow gorge with phyllitic bedrock embankments. Historical debris flows in the Lahnbach basin have resulted from embankment failures, blockage of the gorge, and instabilities along deforested colluvial veneers.

The first 'horrible' bursts of debris over Schwaz occurred about 600 A.D. (Hanausek, 1975, p. 108-109). Devastating flows engulfed the thriving mining centre again in February 1527, in 1539 (?), in 1553, in 1569 when 140 people were said to have been killed, and in July (?) 1669 when 50 people perished (Dalla Torre, 1913, p. 103).

On 26 July 1807, a thunderstorm-cloudburst centred over the Kellerjoch released large amounts of debris from the ravines of the upland basin; debris flows flattened 11 buildings in Schwaz, damaged many others, and killed four people (Aretin, 1808, p. 79-80).

For centuries the town has attempted to protect itself against the ravages of the torrent by a system of stone-masonry dykes designed to guide debris straight through the city to the Inn River. Repeated spillover of debris during storms raised the level of the channel and dykes several metres above the town's streets. Today the bed of the torrent follows the elevated axial ridge of the cone, fringed by a public park. Above the apex of the cone two steelbeam-concrete dams, approximately 10 m high, provide protection against exceptional mass movements. The principal debris sources of the uplands have been neutralized with check dams and snow avalanche bridges. Within the city this increased level of control has permitted building activity even along the slanting outer flanks of the old stone-masonry dykes, and the debris cone now is completely built over (Fig. 179).

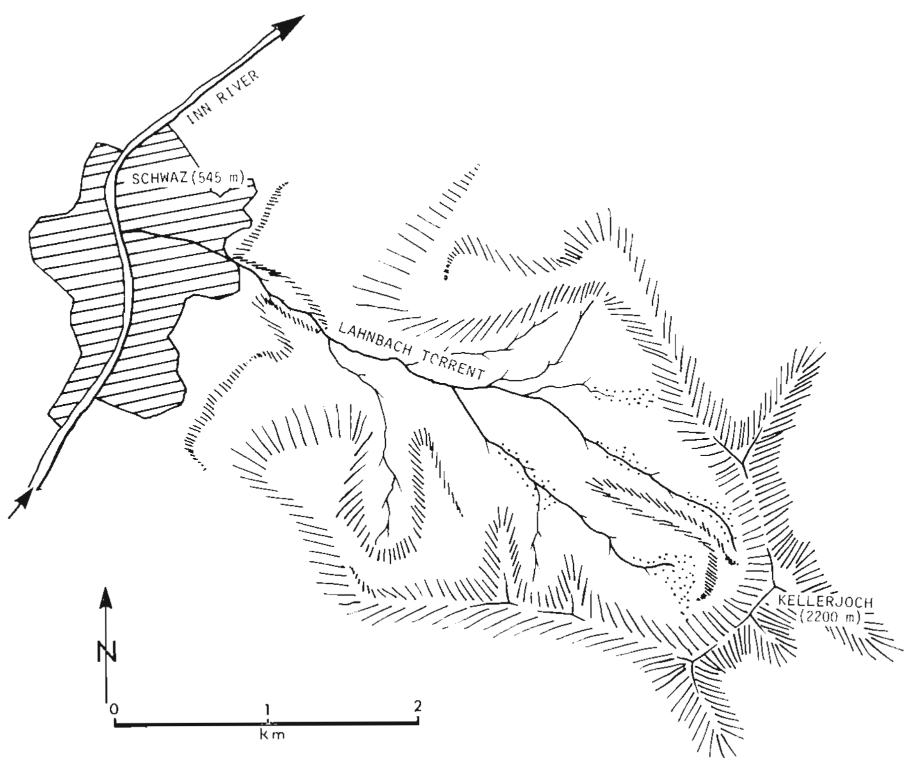


Figure 178: Sketch map of the old mining town of Schwaz on the cone of the Lahnbach Torrent.

Figure 179: View of the built-over cone and the catchment basin of the Lahnbach Torrent; note farms along upland terraces, protective forests along the gorges, and snow avalanche tracks below the crest of the basin. (GSC 204168-B)

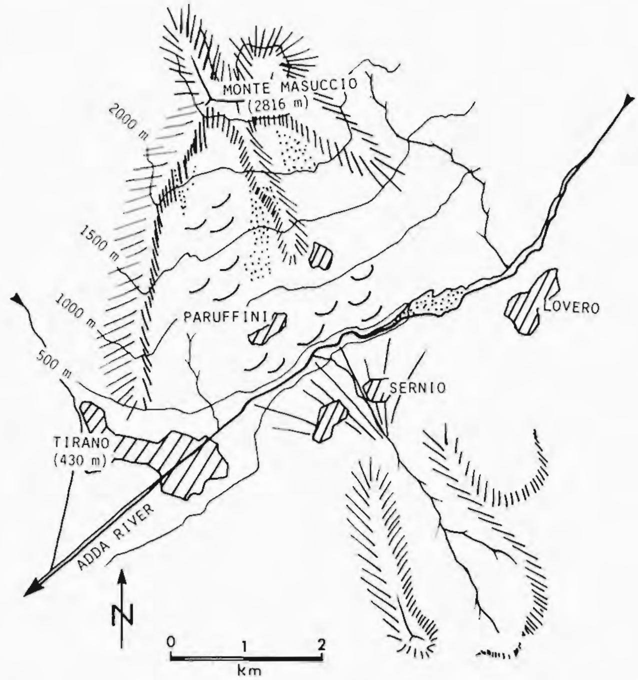


Figure 180: Sketch map of the sagging Monte Masuccio bedrock slope northeast of Tirano.

Monte Masuccio (A73)

Location: Tirano, Valtelline (Adda Valley), Lombardia, Italy (F2)
Date(s): 7 December 1807 (also June 1808)

The massif of Monte Masuccio (2816 m) rises on the north of the Adda River (500 m). Like most other south-facing slopes of the Valtelline, Monte Masuccio hosts several small hamlets and vineyards, which cling to a sagging complex of steeply dipping low grade metamorphic schists of the Austroalpine basement complex covered by pockets of surficial debris (Fig. 180). The toe zone of this sagging slope has been undercut by the Adda River, which, at precisely this point, is forced northward by a large debris cone.

On 7 December 1807, movements along the toe of the sagging slope accelerated, and a large slump mass crossed the Adda River, obliterating the mills of Sernio. These movements probably were triggered by infiltration of water into cracks during the periods of exceptional regional rainfall registered in the months and years preceding the failure.

The slide mass obstructed the flow of the Adda, creating a lake which drowned several small settlements on the floodplain upstream from Sernio. Six months later, in June 1808, the overflow of the lake, possibly enhanced by springtime freshets in the upper basin of the Adda, ruptured the slide dam and a debris flood laid waste to the valley below, including parts of the town of Tirano (Schaubach, 1867, v. 4, p. 78; Böhm, 1886, p. 634).

Today erosion-control structures inhibit undercutting of the Adda River along the toe zone of the sagging Monte Masuccio slope; high masonry walls and dykes guide the boulder-strewn river through the town of Tirano.

Antelao (A74)

Location: Cadore, Bòite Valley, Veneto, Italy (H2)
Date(s): 21 April 1814 (also 25 January (?) 1348, 1729, 1737, 1868)

The Antelao Massif (3263 m) on the northeast side of the Cadore Valley is a steep-walled carbonate pyramid in the South Alpine cover complex (Fig. 181). Below, shattered spurs, towers and ragged ravines open over steep cones of blocky debris.

The upper part of Antelao is composed of thick-bedded dolomite resting on a recessive shale-sandstone unit. The entire panel of rocks dips 25° to the north-northwest. The recessive unit (Raibliano Formation) defines a slightly inward-dipping terrace along the west face of the mountain. The communities of Borca di Cadore (945 m), Villanova, and Cancia (892 m) occupy the slopes between the mountains and the Bòite Torrent. The town of San Vito di Cadore covers the lower portion of a huge debris cone, which probably originated by prehistorical mass movements along the north face of Antelao.

On 25 January (?) 1348, possibly due to a great earthquake that shook the southern Alps, a rock avalanche broke away from the upper cliffs of Antelao and seriously damaged some buildings in Borca di Cadore. Small rockfalls were again registered in 1729 and 1737; these may have been precursors of the great collapse along the northwest face of the mountain in 1814.

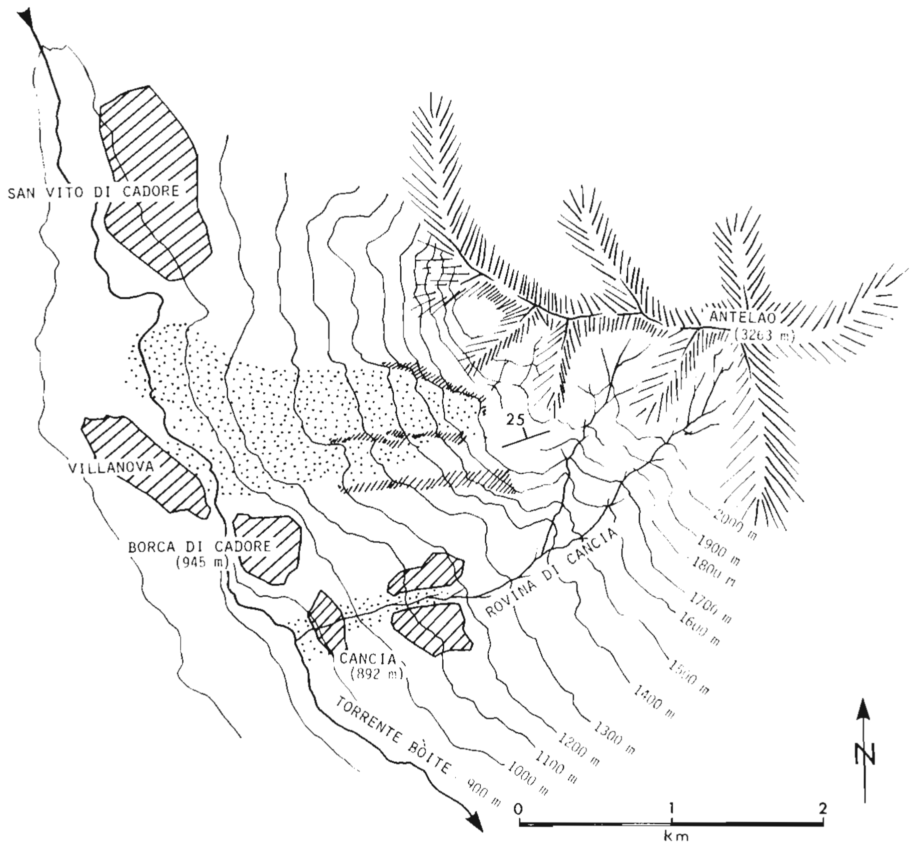


Figure 181: Sketch map of the carbonate pyramid of Antelao and the tracks of historical rock avalanches.

On 21 April 1814, a mass of carbonate rocks approximately $5 \times 10^6 m^3$ in volume slid away above the Raibliano shale horizon. Guided by the northwest-dipping bedding and a steep northwest-trending fracture set, the disintegrating mass of dolomite first moved north-northwesterly and then cascaded westerly from the slanting terrace onto the valley floor (Fig. 182). There the lobe split into two parts: one erased the villages of Taolen and Marceana, killing 300 people; the other blocked the south-flowing Bòite Torrent temporarily (Catullo, 1814; Casal, 1898). The buried villages were not rebuilt after the catastrophe.

In 1868, a year characterized by abnormally deep snowpacks and severe rainstorms throughout the Alps, another segment of Antelao failed. This time, a spur below the summit of the mountain split along a vertical crack and a stream of carbonate rubble swept down the Rovina di Cancia. The rubble cut a swath through the village of Cancia, killing 12 people (Marinelli, 1878, p. 34).

Because of their proximity to the tourist centre of Cortina d'Ampezzo the communities of Villanova, Borca di Cadore, and Cancia presently are experiencing rapid growth. New vacation homes are located on deposits of the 1814 rock avalanche and on the slanting apron of debris fringing the Rovina di Cancia.

Figure 182: Part of the talus-covered basal rupture zone of the Antelao rock avalanche of 1814. (GSC 204168-C)

Perarolo (A75)

Location: Piave Valley, Ceneto, Italy (H2)
Date(s): October 1820

Perarolo (528 m), a small town in the upper Piave Valley, is located on a small bouldery triangle marking the confluence of the Bòite Torrent and the Piave River. Upstream from the town the Bòite Torrent crosses a gorge cut into fractured carbonate shale and evaporites of the South Alpine cover complex.

In October 1820 a part of the evaporitic-calcareous bedrock slope of the Bòite gorge failed and a thick lobe of debris temporarily blocked the torrent. When the barrier broke a massive debris flow overwhelmed Perarolo, depositing as much as 12 m of rubble among the houses, rendering them useless (Casal, 1898, p. 214).

After the catastrophe, the town huddled against the steep western wall of the gorge. Today a dyke, approximately 10 m high, protects the village from the ravages of the torrent (Fig. 183). Two hydroelectric reservoirs in the upper part of the Bòite basin control seasonal discharge. A railroad tunnel also pierces the unstable formations of the gorge.

Figure 183: Masonry and gabion dyke separating the town of Perarolo from the channel of the Bòite Torrent; unstable bedrock slopes upstream from Perarolo in the background. (GSC 204168-D)

Val Badia — Gader Valley (A76)

Location: Dolomite Mountains, Südtirol (Alto Adige), Italy (G2)
Date(s): 19 June 1821 (also 1488 to 1493, June 1827, 1 to 8 November 1841, 14 April 1867)

The north-flowing Gader Torrent drains a large catchment area of subdued shale slopes crowned by rugged carbonate plateaux of the northern Dolomite Mountains (South Alpine cover complex); the crests of the Gader basin attain heights of more than 3000 m (Fig. 184). Extensive veneers of surficial deposits characterize most of the lower shale slopes of the open valleys. In strong contrast, the carbonate successions above weather in vertical walls and towers. The presence of swelling clays in the shales and volcaniclastics favours development of slumps, incipient slide scarps, and extensive earthflows. Mobile shale slopes involve broken bedrock and surficial deposits up to 50 m thick. Displaced fences and road beds suggest long-term movement rates of 10 to 80 cm/year. However, these rates vary greatly from year to year depending on the quantity of water entering the slopes during snowmelt and seasonal storms (Fischer, 1967). The outflow of shale from beneath the plateaux leads to sporadic collapse of carbonate pinnacles. Many of the upper shale slopes are covered by angular carbonate blocks which local legends used to attribute to battles between giants! Chaotic aprons of blocky carbonate debris also are mobilized occasionally into devastating debris flows.

On 19 June 1821, probably prepared by infiltration of snowmelt from heavy snowpacks and by extended periods of summer rain in the region during the preceding decade, creep and slumping involving most of the mountainside below the

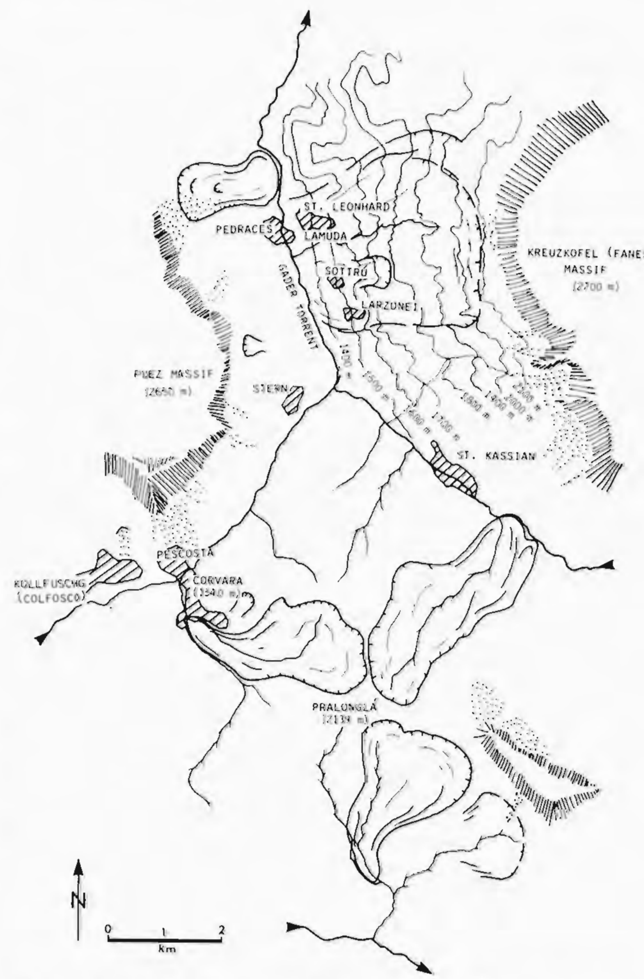


Figure 184: Index map of the upper Val Badia. Note the large earthflow lobes on the gentle shale slopes below the carbonate plateau.

Figure 185: The unstable shale slope of Lamuda below the carbonate cliffs of the Kreuzkofel (Fanes) Massif. The toe of this slope impounded the Gader Torrent in 1821. (GSC 204168-E)

west wall of the Kreuzkofel (Fanes) Massif (2700 m) were noticed southwest of St. Leonhard (Fig. 185). Large cracks soon opened in a broad arc above the hamlets of Lazonei and Sottrú and an earth-debris mass pushed 16 buildings of Lamuda into the rising waters of the Gader Torrent whose flow had been stemmed by the advancing toe zone of the moving slope. A lake known as Lec da Sompunt eventually extended more than 1 km upstream before being drained by a cut across the debris barrier. However, the embankments along the former lake were now extremely unstable and during a sudden snowmelt and rainstorms in June 1827 the Gader Torrent carried away much of the loose slide material, devastating fields and communities in the lower valley (Staffler, 1846, v. 2, p. 280).

During the wet period of the early 19th century other shale slopes in the Dolomite Mountains were in motion. Between 1 and 8 November 1841, a large earthflow advanced towards the town of Cortina d'Ámpezzo (10 km east of Val Badia) demolishing 24 buildings in the hamlet of Pecol. On 14 April 1867, a gigantic earthflow near Pescosta threatened the hamlet with a similar fate.

In the past earthflows on poorly drained shale slopes of the Dolomite Mountains were accepted as a nuisance by the inhabitants of traditionally agricultural settlements in the region. Recently, the Dolomite Mountain region has experienced an unprecedented tourist boom: roads, ski lifts, and residences have invaded formerly open shale slopes (e.g. the Corvara-Pescosta area). Surface drainage is pursued vigorously as the only economically feasible remedial measure, but is often inadequate to completely stabilize creep of the impermeable shale terrain.

Valle di Vanoi (A77)

Location: Valle di Vanoi, Trentino, Italy (G2)
Date(s): December 1825 (also 14 August 1748, autumn 1823, May 1826, 20 September 1829, September 1882, 12 October 1889, 4 November 1966)

The large drainage basin of the south-flowing Vanoi Torrent is underlain by phyllite, schists, and granitic rocks of the South Alpine basement complex (Fig. 186). The basin is bordered on the north by a chain of rhyolite peaks, the Cadena dei Lagorai (2500 m). Much of the large catchment area is underlain by sagging mountainsides of phyllite and pockets of late Pleistocene ice-margin deposits. After reckless deforestation during the 18th century deep cracks and gullying appeared in several tributary basins, initiating a remarkable series of mass movements affecting the length of the Vanoi Valley.

On 14 August 1748, a violent regional rainstorm released an enormous quantity of water onto loose debris near the summit of Cima Folga. The material was mobilized into a coherent debris flow that overwhelmed Canale di Sotto (near the present Canal San Bovo), killing 72 persons (Gorfer, 1977, p. 962).

However, the most spectacular debris movements in the region occurred in the Val del Rebrutt, a small tributary basin west of the Vanoi Torrent. The crest of this basin includes Monte Calmandrino (2079 m) and is underlain by phyllitic bedrock which formerly was covered by a thick blanket of surficial deposits. In 1786, after the basin had been completely denuded of its forest cover, large crown cracks appeared along the upper rim of a north-facing dip slope east of Monte Calmandrino. Landowners recognized that excessive snowmelt and rainstorms had triggered the ground movements and they tried to channel some of the runoff away from the open fissures. Unfortunately, their efforts came too late: gullying and subsidence had advanced to the stage where massive failure was imminent. Finally, in the autumn of 1823 a regional rainstorn set off the first series of debris flows from the Rebrutt basin; the flows crossed the Vanoi Torrent and stemmed its flow for more than an hour. Then the unstable debris barrier burst and a deluge of rock and mud demolished several buildings in Canale. After this, discharge of debris occurred ever more frequently and in December 1825 a newly formed cone across the bottom of the Valle di Vanoi began to impound a lake (Lago Nuovo) which soon drowned 36 houses comprising the hamlet of Ponte (Strele, 1899). In May 1826, during a spring freshet, a large section of the conical dam failed and a deluge of debris burst over the hamlet of Remesori near Canal San Bovo, erasing 13 buildings and killing 52 people (Gorfer, 1977, p. 962). New loads of debris from the Rebrutt basin raised the height of the cone to 90 m above the valley floor and increased the length of the lake to more than 2 km. Below, the Vanoi Torrent now shifted its braided track randomly from one bank to the other. On 20 September 1829, the lake again overflowed and cut deeply into the conical dam. The resulting debris flood, which reduced the lake to one third its size, toppled the church and several buildings in Canal San Bovo.

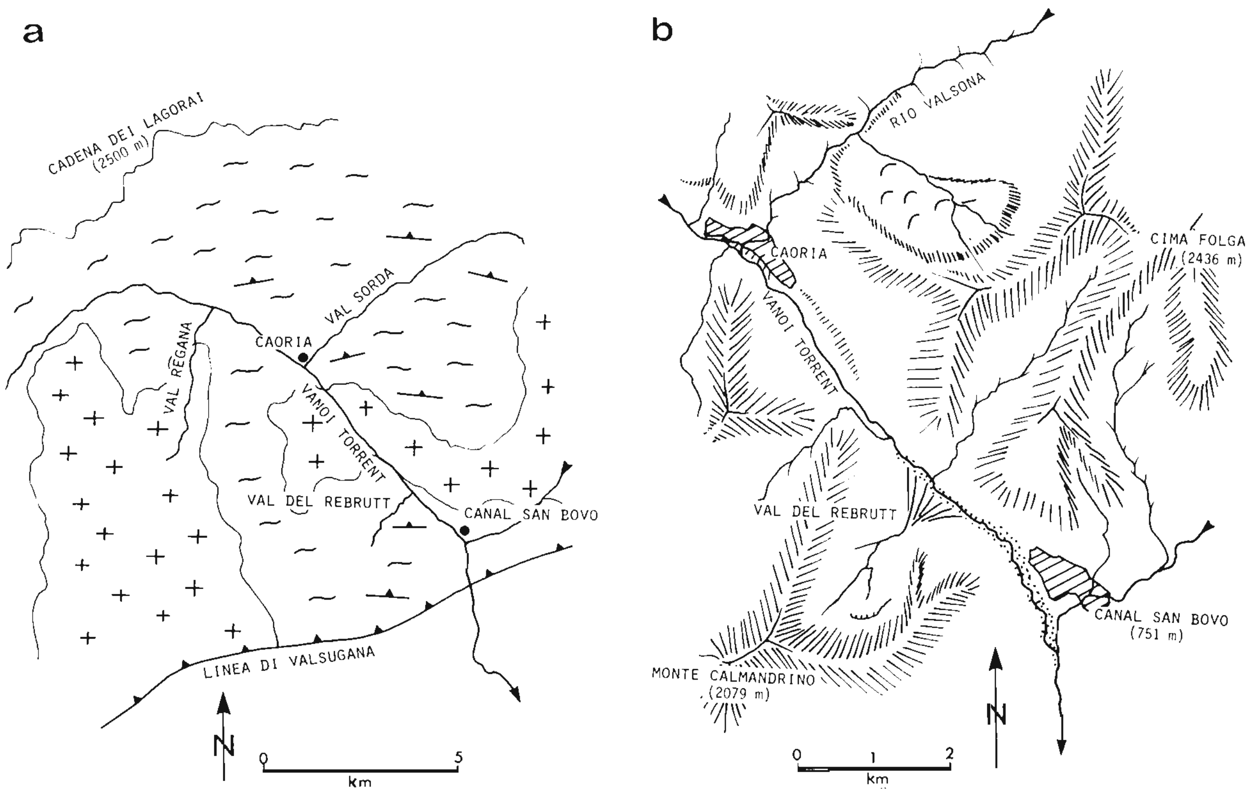


Figure 186: a) Geological index map of the upper Vanoi basin; wavy lines indicate phyllites, crosses indicate granitic rocks. b) Sketch map of the Vanoi Torrent between Caoria and Canal San Bovo; note the Rebrutt fan and sagging foliation dip slopes along Rio Valsona.

In subsequent years supply of available debris in the Rebrutt basin gradually decreased and vegetation reconquered the bleak slopes. Nevertheless the braided reach of the Vanoi Torrent near the Rebrutt cone and the remnants of Lago Nuovo remained a menace for years to come. Late in September 1882, on the heels of a great regional rainstorm, a debris flow from a steep tributary torrent (Val Regana ?) closed the Vanoi Gorge at a point 8 km above the village of Caoria. This bouldery barrier soon collapsed and a flow carrying gigantic granite blocks swept across the lower parts of Caoria into the swampy upper end of Lago Nuovo. The wave train set up by the impact of the debris forcefully struck the Rebrutt cone and lowered the outlet channel of the lake. On 12 October 1889, several debris flows occurred in the Vanoi basin. The most disastrous of these originated from slides in the basin of Rio Valsona. It slammed into buildings of Caoria and filled whatever remained of Lago Nuovo (Strele, 1936, p. 135-136). This last series of catastrophes accelerated the ongoing depopulation of the upper Vanoi region.

Figure 187: Northwestward view along the floodplain of the Vanoi Torrent at the toe of the Rebrutt cone (on the left); note wide-winged check dams designed to localize bedload transport and impede channel migration. (GSC 204168-F)

Figure 188: Recently constructed church on the cone of the Rio Valsona Torrent at Caoria; note the block of metamorphic rock marked with '1889', the date of the debris flow which deposited it. (GSC 204168-G)

During this century the most severe mass movements in the Vanoi region occurred during the great rainstorm on 4 November 1966, when, after four days of rain totalling 300 mm, dozens of buildings, roads, and bridges collapsed in the onslaught of debris avalanches, debris flows, and debris floods. After this event the regional government initiated a concerted effort to check debris movements in the uplands and to control erosion along the banks of the Vanoi (Fig. 187). Protective dams and control structures have been built in and near the communities of Caoria and Canal San Bovo where a considerable number of vacation homes have since sprung up (Fig. 188). Much of the bouldery torrent track along the former bottom of Lago Nuovo between Caoria and Canal San Bovo is now covered by a dense alder forest (Fig. 187). Forests again contribute to the stability of precarious bedrock slopes in the Rebrutt, Valsona and other tributary basins. Downstream from the Rebrutt cone two distinct terraces flanking the Vanoi channel are reminders of the two largest bursts of Lago Nuovo in the 19th century.

Obervellach — Möll Valley (A78)

Location: Kärnten (Carinthia), Austria (I2)
Date(s): 27 June 1828 (also 1619, 1620, 1717, 1757, May 1827, 1840-1854, 17 to 18 August 1966)

The Möll Valley (700 m) west of Obervellach follows the east-trending, south-dipping tectonic contact between mica schists of the Austroalpine basement complex on the south and medium- to low-grade metamorphic rocks of the Pennine Tauern Window on the north (Fig. 189). Mountain ranges flanking the valley attain elevations of approximately 2500 m and are the sources of torrents that commonly flow across sagging bedrock slopes or are carved into Pleistocene colluvium (Weiss, 1969). Prehistorical debris flows from unstable uplands created large cones along the sides of the Möll Valley, giving rise to the sinuous course of the Möll River. In historical time massive debris movements generally followed periods of deforestation or overgrazing of steep south-facing slopes.

In 1619, 1620, 1717 and 1757 the town of Obervellach at the head of the Kaponigbach cone was severely damaged by debris flows as much as 4 m deep. On 27 June 1828, a slide off the dip slope of the Steggraben ravine blocked the Kaponigbach gorge above Obervellach; the pressure of the impounded waters soon caused failure of the debris plug which changed into a massive flow that overwhelmed the town, burying 14 houses (Stini, 1938, p. 23).

At about the time of the last disaster at Obervellach the Klausenkofler Torrent began to deeply erode its unstable embankments. The torrent had been a harmless northwestern tributary of the Möll River until the beginning of the 19th century when a large clearcut denuded a bench of relict colluvium along an unstable foliation dip slope. In May 1827 warm rains and rapid melting of an exceptional snowpack set off the first voluminous debris flows through the steep gorge of the Klausenkofler Torrent into the Möll River. As the torrent joins the Möll River along a relatively narrow stretch of the valley, the flows of debris immediately impounded the river, creating a shallow lake, the Gössnitzsee. Sporadic floods resulting from bursts of the lake persisted for decades and became particularly troublesome during the wet years between 1840 and 1854 (Seckendorff, 1884, p. 202-205).

In later years the torrents of the region were brought under control by masonry check dams in the uplands and by walled embankments on the cones. The Gössnitzsee all but disappeared. Nevertheless, until recently the settlement pattern in the valley reflected the traditional knowledge of recurrent debris flows and flood waves: most villages cluster along south-facing interfan areas well above the floodplain of the Möll River. In the wake of large scale hydroelectric development of the Möll basin, the Gössnitzee was re-established by a dam. New housing developments and residential construction soon shifted onto more hazardous cones.

During the great rainstorm of 17-18 August 1966, most of the torrents of the Möll Valley carried very large loads of debris. One of the most destructive debris flows originated by the failure of a high embankment of relict colluvium in the upland basin of the Wollinitzbach Torrent. A lobe of boulders and uprooted trees accelerated as it descended down the gorge, attaining a volume of $0.3 \times 10^6 m^3$; the debris surged in several pulses over a new section of Kleindorf on the axis of the cone, demolishing eleven buildings and killing three people. The debris reached the Möll River, pushed its channel against the south bank and raised its bed 7 m. Only rapid excavation of a ditch across the debris flow deposit, which eventually attained a volume of $0.6 \times 10^6 m^3$, prevented potentially costly damage to hydroelectric installations upstream from the cone.

After the storm of 1966, check dams and revetments along many channels were supplemented by protective dams across the gorges (Fig. 190).

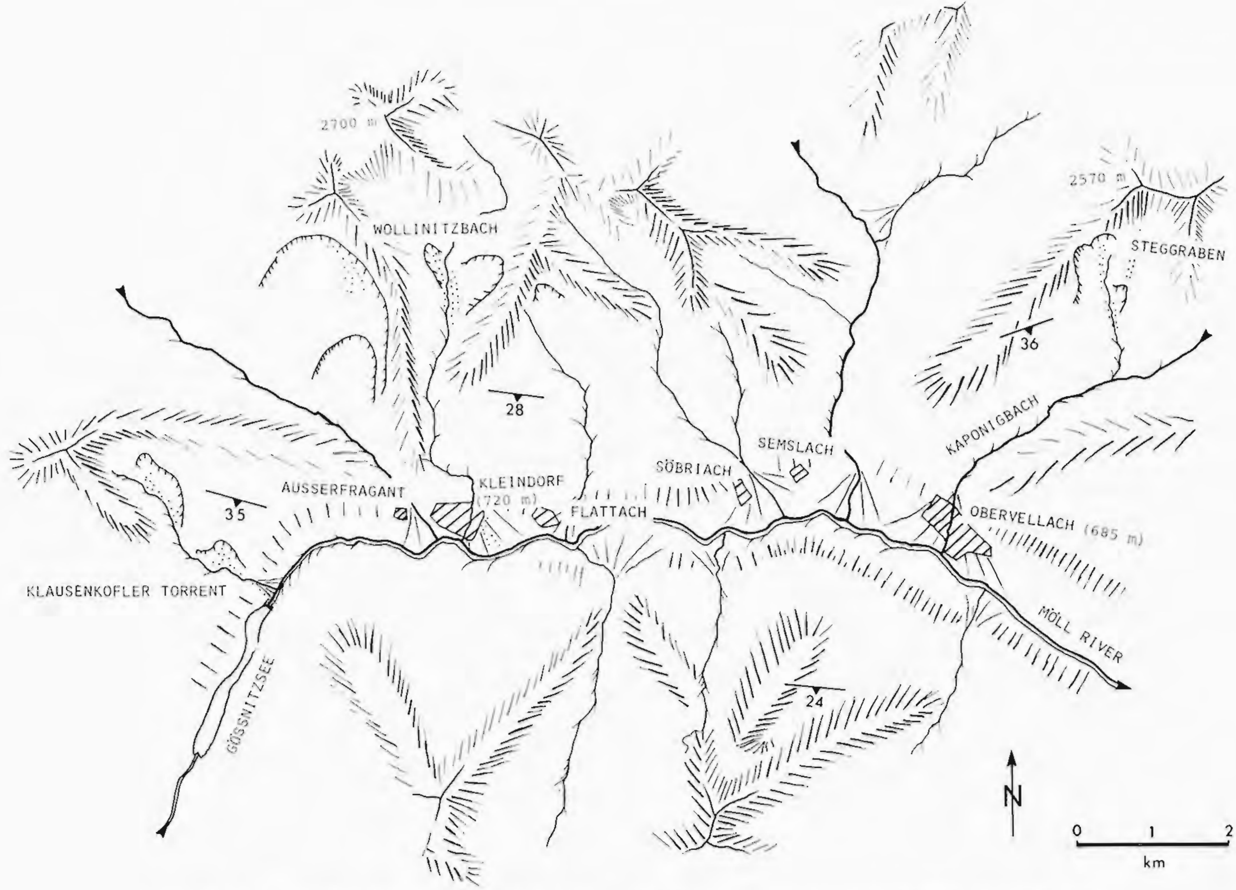


Figure 189: Index map of the Möll Valley showing the principal areas of historical debris flow activity on debris-covered south-facing foliation dip slopes.

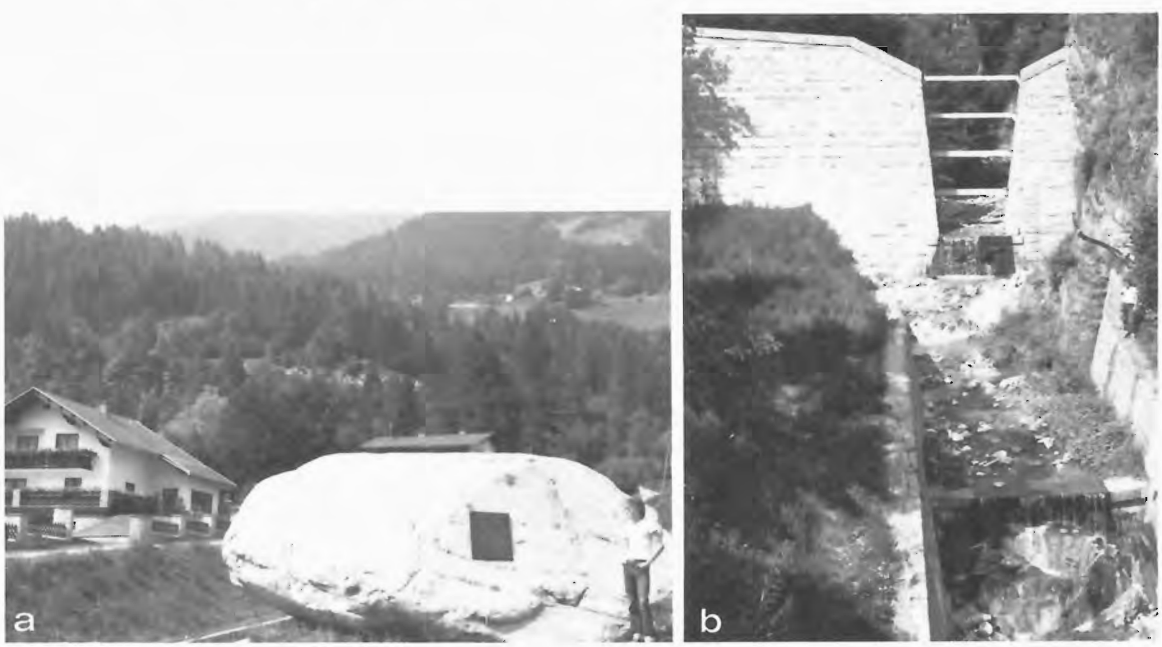


Figure 190: a) Large gneiss block near the apex of the Wollinitzbach cone; commemorative plaque refers to the catastrophe of 1966; note new residence behind the block (GSC 204168-H). b) Protective masonry-concrete-steel dam across the gorge of the Wollinitzbach Torrent; person on the right for scale. Steel beams define a selective discharge section ('filter dam') which prevents the passage of blocky debris flows but permits a gradual removal of accumulated debris by the normal erosive action of the torrent. (GSC 204168-I)

Spriana (A79)

Location: Val Malenco, Lombardia, Italy (E2)
Date(s): 1834

Spriana (754 m) is one of several small settlements clinging to narrow bedrock ledges on the lower slopes of Val Malenco (Fig. 191). Val Malenco is drained by the south-flowing Mallero Torrent which has carved a gorge into massive gneisses of the Pennine basement complex.

Spriana and Scilironi are located on the flanks of an overgrown rockfall cone on the east side of the Mallero gorge. The bouldery cone rises at an angle of 25 to 30° from the Mallero Torrent to a precipitous bedrock chute which in turn opens into a concave rockfall scar at approximately 1800 m elevation. The gneissic bedrock is cut by joints that parallel the trend of the valley.

In 1834 a rock avalanche broke away along a composite fracture zone above Spriana and obliterated most of the village (Montandon, 1933, p. 317). Gigantic blocks of gneiss from this and possibly older falls still pierce the terraced meadows between the old stone-roofed buildings of the community (Fig. 192). Some of these houses have recently been restored into modern homes.

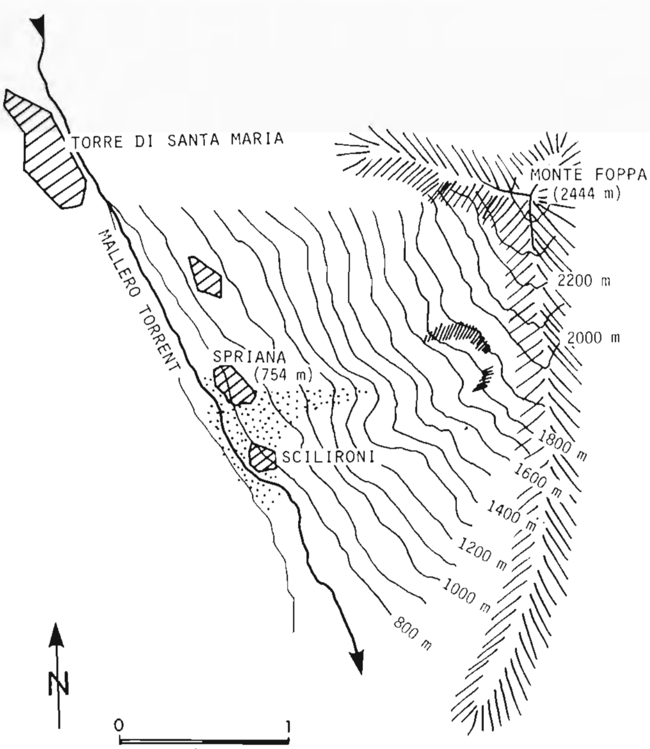


Figure 191: Sketch map of the gneissic bedrock slope flanking the Mallero gorge near Spriana.

Figure 192: Steep rockfall cone and part of Spriana. (GSC 204168-J)

Felsberg (A80)

Location: Chur, Graubünden (Grisons), Switzerland (E2)
Date(s): 13 March 1834

The old village of Felsberg (570 m) spreads over mounds of prehistorical slide debris at the foot of the Calanda Massif (1700-2400 m) north of the Rhine River (Fig. 193). The valley wall behind the village is a composite dip slope of massive Malm limestone of the Helvetic cover complex. On the lower part of the slope bedding dips approximately 35° to the south and together with steep fractures defines a surface of detachment of prehistorical rockslides. Towards the top of the cliff bedding dips between 25 and 30°.

Beginning on 13 March 1834, and continuing until 1867, a series of rockfalls, involving slabs of limestone up to 100 m in diameter, broke away from the headwall of the prehistorical detachment surface at about 1400 m elevation. Some of the blocks came to rest only a short distance behind Felsberg. By 1844 apprehension concerning a major rock avalanche led to the partial evacuation of the community to a new town site, named Neu-Felsberg, 1 km to the east. The anticipated rock avalanche did not materialize although minor rockfalls from the cliffs have been registered from time to time (Heim, 1932, p. 153-154).

In recent years the two Felsbergs — Altdorf and Neudorf — have absorbed suburban growth from the nearby cities of Chur and Domat/Ems; most of the new homes are within reach of a major rock avalanche from the Calanda cliffs (Fig. 194).

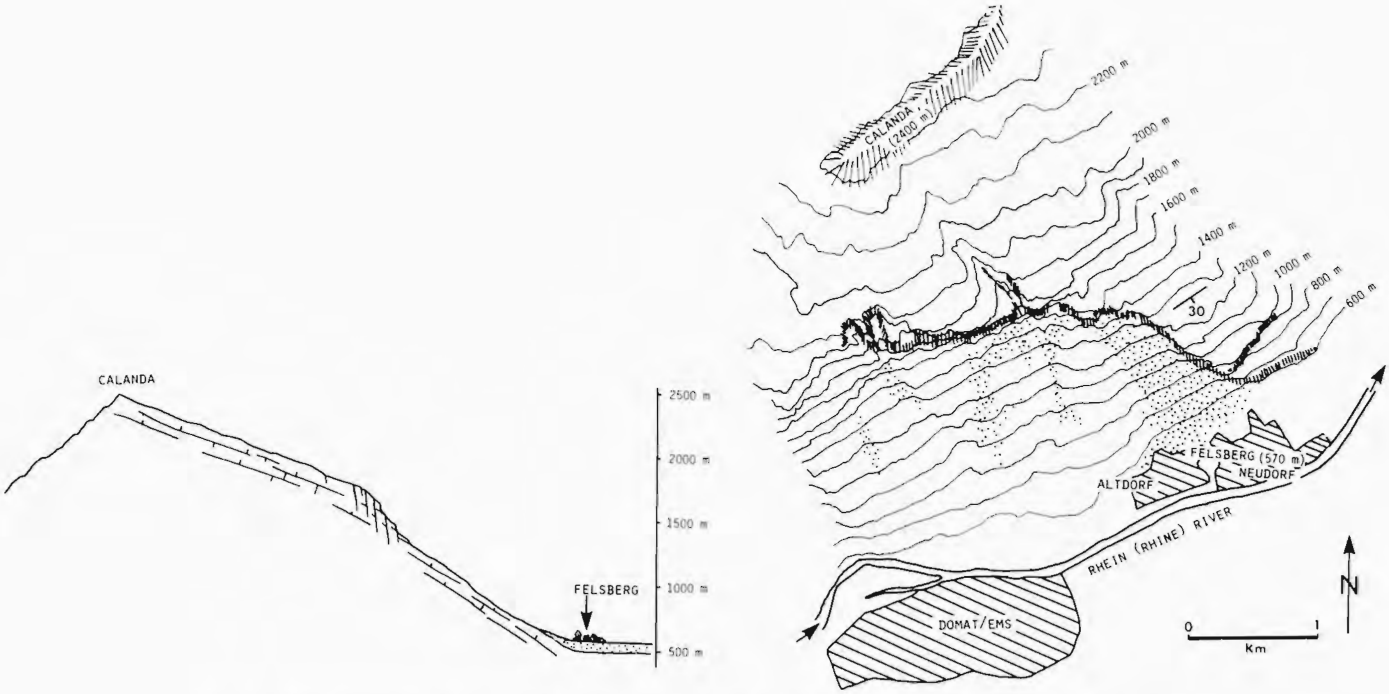


Figure 193: Sketch map and cross-section of the Calanda dip slope near Felsberg; note blanket of rockfall debris below the retrogressive scarp midway up the slope.

Figure 194: View of Felsberg-Altdorf and the carbonate dip slope behind the village; the church rests on a prehistorical slide block. (GSC 204168-K)

Peccia (A81)

Location: Val Lavizzara (Maggia Valley), Ticino, Switzerland (D2)
Date(s): 27 August 1834 (also 1760)

The village of Peccia (840 m) presently skirts the eastern flank of a small debris cone at the junction of the torrential Maggia and Peccia rivers (Fig. 195). The cone is bordered by steep bedrock slopes composed of granitic gneisses of the Pennine core complex. Foliation in the gneisses dips about 30° to the east and is cut by near-vertical north-trending fractures. A short distance north of Peccia the Maggia River flows through a narrow gorge whose east side is a sagging slope involving a total rock volume of more than 150 × $10^6 m^3$. Prehistorical movement along this slope blocked the Maggia channel and created a small aggradational basin on the upstream side.

Accelerated movement of this slope was triggered by intense rainstorms and flood events in September 1757 and July 1758; in 1760 a series of debris flows debouched from the Maggia gorge and leveled the community of Peccia. Shortly thereafter, the community was rebuilt farther to the east, but still on the boulder-strewn cone.

After the intense rainstorm that deluged the south-central Swiss Alps on 27 August 1834, the toe of the sagging slope again obstructed the flow of the Maggia River and bouldery flows once more demolished Peccia (Preiswerk, 1918; Montandon, 1933, p. 308). During this catastrophe approximately 50 × $10^6 m^3$ of material were in motion on the lower part of the sagging slope.

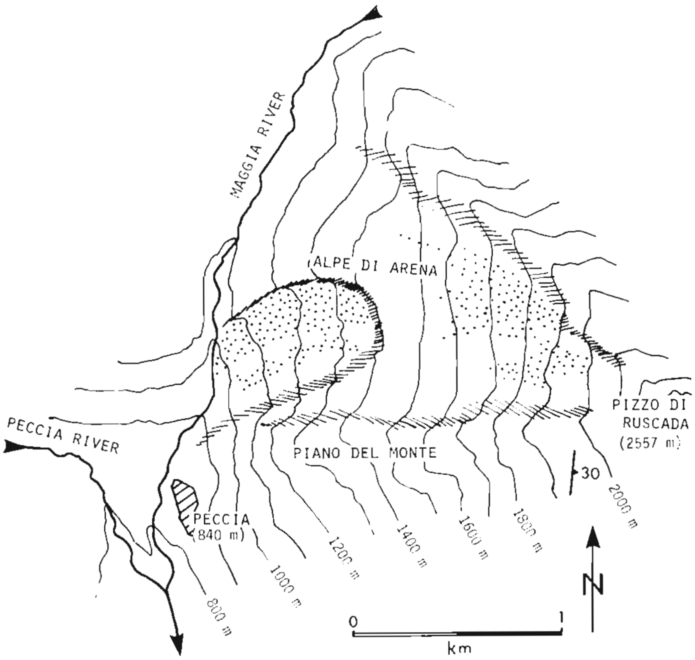


Figure 195: Sketch map of the sagging slope below Pizzo di Ruscada and subsidiary bedrock slump below Alpe di Arena; the gorge of the Maggia River was repeatedly blocked by movements of this slope.

Crodo (A82)

Location: Valle Antigorio, Piemonte, Italy (D2)
Date(s): 27 August 1834

Crodo (534 m) crowds the southern flank of the steep debris cone of the Alfenza Torrent along the west side of the Toce River (Valle Antigorio). Bedrock underlying the mountains in the vicinity of Crodo consists of massive, gently southeast-dipping gneisses of the Pennine core zone. The Alfenza basin above Crodo, however, is a huge terrace of sagging bedrock rimmed by the southern summit ridge of Monte Cistella (2880 m). Embankments of unstable bedrock and relict colluvium along the gorge between Mozzio and Cravegna sporadically fail due to deep-seated creep of the sagging rock mass (Fig. 197).

On 27 August 1834, during a severe rainstorm, slumps along the Alfenza gorge stemmed the flow of the swollen torrent. A massive plug of bouldery debris containing slabs of bedrock up to 12 m in diameter burst forth from the gorge onto the Alfenza cone and devastated half the village of Crodo (Montandon, 1933, p. 317).

Until recently the axial sector of the Alfenza cone was left undeveloped because of the threat of debris flows from the unstable uplands. However, check dams and protective structures have been completed above the cone and the first houses have sprung up on its central sector (Fig. 198). Sustained lateral pressure along sagging channel embankments has threatened to crush the check dams and will eventually necessitate repair or replacement of these structures.

Figure 196: Deep but dormant crown cracks in massive gneisses along the headwall of the Alpe di Arena detachment zone. (GSC 204168-L)

Since that time the town has survived, perched against the precipitous and not entirely stable bedrock spur of Piano del Monte. The Maggia River has been diverted into a hydroelectric tunnel just above the slide-blocked gorge. The sagging slope now displays an undisturbed mantle of mature forest, suggesting that at present it is dormant (Fig. 196).

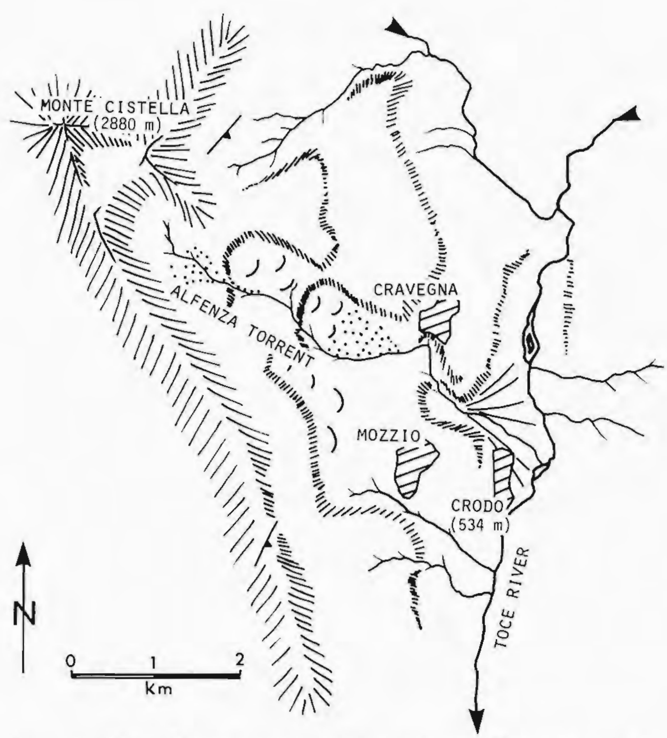


Figure 197: Sketch map of the Alfenza Torrent which flows through a sagging bedrock slope of gneisses overlain locally by thick surficial deposits.

Figure 198: a) View of the Alfenza cone; village of Crodo on the left; note densely forested slopes bordering the Alfenza gorge and relatively new residential buildings along the formerly avoided central sector of the debris cone (GSC 204168-M). b) Damaged wing of a check dam in the upper Alfenza Torrent; the structure is being crushed by lateral stresses exerted by the sagging slope (GSC 204168-N).

Torrent de St. Barthélemy (A83)

Location: Rhone Valley, Vaud, Switzerland (B2)
Date(s): 26 August 1835 (also 563 A.D., 11 October 1635, 12 May 1636, 25 June 1636, 26 September to 9 October 1926)

The channel of the St. Barthélemy Torrent leads from the west bank of the Rhone River (430 m) over the huge debris cone of Bois Noir to the hamlet of La Rasse (600 m) and up into the steep Gorge du Foillet. The gorge eventually opens towards the rugged and ice-covered summit ridge of the Dents du Midi (3100 m). The gorge of the torrent parallels the north-dipping contact between crystalline basement and sedimentary cover of the Helvetic zone (Fig. 199, 200). Vertical walls and unstable pinnacles of calcareous flysch and massive limestone along the Dents du Midi ridge frequently are the source of rockfalls. Glaciers (Plan Névé) and firn fields cover some of the bedrock shelves above the Gorge du Foillet. In the past, temporary blockage of slide debris in the Gorge du Foillet set off violent flows of debris to the Bois Noir cone.

During the decline of the Roman Empire the Bois Noir cone was probably the site of a military fortress and a town, known as Tauredunum, which controlled access to the upper Rhone Valley. In 563 A.D., a massive failure in the St. Barthélemy basin triggered a large debris flow (or flows), which demolished the settlement and blocked the flow of the Rhone River; the impounded river eventually broke through the low dam and a devastating flood wave raced downvalley to Lake Geneva (Lac Léman). Many people were said to have been killed (Montandon, 1933, p. 280-281).

On 11 October 1635, a series of slides from the summit ridge of the Dents du Midi produced pulses of debris which reached the cone. On 12 May and 25 June 1636, the masses of accumulated debris again were of sufficient volume to block the Rhone River temporarily (Montandon, 1933, p. 302-303).

On 26 August 1835, a small rock avalanche from the Cime de l'Est deeply scoured the ice fields of Plan Névé; the mixture of ice and rock descended through the gorge in four major pulses. The first pulse was the most destructive: it buried parts of the hamlet of La Rasse and destroyed much

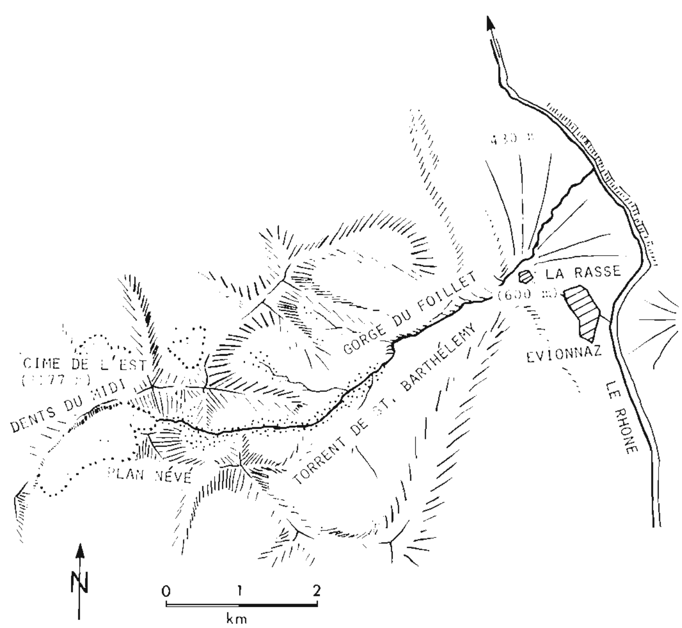


Figure 199: Sketch map of the Torrent de St. Barthélemy, its partly glacier-covered source area, and the large debris cone extending across the floor of the Rhone Valley.

agricultural land on the cone; the Rhone was dammed briefly (Heim, 1921, v. II, p. 461). This catastrophe was witnessed by a group of geologists, among them Elie de Beaumont. In 1847 de Beaumont wrote about his observations:

. . . 'I have seen in these muddy overflows a small-scale picture of the "phénomène erratique" as I conceive it'..

The term 'phénomène erratique' was then used for the seemingly incomprehensible occurence of erratic blocks of crystalline schists in the sedimentary foreland of the Alps. A catastrophist, Elie de Beaumont could not yet accept a glacial interpretation for erratic blocks which at the time was rapidly gaining support among European naturalists.

In September 1926, debris and remnants of snow avalanches from the previous winter filled the upper gorge. Torrential rains that began on 26 September mobilized this debris, generating flows that ravaged the cone until 9 October. The total volume of material brought down during this cycle was approximately $1 \times 10^6 m^3$ (Mariétan, 1927; Montandon, 1933, p. 317-318 and p. 327-328). Other debris flows in 1927 and 1930 interrupted the transportation routes to the upper Rhone Valley.

By 1939 several masonry check dams had been erected in the upper basin of the torrent to prevent undercutting of the colluvial embankments along the channel (Fig. 201a). In 1970 the torrent again showed signs of serious debris flow activity, and in 1975 a concrete dam, 45 m high, was erected across the upper gorge (Fig. 201b). Aggradation behind this structure will eventually render it ineffective. A large part of the impressive cone of Bois Noir is still covered by a pine forest (Fig. 201c); the community of Evionnaz and several recreational developments are now encroaching onto the southern sector of the cone.

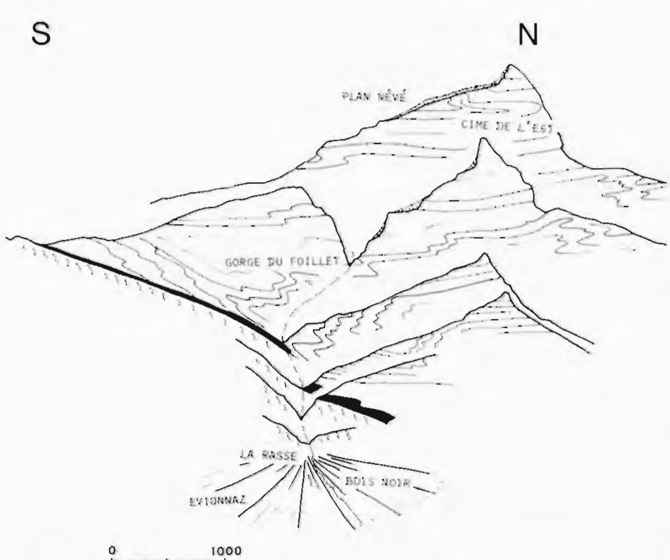


Figure 200: Schematic geological cross-sections along the Gorge du Foillet.

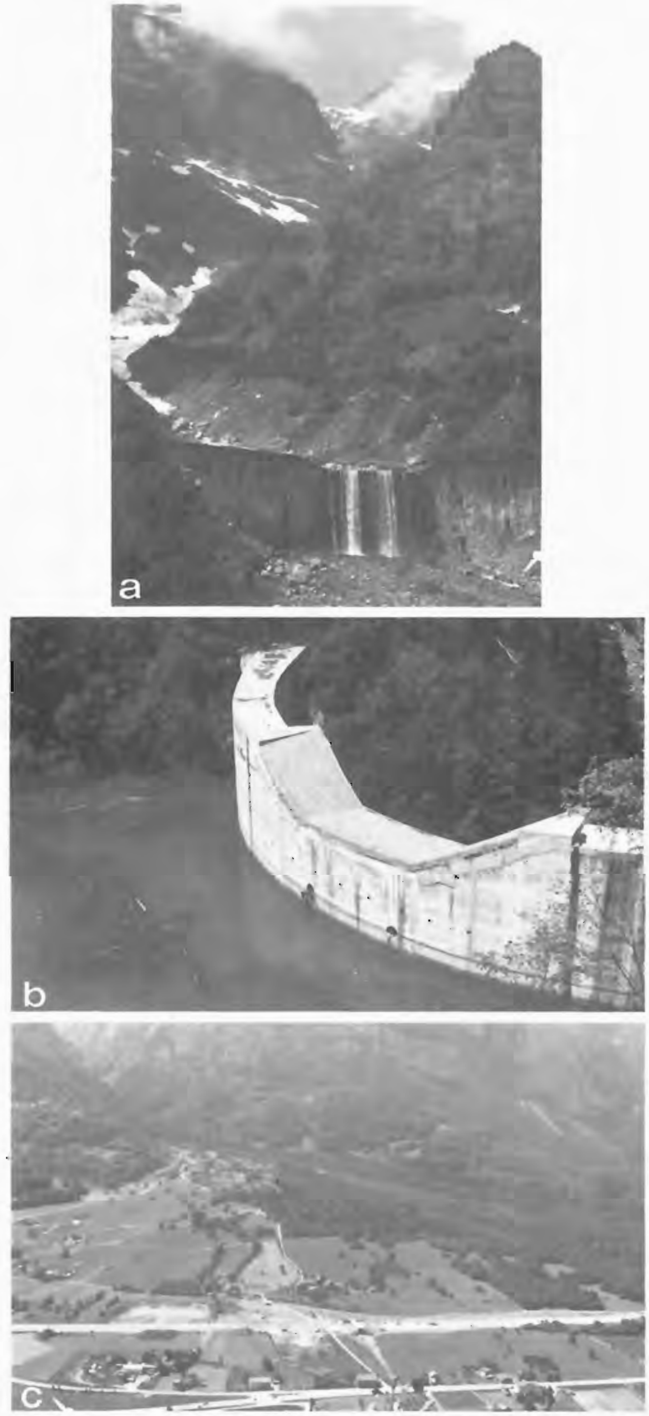


Figure 201: a) View towards the rugged summit ridge of the Dents du Midi from the top of the Gorge du Foillet; note large masonry check dam built in the 1930s (GSC 204168-O). b) Recently completed steel-concrete dam across the Gorge du Foillet; note temporarily impounded snowmelt runoff behind the structure (photograph taken in 1981) (GSC 204168-P). c) View across the Bois Noir cone; note protective pine forest on the northern sector of the cone (right); farms and residences dot the southern sector (GSC 204168-Q).

Verrès (A84)

Location: Dora Baltea Valley, Valle d'Aosta, Italy (C3)
Date(s): 1840 (also 11th century, 1408, 1620, 10 September 1857)

Near the town of Verrès (391 m) the Dora Baltea River breaks through mountain ranges composed of southeast-dipping metamorphic rocks of the Pennine and South Alpine basement complexes. Closely spaced settlements in the narrow valley cluster along debris cones and fans marking the junctions of tributary torrents with the flood-prone river (Fig. 202).

One of the most prominent of these cones is the rubbly surface of Arnad (361 m), which used to be known as 'Glarier d'Arnaz' (glarea = blocks, gravel). It spreads below a precipitous bedrock basin underlain by massive gneisses. Steep ravines feather out in the summit ridges of Monte Crabun (2710 m). Wedge failures of bedrock along these ravines develop into channelized flows to the cone. Destructive bursts of blocky debris on the Arnad cone occurred in the 11th century, in 1408, in 1620, and on 10 September 1857, when 30 people lost their lives (Montandon, 1933, p. 319-320). The community of Arnad has persevered in the

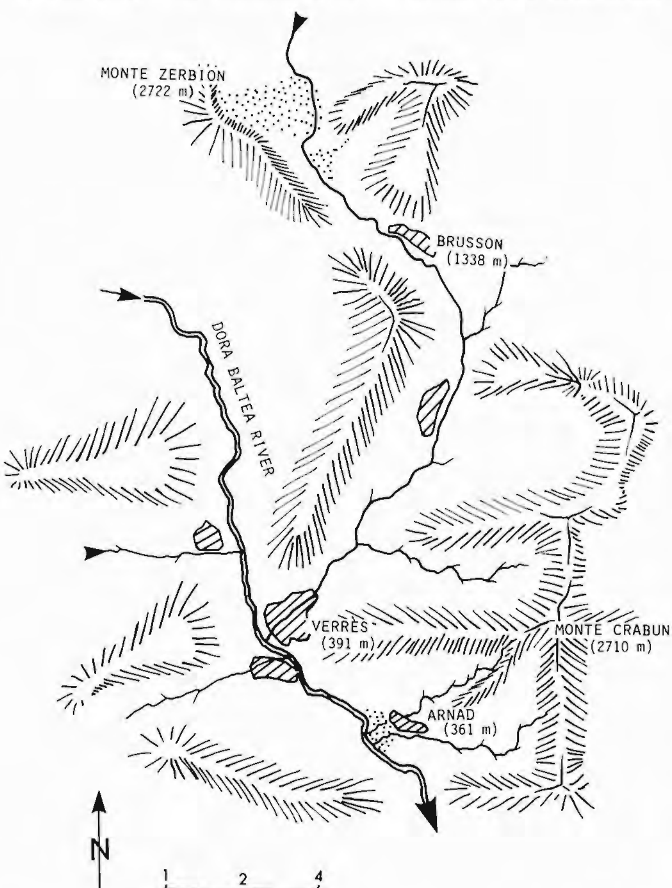


Figure 202: Index map of the lower Dora Baltea Valley near Verrès and Arnad.

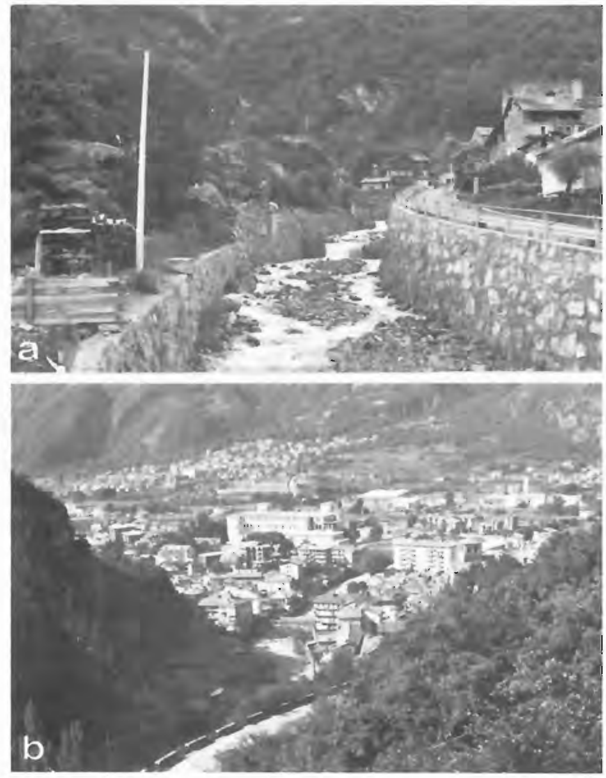


Figure 203: a) View of the apex of the built-over cone of Arnad; large gneissic blocks are visible to the left of the trained torrent channel. Paved masonry walls are about 2 to 3 m high (GSC 204168-R). b) View of Verrès from the gorge of the Evancon Torrent; fan apex and lowermost gorge are densely populated (GSC 204168-S).

face of these flows and recently has expanded onto blocky debris left by the last catastrophe (Fig. 203a). The lower parts of the 'Glarier' are now shared by new housing, industrial sites, and vineyards; the channel has been lined with high masonry revetments.

The problem of Verrès is slightly different. This town has grown at the mouth of the Evancon Torrent, on the gentle debris fan above the east bank of the Dora Baltea River. The large catchment basin of the Evancon Torrent (Val Ayas) is underlain by intensely fractured ultramafic and metamorphic rocks. Along the eastern flank of Monte Zerbion (2722 m) a prehistorical (?) slide mass of approximately 200 × $10^6 m^3$ crosses the valley floor and in the past has supplied abundant debris to the torrent (Fig. 202). In 1840 coarse bedload apparently jammed the narrow gorge above Verrès; the explosive release of the accumulated debris crushed a number of buildings in the town's centre, killing 80 people (Montandon, 1933, p. 318).

Today Val Ayas faces tremendous development pressure and many hazardous reaches along the torrent, including the floodplain below the Monte Zerbion slide mass, host permanent dwellings or campgrounds. Verrès still clings to the exit of the Evancon gorge and the torrent crosses the built-over fan in a masonry-walled channel (Fig. 203b).

Brannenburg (A85)

Location: Degerndorf, Inn Valley, Bayern (Bavaria), Germany (H1)
Date(s): 9 to 18 August 1851

Brannenburg (509 m) is a small community on the north flank of a large debris cone west of the Inn River in the front ranges of the eastern Alps (Fig. 204). This cone is crossed by the Kirchbach Torrent whose catchment area straddles the tectonic contact between south-dipping Austroalpine carbonate complexes and intensely deformed Helvetic flysch terrain. Approximately 1 km west of the apex of the cone the southern rim of the drainage basin is a steep scarp slope (Schrofen) consisting of a lower recessive evaporitic shale-dolomite unit and an upper cliff-forming dolomite.

The summer of 1851 was extremely rainy throughout the front ranges of the Alps; the Inn River flooded at least three times. On 9 August 1851, excessive infiltration of rainwater caused the collapse of approximately $3 \times 10^6 m^3$ of carbonate-evaporite rock from the Schrofen face. By 18 August the fragmented carbonate mass had spread across the channel of the Kirchbach Torrent. Saturated by the impounded waters behind the slide mass the lobe of debris began to advance towards Brannenburg which was spared destruction by a ledge of bedrock in the path of the debris masses (Gümbel, 1861, p. 290). However, in the small hamlet of Gmain, south of Brannenburg, six buildings were crushed before the debris lobe came to a halt.

In the years since this disaster a mature forest has grown on the bulky debris lobe and check dams along the torrent prevent mobilization of debris from its embankments. The lower fringe of the debris lobe recently has been invaded by large hotels. The breakaway scar on the Schrofen, although still bare, shows only traces of rockfall activity.

Upper Vintschgau-Val Venosta (A86)

Location: Vintschgau (Val Venosta), Südtirol (Alto Adige), Italy (F2)
Date(s): 17 to 18 June 1855 (also 18 to 19 May 1847, 7 June 1849, July 1851, 3 June 1855)

The upper Vintschgau (Val Venosta) is a wide, glacially sculptured valley, flanked by mountains composed of high grade metamorphic rocks of the Austroalpine basement complex. It is drained by the Etsch (Adige) River which heads in two lakes (Reschensee and Haidersee) near the border between Italy and Austria (Fig. 205). Tributaries of the Upper Etsch River enter the valley over huge debris cones which restrict the flow of the river. Catchment basins are rimmed by peaks 2000 to 3000 m in elevation. In the past many of the steep slopes of the upper Vintschgau have suffered from overgrazing and deforestation. Mass movements have occurred repeatedly during periods of intense rain and delayed snowmelt. It comes as no surprise that local legends tell of disappeared cities, bursting lakes, and irate dragons!

The largest of the debris cones, reputedly the largest in the Alps, is the 'Malser Heide' (Fig. 206). Dropping off from an astonishingly small catchment basin, this cone has forced the Etsch River against the west side of the valley for a distance of almost 10 km. It also impounds the Haidersee, a lake whose level once depended on the subtle interplay between debris accumulation on the cone and erosional downcutting by the torrential Etsch River. Numerous historical debris flows have swept across the barren cone and the hamlet of Plawenn near its apex was destroyed at least three times. Below the Malser cone the floods of the unruly Etsch River forced other settlements onto cones and fans along the sides of the valley, exposing them to sporadic debris flows from other basins.

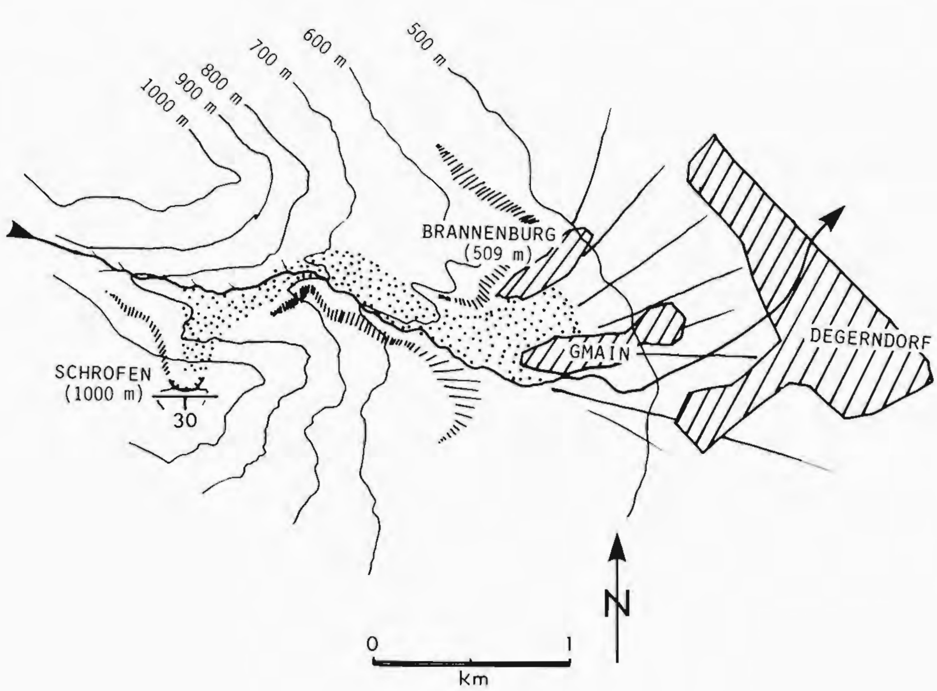


Figure 204: Sketch map of the Schrofen scarp face near Brannenburg and the blocky debris lobe of 1851.

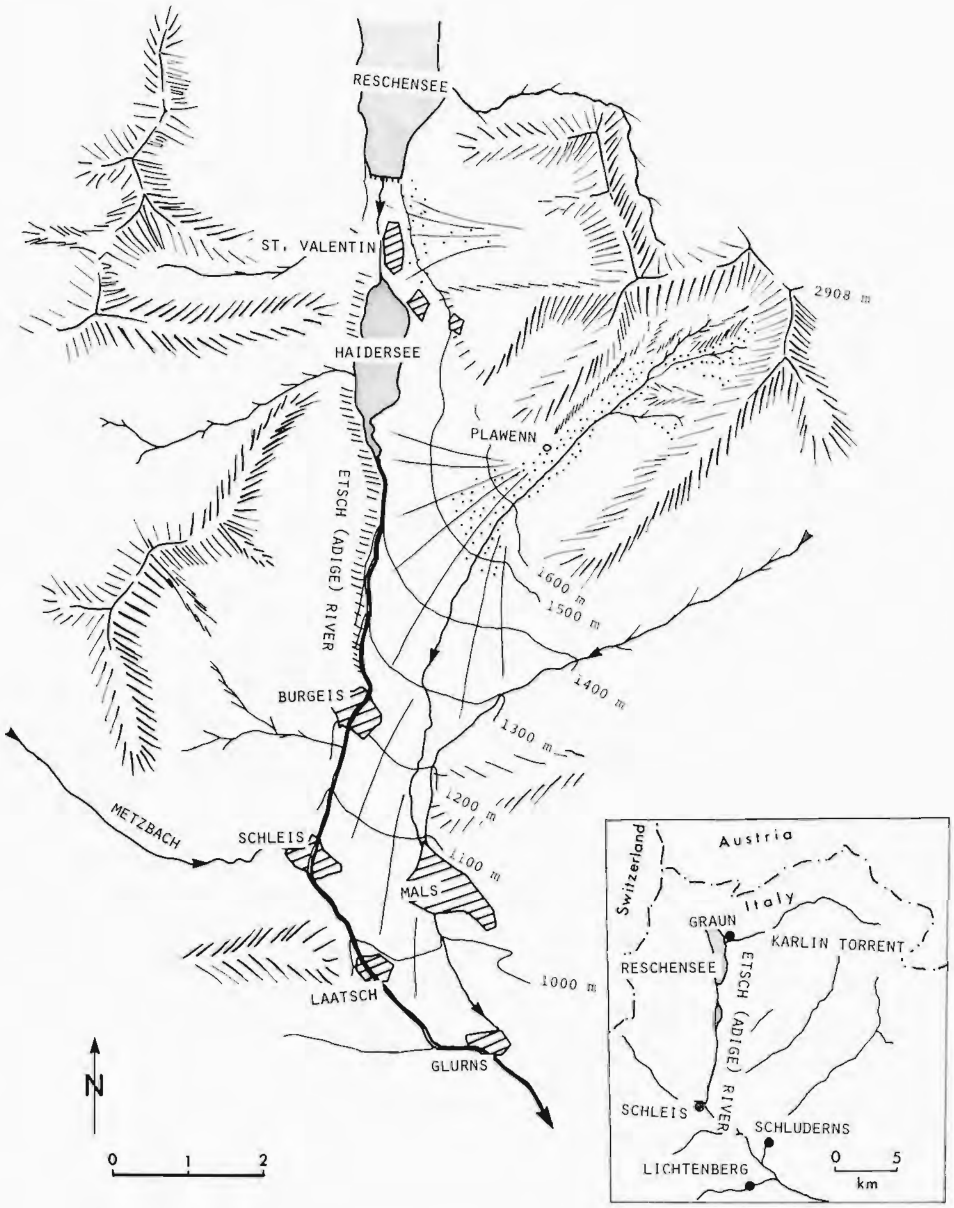


Figure 205: Index and sketch map of the upper Vintschgau (Val Venosta); note the huge debris cone of Mals east of the Etsch (Adige) River.

In 1844, 1845 and 1846 the upper basin of the Etsch River experienced three major floods: on 25 October 1844, after heavy autumn rains; on 13 August 1845, after a lengthy period of snowmelt, rainstorms and warm winds from the south; and on 17 May 1846, after rapid snowmelt. These unusual climatic events triggered numerous failures in colluvium and bedrock in the upland basins; in the following decade torrents released great quantities of bedload and occasional debris flows over their cones.

On 18 and 19 May 1847, extensive mass movements set off pulses of debris, totalling more than $0.3 \times 10^6 m^3$, which overwhelmed the community of Lichtenberg (900 m); twenty-six houses were completely obliterated. Only two years later, on 7 June 1849 another sudden burst of debris flattened other buildings in the same community and eight people were killed. In July 1851 and on 3 June 1855 debris flows again erupted over Lichtenberg. After these disasters a large stone-masonry dam was constructed along the eastern sector of the fan to deflect the flows along the torrent away from the community.

In the summer of 1855, when the old community of Lichtenberg received its last blow, other human works along the Etsch were demolished in one of the greatest natural disasters to befall the region. The spring melt of 1855 had been long delayed and when temperatures rose abruptly to their seasonal values the heavy winter snowpack quickly disappeared. Runoff was intensified by heavy rains. On 17 and 18 June 1855, pulses of debris from the Metzbach Torrent buried dozens of buildings in Schleis. Meanwhile the rising waters of the Mittersee (now part of the Reschensee hydroelectric reservoir) and the Haidersee in the upper Etsch basin overflowed, thus causing deep erosion of the outlet of the Haidersee; floods breaching the channel of the Etsch River demolished buildings in four small communities along its banks (e.g. Burgeis); it also destroyed most of what was left of Schleis (Sonklar, 1883, p. 100-102; Strele, 1936, p. 133-134; Stacul, 1979, p. 50).

Early attempts to control the Etsch River were directed towards systematic lowering of the lake levels. After the catastrophe of 1855 more attention was paid to training the river channel below the lakes. More recently, Mittersee and Reschensee have been combined to form the Reschensee hydroelectric reservoir that inundated parts of the settlement of Graun. The glacial Karlin Torrent, whose cone used to separate the two lakes, has accumulated a large delta into the unsightly reservoir. Below Haidersee some communities (e.g. Burgeis) have recently expanded onto the formerly barren surface of the Mals cone (Fig. 206); this development has been accompanied by the construction of protective dams above communities threatened directly by debris flows.

Figure 206: View of the large cone of Mals and the settlement of Burgeis in the foreground. (GSC 204168-T)

Bilten (A87)

Location: Linth Valley, Glarus, Switzerland (E1)
Date(s): 29 to 30 April 1868

The village of Bilten (430 m) is situated on a small debris cone at the foot of a bedrock cliff composed of conglomerate and mudstone of the Molasse Zone (Fig. 207). Small ravines extend from debris-covered benches in the uplands to the Linth Valley.

The winter of 1867-68 produced exceptionally deep snowpacks throughout most of the north-central Alps. In early spring wet snow avalanches were common. One of these large avalanches came to rest above a northeast-facing conglomerate rib 500 m above Bilten. As the snow avalanche melted, underlying thick surficial deposits became saturated and began to slip towards the edge of the cliff. There the debris mass broke away piece by piece and collected in a bedrock ravine above the apex of the debris cone of Bilten. The spatulate lobe attained a volume of $0.18 \times 10^6 m^3$ and engulfed 20 buildings in Bilten (Heim, 1932, p. 39-40).

Today there is little to be seen of the 1868 debris flow deposit; large apartment buildings have been built on the upper segment of the debris cone. The torrent above the cone has long been controlled by an array of check dams.

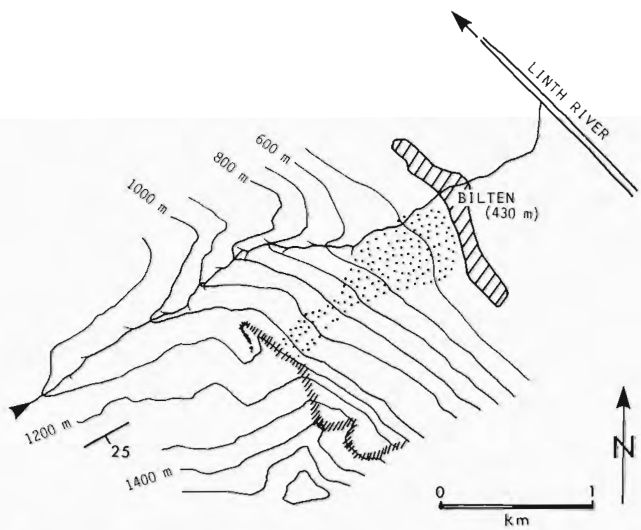


Figure 207: Sketch map of Bilten and the snowmelt-triggered debris flow of 1868.

1868 Ticino Rainstorm (A88)

Location: Ticino and Graubünden (Grisons), Switzerland
Date(s): 17 September to 7 October 1868

In autumn of 1868 the south-central part of the Alps suffered one of the most severe rainstorms in the recorded history of the region. The year 1868 had begun with a winter and spring characterized by exceptional snowpacks leading to numerous snow avalanches in eastern Switzerland and western Austria. The summer that followed was hot and dry. Then in August and early September thunderstorms and localized downpours struck the western Alps. These storms set the stage for a 20-day period of almost uninterrupted precipitation beginning on 17 September. At first, heavy snow fell in the high mountains between the Ticino and Rhine rivers. Then, on 25 September the snowfall changed to torrential rain. Exceptionally warm winds from the south beginning on 27 September and lasting until 4 October brought more heavy rain and unseasonable temperatures to most upland basins. Locally rainfall amounted to more than 250 mm/24 h and on the Passo del San Bernardina (2100 m) more than 1600 mm of rain was recorded between 17 September and 6 October (Salis, 1868; Hellwald, 1869).

Landslides, debris flows, and floods afflicted much of the region. Rescue operations could only begin after the storm had moved eastward into the Trentino region. In November a sharp drop in air temperatures brought stabilizing winter weather.

The enormous impact of this natural disaster spurred implementation of improved forest practices, and construction of torrent control works in eastern Switzerland and northern Italy.

Nolla (A89)

Location: Thusis, Hinterrhein (Rhine) Valley, Graubünden (Grisons), Switzerland (E2)
Date(s): 27 to 28 September 1868 (also 15 August 1585, 1705 to 1719, 30 November 1807, 1817, 1834, 5 July 1938)

The Nolla Torrent is a western tributary of the Hinterrhein (Rhine) River (Fig. 208). The Nolla Torrent joins the Hinterrhein Valley where it opens into the broad Domleschg plain. Thusis (730 m), strung out along a terrace above the river near the mouth of the torrent, used to be a major medieval supply and trading centre.

The Nolla Torrent has two principal branches: the 'Schwarze Nolla' (=black Nolla) fed by tributary ravines from the Heinzenberg slope, an undulating surface underlain by sagging masses of Pennine slates; and the 'Weisse Nolla' (=white Nolla) draining Pennine carbonate cliffs south of the Heinzenberg slope.

A chronicle of the early 16th century describes the Nolla as a pleasant creek with mills along its lower course. Only a few decades later it had acquired an evil reputation (Clavuot, 1973, p. 44). The cycles of destructive debris flows that began to ravage the Domleschg plain were probably initiated by intense cultivation and deforestation of the Heinzenberg slope, and a generally deteriorating summer climate during the late 16th century. The 'roar' of moving debris in the

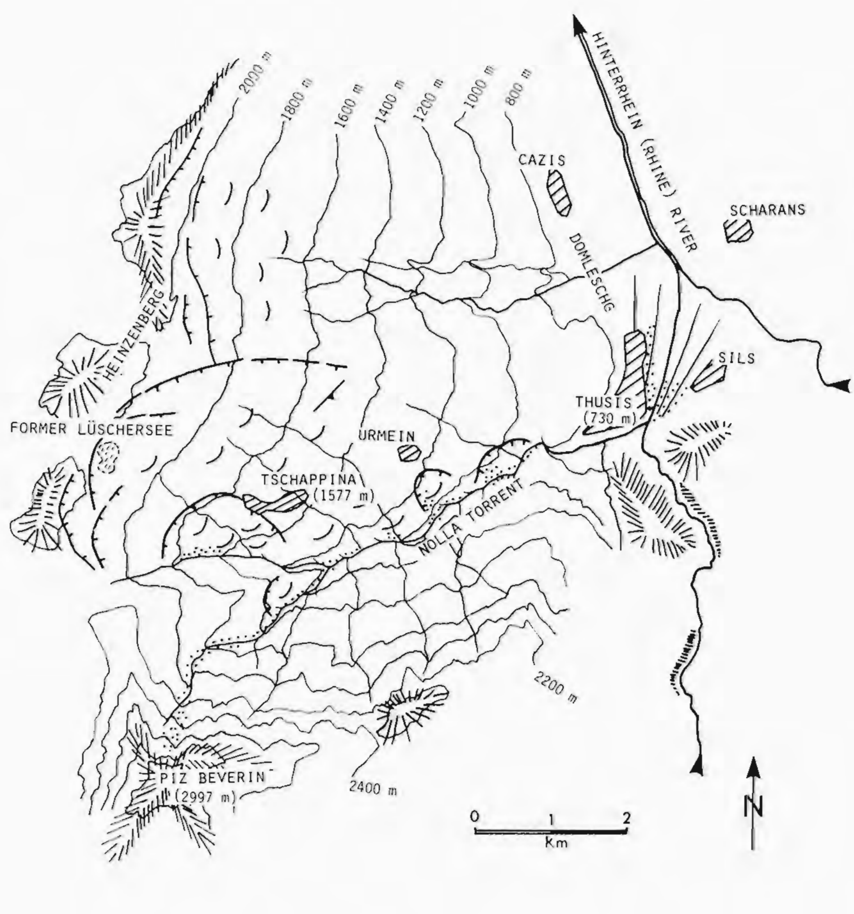


Figure 208: Sketch map of the Nolla basin and the sagging Heinzenberg slope; broken lines with barbs indicate the main head scarps of the sagging low grade metamorphic terrain.

gorge of the torrent was soon attributed to an infuriated dragon which supposedly had taken up quarters in the sag ponds and other depressions of the fog-shrouded uplands.

On 15 August 1585, after rains had fallen for most of the summer, debris flows erupted from the Nolla gorge and stemmed the flow of the Hinterrhein River. Rupture of the blocky welt which stretched across the valley sent a deluge of mud and rocks over the cultivated Domleschg. Between 1705 and 1719 a series of debris flows raised the channel of the Hinterrhein River to such a level that flooding became an annual disaster (Salis, 1892, v. 2, p. 13).

A particularly severe cycle of debris flows began during the wet years of 1805 and 1806. Finally, on 30 November 1807, after another exceptionally wet summer and early autumn, large slides occurred along the Nolla gorge below Tschappina (1577 m). Houses at Tschappina swerved off their foundations and several people were killed. The debris flows that soon debouched on the debris fan below Thusis raised the braided channel of the river by 12 m. Similar flows again shifted and raised the channel in 1817 and 1834, inflicting severe hardship on downstream communities whose livelihood depended on crops grown on the Domleschg plains.

In this period of distress the first attempts to train the Hinterrhein River were undertaken. However, progress was not satisfactory because of undiminished discharge of debris from the unbridled Nolla Torrent during every rainstorm. During the great regional storm between 27 and 28 September 1868, more than $1 \times 10^6 m^3$ of blocky debris reached the channel of the river. Mass movements were accentuated by the overflow of the Lüschersee, the largest sag pond at the head scarp of the Heinzenberg slope (Salis, 1892, v. 2, p. 13-14).

It was now recognized that without a comprehensive control program along the Nolla Torrent the dyking of the Hinterrhein River was doomed to failure. Thus began a long and ultimately successful struggle against the Nolla dragon! Short-term neutralization of the main debris sources was achieved by masonry spurs and check dams along the unstable embankments of the torrent (Fig. 209). Channel revetments and flood control dykes were added farther downstream. Later, the steepest embankment sections were reforested and the Lüschersee sag pond drained. Runoff from the alpine meadows of the Heinzenberg crest has been and still is being carefully channeled along drainage ditches (Fig. 210) which are adjustable to allow for continued deep-seated creep amounting to between 10 and 20 cm/year and involving slaty bedrock to a depth of approximately 150 m (Jäckli, 1957, p. 56). Many check dams on the Nolla crack slowly under stresses propagated from the Heinzenberg slope and are buried by slumps and small debris flows. They have to be replaced from time to time by ever stronger concrete dams to maintain a reasonably 'clean' torrent.

The efficiency of this system was tested when a rock avalanche from the carbonate crags of Piz Beverin (2997 m) overloaded the channel of the Weisse Nolla on 5 July 1938. Although more than $0.2 \times 10^6 m^3$ of debris was transported through the Nolla channel, both the torrent and the Hinterrhein maintained their positions during this phase of aggradation (Jäckli, 1957, p. 32 and 56). At present the difficult problems of the past in the upper Domleschg are almost forgotten; the bouldery surface of aggradation of 1868 is now a terrace 10 m above the Hinterrhein River and hosts an open pine forest, pleasant camping facilities, and an outdoor sports centre. Ski facilities and new hotels dot the Tschappina upland area.

Figure 209: Slumping bedrock embankments along the Nolla channel as seen at the turn of the century; note shifted and partly crushed masonry check dams (from Salis, 1914).

Figure 210: View of surface drainage ditches on the undulating upper part of the Heinzenberg slope; snowmelt runoff is thus prevented from accumulating in and infiltrating along sag ponds on the slope. (GSC 204168-U)

Campo (A90)

Location: Valle di Campo, Ticino, Switzerland (D2)
Date(s): 27 September to 4 October 1868 (also 1859)

The Rovana Torrent in the Valle di Campo is an east-flowing tributary of the Maggia River which drains rugged mountains underlain by high grade schists and gneisses of the Pennine core zone (Fig. 211, 212). The north side of the Campo valley is partly a sagging dip slope; foliation planes are inclined approximately 20° towards the Rovana channel. The south side of the valley is a steep scarp slope mantled by rockfall cones. As in other districts of the Ticino, settlements are preferably sited on gentle, sunny south-facing dip slopes. Campo (1400 m) is an old village located on a terrace several tens of metres above and north of the Rovana gorge. The gorge cuts the toe of an enormous ($200 \times 10^6 m^3$) sagging mass of mica schist below the Bombogn Massif (2330 m).

Up to 1850 no dramatic slope movements were registered at Campo, although sporadic slumping was noticed along the toe of the slope around 1780. In 1851 large-scale logging was carried out in the uplands west of Campo and the channel of the Rovana Torrent was used to drift logs down the valley. Repeated 'flushing' of the torrent during the log drifting began to undercut the terrace of Campo with the result that the western half of the sagging slope began to slide towards the gorge. Soon about $120 \times 10^6 m^3$ were in motion and in 1859, during an extended period of rain, several buildings in Campo collapsed. As a first countermeasure against the impending destruction of an ancient church and surrounding farmhouses log drifting was discontinued. But the deep-seated creep of the broken terrain continued. In late September or early October 1868, when a catastrophic rainstorm deluged the Ticino, a small rock avalanche became detached from the scarp face along the south flank of the Rovana gorge; its debris lobe pushed the raging waters of the torrent against the toe of the Campo terrace. Erosion of the toe zone and infiltration of water into open cracks along the head scarp of the sagging slope again accelerated motion of the slide mass and another group of buildings in Campo soon veered off their foundations. After this disaster the rate of movement of the slope decreased to a tolerable level. Since

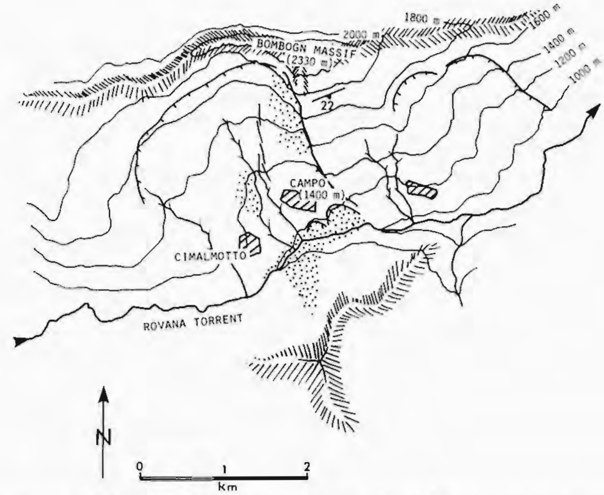


Figure 211: Sketch map of sagging bedrock terrain in the vicinity of Campo in the upland basin of the Rovana Torrent.

1889 stone-masonry spurs, groins, and check dams have been installed to keep the Rovana from eroding its unstable northern embankment. Nevertheless, high rates of infiltration in the uplands during years of abnormal snowmelt or rainfall tend to accelerate the steady creep of the rock mass (e.g. 1927 and 1951). Between 1892 and 1960 the church of Campo moved a total of about 15 m horizontally and settled almost 3 m vertically. Distortions of the buildings are corrected from time to time (Fig. 213). In the same period the Rovana Torrent has lowered its channel 14 m and eroded approximately $3.5 \times 10^6 m^3$ from the toe of the slide mass. Most of this debris has been carried to the Maggia River and has contributed to a recurrent flood problem on the Maggia delta along the north arm of Lago Maggiore.

Between 1964 and 1966 piezometers were installed in the slope. They indicated that artesian water pressures exist below the terrace of Campo. Plans were drawn up to eventually drain the sagging rock mass by means of a tunnel more than 1 km long and extending into the slope from beneath the eastern detachment zone (Heim, 1932, p. 49-55; Hirsbrunner, 1960, p. 20-35; Lichtenhahn, 1971b).

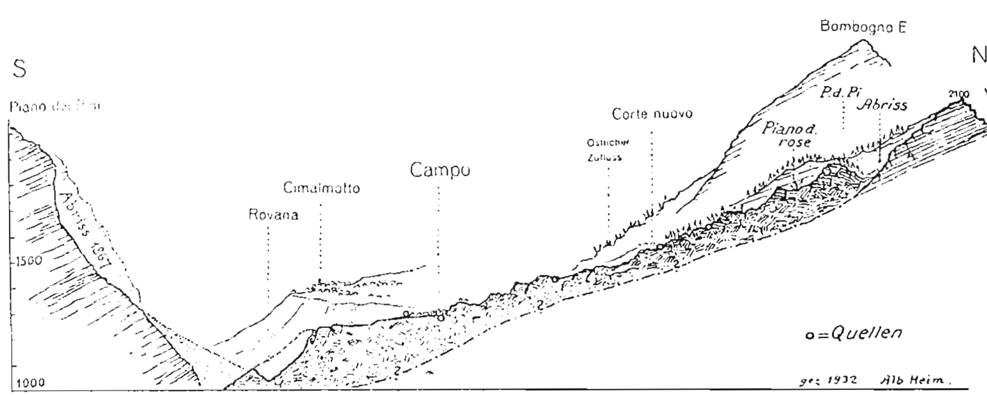


Figure 212: Cross-section of the slide mass underlying the village of Campo and overlying the foliation dip slope of the Bombogn Massif (from Heim, 1932).

Figure 213: a) The old church of Campo which during the last century has shifted at least 15 m horizontally (GSC 204168-V). b) Close-up view of the cracked and distorted church wall (GSC 204168-W).

Bec Rouge (A91)

Location: Tarentaise, Savoie, France (B3)
Date(s): May-July 1877 (also 30 August 1882, 3 June and 16 September 1883, 20 July 1895)

The Bec Rouge Massif (2515 m), a pointed ridge east of the upper Isère River (900 m) is composed of intensely fractured conglomerate, quartzite, and slate of the Pennine basement complex (Fig. 214). Bedding dips 30 to 40° to the south. At the foot of the dip slope the narrow gorge of the Torrent de St. Claude drops westward to the Isère River passing between the hamlets of Le Miroir and La Masure (1200 m). Parts of these hamlets rest on prehistorical landslide rubble. In the early 19th century most of the slopes below the summit ridge of Bec Rouge were denuded by extensive clearcuts.

Spring of 1877 was characterized by long delayed snowmelt of an exceptional winter snowpack. Snowmelt infiltrated the fractured bedrock of the Bec Rouge slope and the first threatening rockfalls were registered in Le Miroir in May 1877. During the following months the slope above opened along an arcuate array of cracks and a rock mass of approximately $3 \times 10^6 m^3$ began to slide away on a complex zone of rupture along its eastern flank. Except for incessant rockfalls the motion of the slide mass was slow. Eventually, the disintegrating toe settled into the gorge and blocked the Torrent de St. Claude over a distance of 0.5 km. Some buildings in Le Miroir were destroyed by shifting ground and rockfalls. A more serious problem was posed by the small lake that had been impounded. Violent overflows from this lake destroyed the hamlet of Champet on the cone of the Torrent de St. Claude: first, a regional rainstorm on 30 August 1882, mobilized $0.2 \times 10^6 m^3$ of debris from the rubbly toe of the slide mass, which burst over fields and houses of Champet; then, on 3 June and 16 September 1883, even more violent flows devastated the remaining fields and buried the abandoned ruins of the hamlet; finally, on 20 July 1895, during a local thunderstorm, $0.6 \times 10^6 m^3$ of debris covered the entire cone and took out the road to the upper Isère region. At this stage six transverse masonry check dams were built to prevent further erosion along the toe of the slide mass (Fig. 215). Drainage works on the landslide itself reduced continuing downward creep and reforestation impeded the development of new ravines in the scarred uplands (Mougin, 1914, p. 721-725, and 1931, p. 525-531).

With these remedial measures the destructive mass flows to the cone ceased and soon a mantle of vegetation covered the abandoned site of Champet. In recent years the cone has become a major gravel pit. Some of the old buildings of Le Miroir have been abandoned, others have been converted into vacation homes. The upland cliffs of Bec Rouge adjacent to the slide scar of 1877 are in a state of precarious equilibrium.

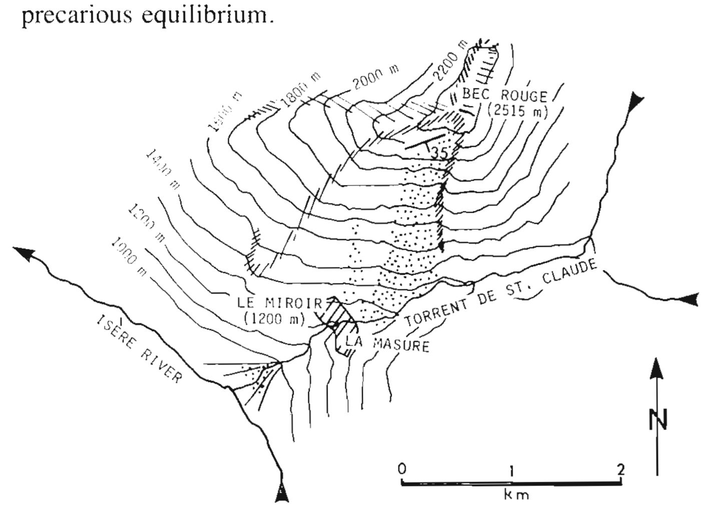


Figure 214: Sketch map of the detachment zone and slide deposits on the Bec Rouge dip slope.

Figure 215: Masonry check dams along the rubbly toe of the Bec Rouge slide; check dams have prevented the torrent from eroding the loose landslide deposits. (GSC 204168-X)

Ahrn Valley (A92)

Location: Südtirol (Alto Adige), Italy (G2)
Date(s): 17 August 1878 (also 14 to 15 September 1867, 30 June 1959)

The upper Ahrn Valley (1000 m) is drained by the southwest-flowing Ahr Torrent which originates in numerous high-gradient tributary ravines joining the main torrent almost at right angles (Fig. 216). The uplands are underlain by south-dipping schists of the Pennine Tauern Window. At their mouths the tributary ravines flow across terraces of late Pleistocene ice-margin deposits.

Medieval copper mining and smelting led to early deforestation of the high basins which later was aggravated by overgrazing of the alpine pastures. Deep erosional ravines developed in the unconsolidated deposits flanking the Ahrn Valley (Bazing, 1872). Old chronicles and recent excavations tell a long and sad story of debris flows, floods, heavy damage, and loss of life in the settlements of the region (Stacul, 1979, p. 101-102).

Between 14 and 15 September 1867, an intense heavy rainstorm pounded the upper Ahrn Valley, triggering numerous debris flows. One of them destroyed 25 buildings in the village of St. Jakob and blocked the Ahr Torrent to form a small lake. When the debris barrier burst a deluge of mud and rocks blanketed agricultural land below (Strele, 1936, p. 134).

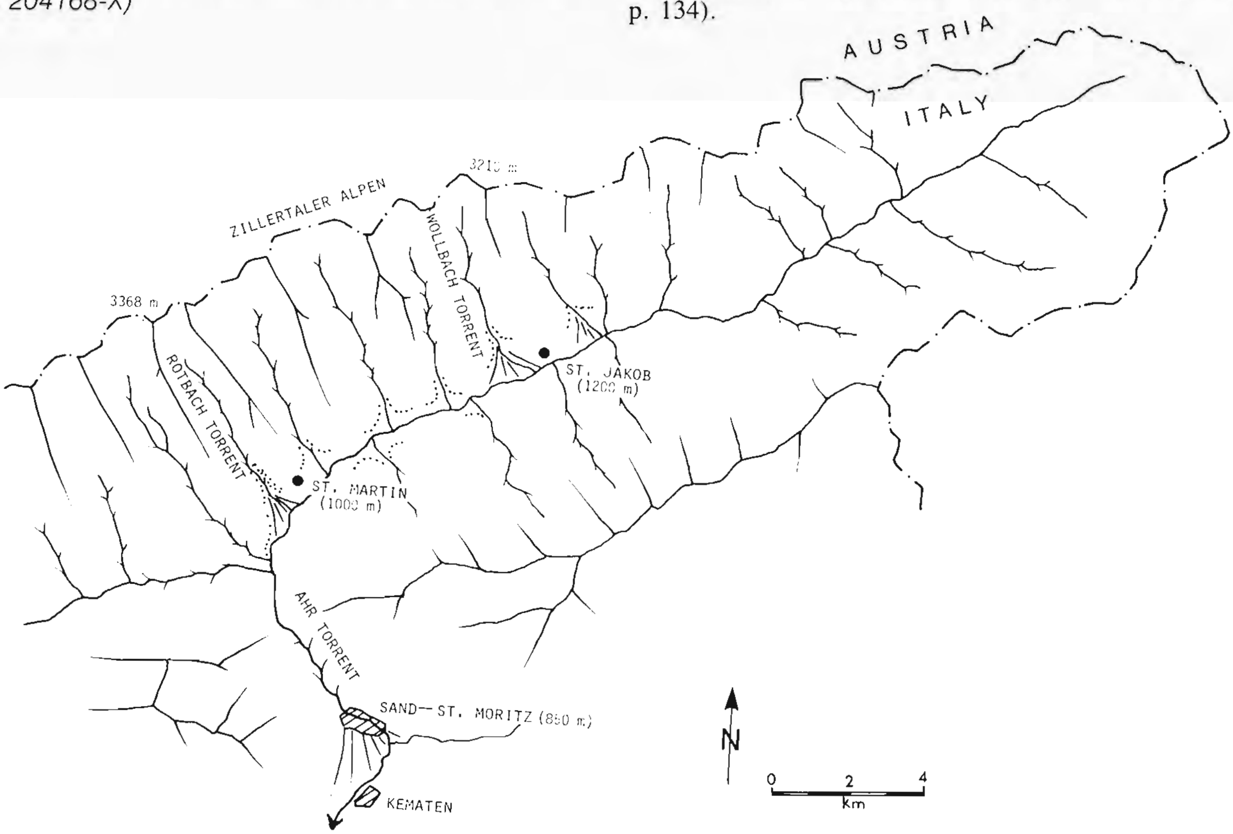


Figure 216: Index map of the Ahrn Valley; note closely spaced joint-controlled tributary torrents.

Between 16 and 17 August 1878, a highly localized storm swept across the valley. It was accompanied by an exceptional rise in air temperatures and led to complete melting of an unseasonal snow cover. Runoff-triggered debris avalanches caused deep erosion of the benches of surficial debris throughout the valley. About 14 $\times$ $10^6 m^3$ of debris was deposited at the mouth of the Rotbach Torrent, burying a copper smelter and other buildings to a depth of more than 10 m and impounding the Ahr Torrent into a lake that soon flooded the village of St. Martin. Downstream, the people fled from their homes in anticipation of a sudden failure of the unstable debris wedge. On 17 August 1878, the expected flood surged through the communities of Sand-St. Moritz and Kematen, burying houses and rendering fields and crops useless (Daimer, 1879; Lehmann, 1879, p. 43-44). The destructive force of the storm also spilled over the mountain range to the north; there too roads, buildings and people disappeared in the onslaught of the torrential runoff (Sonklar, 1883, p. 107-108).

A concentrated effort to stabilize embankments along the Ahr and its tributary torrents began after 1884. Check dams and reforestation have tranquilized many notorious debris sources, but others continue to cause problems. A severe rainstorm on 30 June 1959, again resulted in a debris flood along the Ahr Torrent which took out bridges and demolished several buildings. In recent years the Ahrn Valley has become a thriving tourist centre hosting modern hotels and recreation facilities.

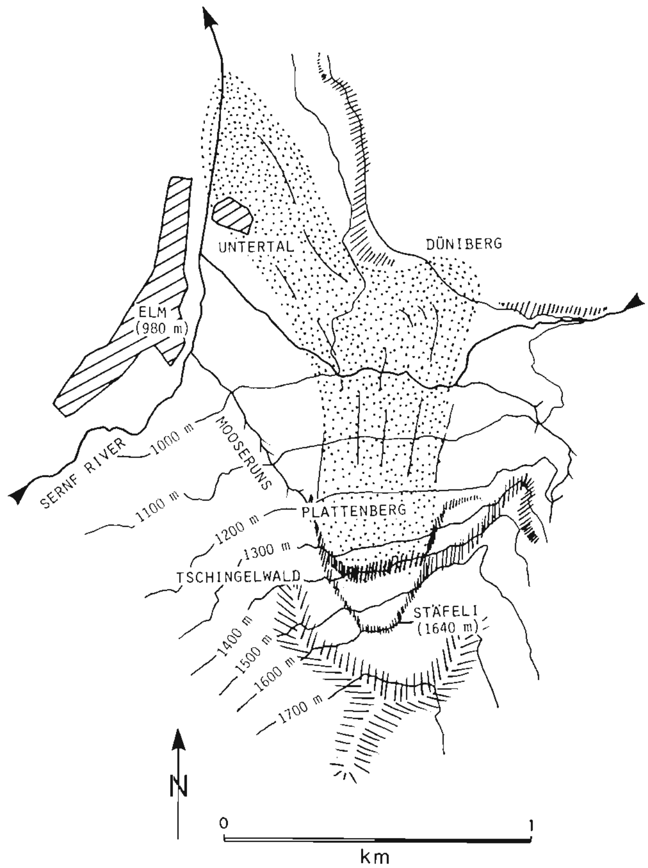


Figure 217: Sketch map of the Tschingelwald scarp slope near Elm and the rock avalanche lobe of 1881.

Elm (A93)

Location: Sernf Valley, Glarus, Switzerland (E2)
Date(s): 11 September 1881

Elm (980 m) is a small village in the north-trending Sernf Valley (Fig. 217). Near Elm the Tschingelwald ridge, composed of Eocene slaty flysch of the Helvetic cover complex, rises along a north-northwesterly facing scarp slope with a general inclination of about 40°. Cleavage and bedding of the slaty rocks dip between 30 and 50° into the mountain and to the south-southeast (Fig. 218).

The catastrophe of Elm in 1881 became famous because the events leading up to and accompanying the failure were carefully documented by Buss and Heim (1881) and Heim (1882a, 1932). Their work still stands among the most valuable descriptions of streaming rock avalanches. It was based on interviews with eyewitnesses and on geological observations. Particularly significant were the observations of Schoolmaster J. Wyss who, watch in hand, timed the initial failure, and of the boy Fridolin Rhyner who ran for his life, barely escaping the streaming rubble.

The failure of the Tschingelwald slope had been partly prepared by natural processes, and partly by the reckless extraction of slate at the Plattenberg quarry (1200 m) near the foot of the cliff. In 1760 a minor failure along the western flank of the Tschingelwald created a steep ravine, the Mooseruns. In 1856 scarplets and minor subsidence were noted in the upper segment of the slope, along a bench known as Stäfeli (1640 m). Then, in 1868 the Plattenberg slate quarry was opened and, after a period of ten years, the quarry extended for a distance of 180 m along the slope. Towards 1879 an increasing number of cracks, oriented parallel to the slope, opened above the quarry and soon rockfalls began to threaten the operation. Nevertheless, extraction of slate continued even after gaping fissures appeared in the working face. Major cracks also developed higher up on the Tschingelwald and in spring 1881 one of these captured runoff from snowmelt. The water which disappeared into the crack drained through the fractured mountainside and reappeared as a spring 40 m below the quarry. In August 1881 a large composite fracture – 'der grosse Chlagg' – opened across the mountain at an elevation of 1500 m. The crack stretched several hundred metres eastward from the head of the Mooseruns ravine and displaced the Tschingelwald slope along a 4-5 m high head scarp.

Between 25 August and 10 September 1881, about 250 mm of rain fell in the Sernftal region. Rockfalls and cracking noises from inside the Tschingelwald at first merely frightened the quarry personnel, but later work was stopped along the quarry face. On 10 September an investigative

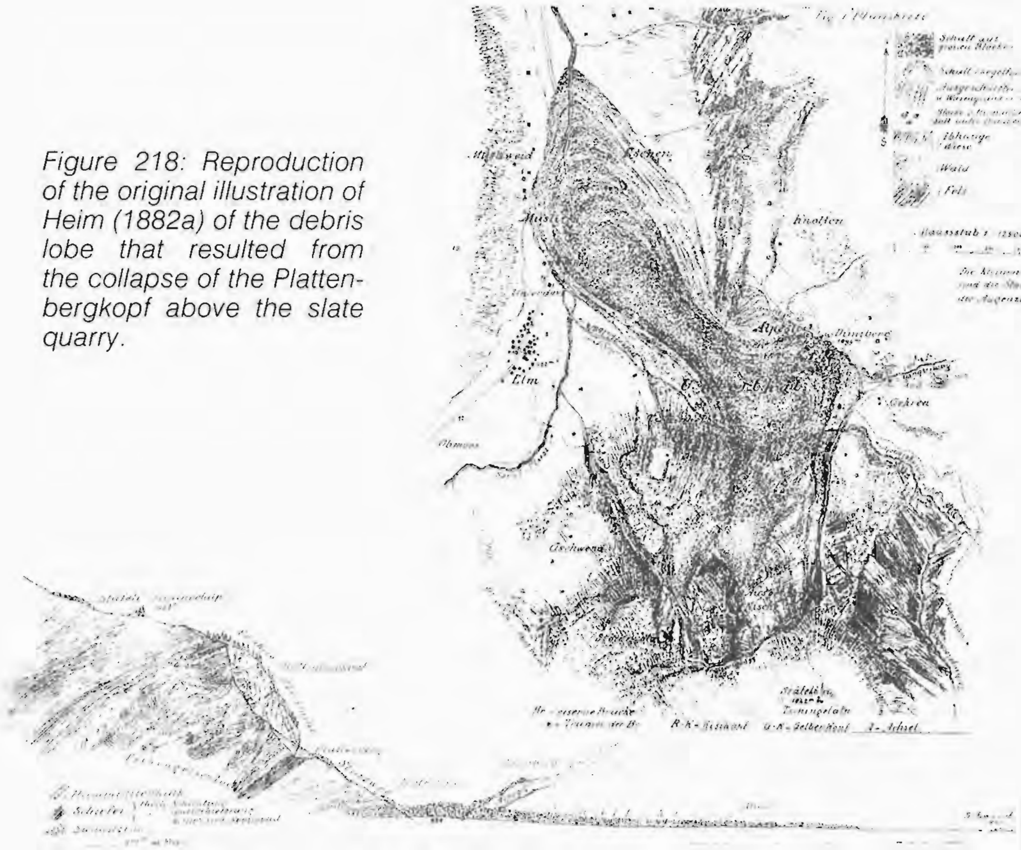


Figure 218: Reproduction of the original illustration of Heim (1882a) of the debris lobe that resulted from the collapse of the Plattenbergkopf above the slate quarry.

commission made up of local authorities climbed the Tschingelwald and noticed cracks, subsidence, tumbled trees, disappearing rivulets, and an almost uninterrupted rockfall activity. Some of the commission members felt that a removal of the tall conifers from the slope might do some good, but the district forester refused to do this and expressed his presentiments of an imminent rockslide. On 11 September 1881, rockfalls and small avalanches continued to break away from the slope and at 4 p.m. rows of trees collapsed backwards into the gaping fissure along the Grosse Chlagg. At 5:15 p.m. a small rock avalanche descended from the east side of the quarry, covered part of the terrace below and catapulted onto the valley floor, burying several buildings that had already been abandoned. At 5:32 p.m., a larger mass of rock came loose along the western margin of the Grosse Chlagg and the resulting rock avalanche buried the hamlet of Untertal and 20 people. The two failures left a downward tapering wedge of slate of approximately $10 \times 10^6 m^3$ without lateral support. Sensing disaster people and animals stampeded away from the Tschingelwald. At 5:36 p.m. the remaining unsupported wedge came loose along the full length of the Grosse Chlagg, now a scarp 500 m long. The wedge of slate first collapsed onto the Plattenberg quarry and then shot across the valley towards the Düniberg. In the words of Schoolmaster J. Wyss:

'I saw the mass first falling vertically and then gushing forth horizontally from the platform of the Plattenberg, whereby the lower and farther projecting part of the mountain was squeezed outward by the pressure of the upper part falling onto it and was thus blasted out into the air (Buss and Heim, 1881, p. 35).'

For a few seconds the Untertal Valley could be seen underneath the cloud of rock and dust as it became airborne at the Plattenberg terrace. In its fall the rock avalanche attained a velocity of at least 80 m/s (Heim, 1932, p. 148) and, according to witness H. Elmer, crashed into some houses 'as if a bowling ball had hit the pins'. The front of the avalanche surged up the Düniberg slope and then turned left. A blast of air and dust preceded the rock stream and pushed a few people to safety. Buss and Heim (1881, p. 143-144) noted from the eyewitness accounts that ....

'this preceding blast of wind acted powerfully only in the direction in which the rock masses shot forward, but that laterally it was entirely insignificant'.

Furthermore,

'it was stronger near the breakaway area in Unterthal than farther away at Müsli .... Along the eastern fringe of the debris stream, between the rock spur of Knollen and the hamlet of Eschen, bales of hay not more than 2 m from the margin of the debris lobe remained undisturbed.'

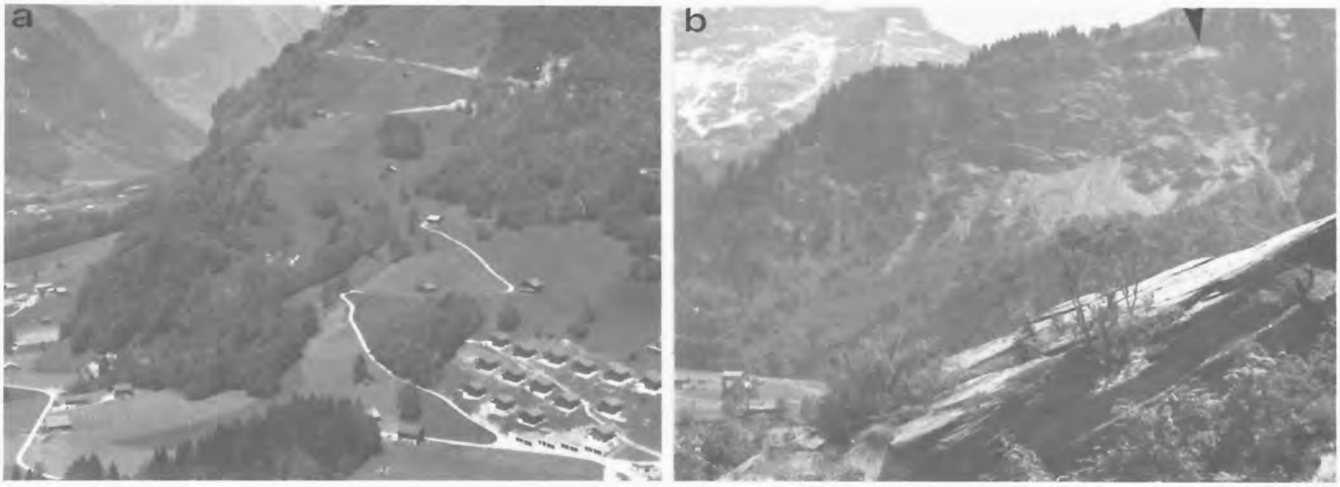


Figure 219: a) View of recently built vacation homes on the Düniberg slope as seen from the former Plattenberg terrace. (GSC 204168-Y) b) View towards the Tschingelwald failure zone (arrow); slate block in the foreground is one of the few that remain from the rock avalanche of 1881. (GSC 204168-Z)

According to witness M. Rhyner-Kubli 'the mass appeared to slide along the ground'. A bituminous smell hung in the air and after the blocky mass had come to a sudden halt the Sernf River was briefly blocked. The catastrophe claimed the lives of 115 people in the community of Elm; most of them had been overtaken by the debris in their desperate run from the cliff.

In the breakaway area a rock spur of approximately $1 \times 10^6 m^3$ remained precariously balanced above the detachment surface whose inclination towards the valley was approximately 50°. Had this unstable mass, the Risikopf, failed as a unit it probably would have resulted in a rock avalanche which might have reached Elm. Between 12 and 16 November 1881, the wide crack that separated the Risikopf from the main wall of the Grosse Chlagg widened noticeably. In December the Swiss army directed a barrage of artillery fire towards the Risikopf to bring it down. Although no bodily collapse was achieved the barrage probably contributed to the eventual disintegration of the rock spur in a series of harmless falls (Heim, 1932, p. 163-164).

Within a few decades the lobe of shattered slate fragments in the valley weathered into a smoothly rolling surface, soon to be reoccupied by fields and roads. Today, a few slabs of slate dot the meadows of Elm (Fig. 219b). On the lower slopes of the Düniberg a colony of bungalows has recently sprung up, facing the scar above the overgrown Plattenberg (Fig. 219a).

The detailed observations of Heim and the conclusions he drew from them have led to important hypotheses towards a better understanding of the baffling reach of dry rock avalanches (Hsü, 1975, 1978).

Bocca di Brenta (A94)

Location: Cima Tosa, Brenta Group, Trentino, Italy (F2)
Date(s): May 1882

The Brenta Group is a rugged mountain chain in the western Trentino. It is composed of flat lying and gently dipping carbonates of the South Alpine cover complex. The Cima Tosa-Bocca di Brenta area consists of well stratified Dolomia Principale which weathers in towers and spires of breathtaking beauty.

During a rainy night in May 1882 a northerly spur of the Cima Tosa (3100 m) collapsed into the valley leading from the Alpe Brenta to the Bocca di Brenta. The failed slab was several hundred metres high and about 6 to $10 \times 10^6 m^3$ in volume. It fell 200 m, crashed onto a shelf, and splintered apart 'like a fluid'. The sheet of debris (mixed with ice and snow?) erased Alpe Brenta (1700 m) and split into three tongues before coming to rest. The 'roar' of the collapse was heard in Madonna di Campiglio, 10 km to the north (Richter, 1885, p. 72-73; Schwinner, 1912, p. 175-177).

1882 Rainstorm (A95)

Location: Saoutheastern Alps (Trentino, Südtirol, Tirol Kärnten)
Date(s): 13 to 17 September and 26 to 29 October 1882

The summer of 1882 was characterized by long periods of rain throughout the Alps. In September the frequency and intensity of precipitation increased markedly because a persistent low-pressure zone had settled over central Europe.

Between 11 and 13 September it rained hard along the southern front ranges of the Alps and snow fell in the higher regions. Then, on 14 September, strong warm winds swept the southern valleys creating a chaotic mix of warm and cold air masses above the mountains, resulting in local cloudbursts and rapid snowmelt. Between 13 and 17 September rainfall in the valleys reached values between 250 and 400 mm. At high elevations slope failures and the gullying of colluvial veneers created unprecedented debris transport in many torrent basins. The rivers were in flood.

Between 26 and 29 October, after a calm spell of four weeks the sequence of thunderstorms, snowfall, southerly winds, snowmelt, and intense rain squalls repeated itself once more. Between 150 and 400 mm of precipitation were recorded in valley communities (Koch, 1883; Sonklar, 1883, p. 108-116).

The impact of debris avalanches, debris flows, floods, and windfall in terms of lost property and human lives was enormous. A concerted effort was made by the Austrian government to assist the affected regions in cleanup operations and in installing new protective and control works along the most dangerous torrents. In 1884 countrywide systematic torrent control became Austrian government policy. This program has been broadened since then.

Toblach-Dobbiaco (A96)

Location: Puster Valley, Südtirol (Alto Adige), Italy (H2)
Date(s): 16 September 1882 (also 1754, 28 May 1821, 15 October 1823, summers of 1856 and 1857, 19 June 1871)

The town of Toblach (1250 m) spreads across the apex of the large debris fan of the Sylvesterbach Torrent on the north side of the Puster Valley (Fig. 220). The Puster Valley parallels a west-trending lineament separating South Alpine basement gneisses on the north from Mesozoic cover rocks on the south. The Sylversterbach fan also creates a major drainage divide between the west-flowing Rienz River and the east-flowing Drau River. The most significant tributary of the Sylvesterbach is the Pfannbach Torrent, which drains the bowl-shaped south slope of the Pfannhorn Massif (2663 m) marking the border between Italy and Austria. The lower Pfannbach basin is fringed by terraces of late Pleistocene ice-margin deposits up to 50 m thick.

During the 18th and early 19th centuries several densely forested catchment basins north of the Puster Valley including the Sylvesterbach were clearcut. Colluvium and Pleistocene ice-margin deposits in these basins soon showed signs of gullying and debris avalanching, and an increase of bedload transport by the torrent was noted. In the Sylvesterbach basin the first destructive debris flows occurred in 1754; they demolished several buildings in Wahlen, a hamlet then located along the torrent channel just above Toblach. Other serious flows followed on 28 May 1821, and on 15 October 1823, when 40 buildings in Toblach were seriously damaged. During heavy cloudbursts in the summers of 1856 and 1857 the remaining buildings of Wahlen disappeared in debris and the entire hamlet was rebuilt on a terrace west of the Sylvesterbach. By this time the torrent channel above Toblach had changed into a wide braided track. On 19 June 1871, during a period of intense rain onto a residual snowpack, masses of debris spread across Toblach. In the source area gullying now had reached such proportions that every minor rainstorm released bouldery flows.

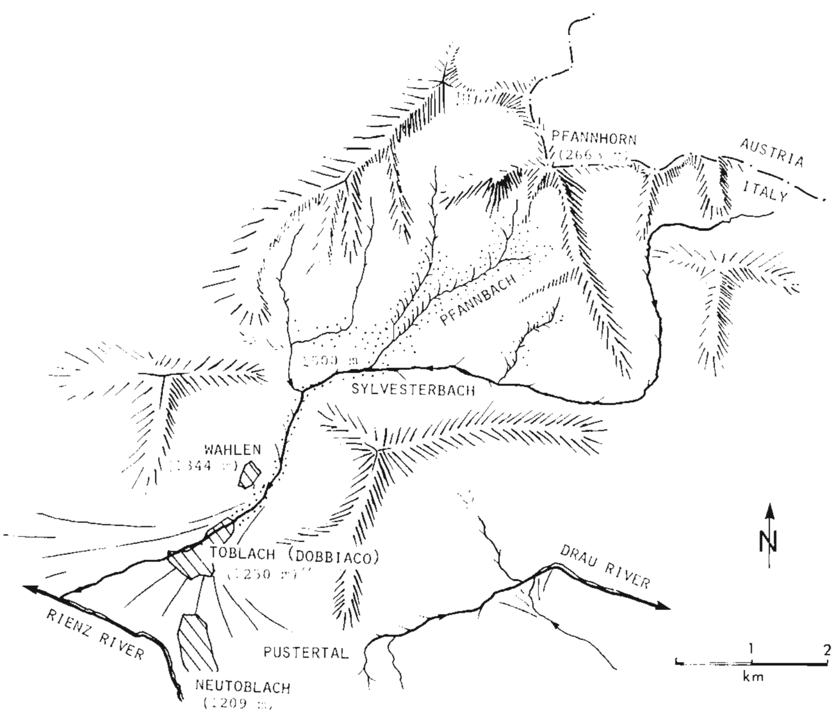


Figure 220: Sketch map of the Sylvesterbach basin, the main debris source area in the Pfannbach tributary catchment, and the fan of Toblach (Dobbiaco).

Between 15 and 17 September 1882 a three-day rainstorm deluged the southeastern Alps. Colluvial embankments along tributary gullies collapsed and pulsating debris flows engulfed Toblach destroying some 40 houses. Other buildings were covered deeply by sand and gravel, but fortunately for the inhabitants they could be reclaimed after the storm. They simply had gained basements they did not have before (Fig. 221a).

Other towns of the Puster Valley did not fare much better than Toblach. Welts of fresh debris converted the valley bottom into a string of lakes submerging all the lowlying land spared by debris flows (Bazing, 1872; Seckendorff, 1884, p. 166-189; Strele, 1899, p. 5; Dalla Torre, 1913, p. 217; Stacul, 1979, p.99).

During the last hundred years a gallant battle has been waged with increasing success to neutralize the debris sources in the Sylvesterbach basin. The lower reach of the Sylvesterbach Torrent has been confined to a lined channel along the western flank of the cone. Near the head of the fan the wide track is bordered by stone-block levees and two strips of protective forest. In the uplands several generations of check dams along deep V-shaped ravines bear witness to the difficult struggle against gullying (Fig. 221b). An attempt is also being made to raise the treeline by systematically planting ecologically appropriate spruce and larch species. Both Toblach and Wahlen continue to grow as prosperous agricultural-tourist communities and some of the recent development has been taken up by Neu-Toblach along the periphery of the cone.

Figure 221: a) House in Toblach which survived the debris floods and flows of the late 19th century; the first floor has been converted into a basement (GSC 204169-A). b) Erosional ravines in late glacial ice-margin deposits in the Pfannbach basin; note check dams (GSC 204169-B).

Altdorf — Spiringen (A97)

Location: Uri, Switzerland (D2)
Date(s): 29 May 1887 (also between 1110 and 1125, 14 June 1910)

The town of Altdorf (500 m) is built on a gently sloping triangular plain bounded by the floodplain of the Reuss River on the west, the debris fan of the Schächen Torrent on the south, and a steep bedrock face on the northeast (Fig. 222). The Schächen Valley follows the strike of gently northdipping Helvetic flysch and carbonate rocks. The settlement of Spiringen (990 m) nestles on a terrace above the northern embankment of the torrent.

Between 1110 and 1125 parts of Bürglen apparently were buried by slides off the south side of the valley (Montandon, 1933, p. 283).

On 29 May 1887, possibly as a result of rapid infiltration of snowmelt, a rock slab approximately $0.5 \times 10^6 m^3$ in volume failed along a steep composite fracture surface in slaty flysch below the summit ridge of Spitzen (2400 m) south of Spiringen (Fig. 223). The disintegrating slab cascaded some 1300 m to the bottom of the Schächental, surged over a hamlet and killed 7 people. The front of the debris stream charged 50 m up the north side of the valley and, for a brief spell, dammed the west-flowing Schächen Torrent (Heim, 1932, p. 122-123).

In June 1910 the mountains of eastern Switzerland and western Austria were still covered deeply by snow when intense squalls of warm rain deluged the Schächen basin. On 14 June 1910, when 100 to 200 mm/24 h were registered in Altdorf, a total of $0.2 \times 10^6 m^3$ of debris was mobilized from the Spitzen rock avalanche deposit and surficial deposits along the Schächen Torrent. An industrial installation south of Altdorf was seriously endangered by this debris. At the same time, a rockfall from the cliffs above Altdorf descended through a chute and crashed into a building, killing its 11 occupants (Heim, 1932, p. 68).

In the aftermath of these events a large number of check dams were erected along the channel of the Schächen Torrent. They have functioned satisfactorily since then, although recently aggradation at the confluence of the torrent and the Reuss River has required modification of the channel works. Since 1972, several protective dams also have been constructed below ravines near Altdorf (Fig. 224).

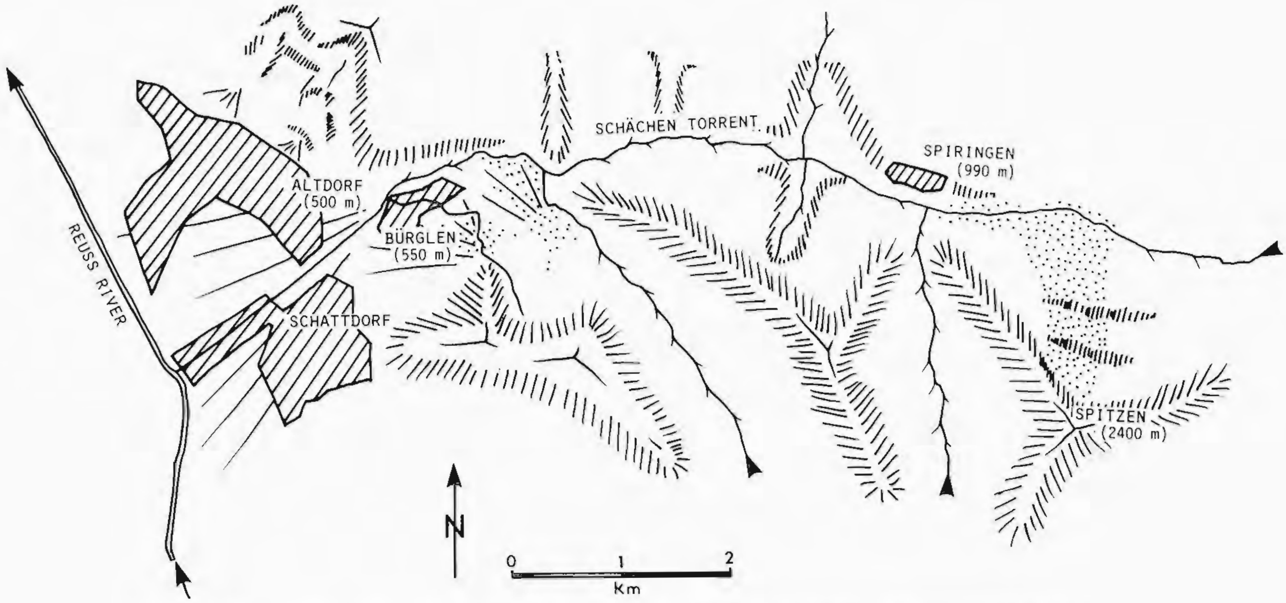


Figure 222: Index map of the lower Schächen Torrent showing the location of the slide deposits of Spiringen and the debris fan that hosts the communities of Altdorf and Schattdorf; steep debris-rockfall chutes occur on the northeastern outskirts of Altdorf.

Figure 223: Detachment zone of the Spitzen rockslide; failure occurred along a stepped composite fracture surface cutting across the bedding. (GSC 204169-C)

Figure 224: Protective dam against debris avalanches under construction near Altdorf; note stone block armour and discharge section designed to drain water from complex debris-snow avalanches. (GSC 204169-D)

Martell Ice Lake (A98)

Location: Martell Valley, Vintschgau (Val Venosta), Südtirol (Alto Adige), Italy (F2)
Date(s): 17 June 1891 (also 1127 (?), 16 June 1888, 5 June 1889, 1 June 1895)

The Martell Valley is a relatively straight and narrow tributary of the Vintschgau (Val Venosta) carved into South Alpine basement rocks (Fig. 225). It is drained by the Plima Torrent which flows from glaciated cirques below the Monte Cevedale Massif (3778 m) to Morter (850 m) in the Vintschgau.

Sudden discharge of water and debris from the head of the basin has always been feared by the people of Morter: as early as 1127 (?) a debris flood practically demolished the town. Debris floods of glacial origin once again threatened the town towards the end of the 19th century.

The Plima Torrent flows from two glaciers in the upper Martell Valley: the Langenferner and the Zufallferner. During the Little Ice Age the two glaciers formed a single terminal moraine a short distance below their confluence. After 1850 both glaciers retreated, the Langenferner at a greater rate than the tributary Zufallferner. During a minor readvance in the 1880s the Zufallferner crossed the Martell Valley bottom and impounded a lake that extended to the snout of the Langenferner (Fig. 225).

On 16 June 1888, the lake drained suddenly through a natural tunnel underneath the ice of the Zufallferner. Picking up loose bouldery debris below the ice dam, the raging flood waters ripped most of the bridges in the valley off their foundations. Fortunately, the roaring sound of the approaching debris flood alerted the inhabitants of the valley and they were able to take refuge on high ground before the frontal wave struck their homes.

On 5 June 1889, a newly formed lake burst again. Approximately $0.6 \times 10^6 m^3$ of water rushed through the subglacial tunnel and the lake was empty within half an hour; again the frontal wave of the flood changed into a debris-laden surge. Damage to structures was twice as serious as during the preceding flood and two lives were lost. After the catastrophe the glaciologists E. Richter and S. Finsterwalder visited the Martell Ice Lake. E. Richter suggested the installation of a signal service and urged the regional government to erect a retention dam below the Zufallferner. Although the funds for this undertaking were approved quickly, local disputes over the supposed mechanisms of the ice floods prevented an immediate start of construction.

In June 1891 the lake, now closely monitored, rose again at a rate of 1 to 2 m a day. Finally, on 17 June 1891, alarm shots were fired at the mouth of the subglacial conduit announcing the burst of some $0.7 \times 10^6 m^3$ of water through the collapsing gap of the ice barrier. In the hamlet of Gand people had just removed their most precious belongings from

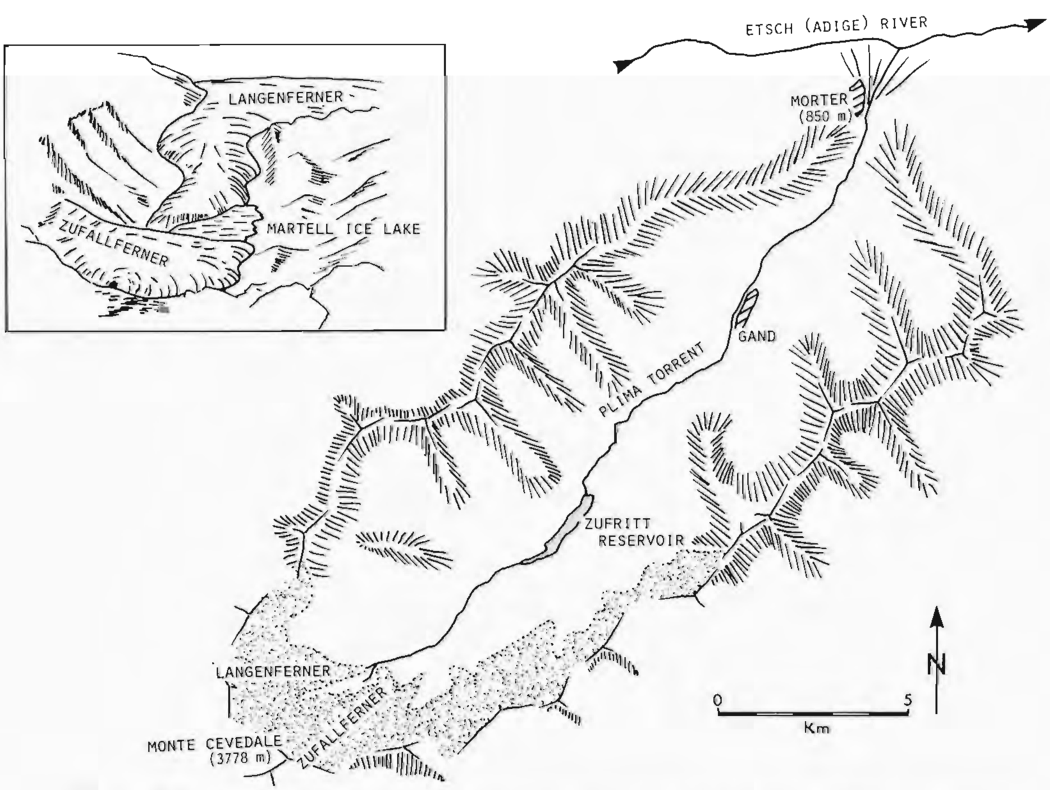


Figure 225: Index map of the Martell Valley showing the location of the Langenferner and Zufallferner. The Martell Ice Lake formed behind the toe of the Zufallferner glacier in the late 19th century.

their homes when the frontal wall of rocks, trees, and water approached the settlement with a frightful roar. The abandoned buildings were completely demolished.

Between 1892 and 1893 the proposed retention dam was built approximately 1 km below the Zufallferner. The dam was 330 m long and had a maximum height of 15 m, allowing for retention of $0.7 \times 10^6 m^3$ of water. A tunnel was blasted through a rock abutment to serve as a spillway. The structure served its purpose during the last burst of the ice lake on 1 June 1895 (Richter, 1889a; Mayr, 1895; Hueber, 1906).

After this series of ice floods the Zufallferner withdrew from the bottom of the valley and the Martell Ice Lake disappeared. Today the Martell Valley has become a major tourist area and the Plima Torrent has been impounded into the hydroelectric Zufritt Reservoir a few kilometres below the historical retention basin.

Kollmann-Colma di Barbiano (A99)

Location: Eisack (Isarco) Valley, Südtirol (Alto Adige), Italy (G2)
Date(s): 17-18 August 1891

Kollmann (490 m) is a small community crowding the steep cone of the Gonderbach Torrent west of the Eisack (Isarco) River (Fig. 226). The two tributaries of the Gonderbach Torrent drain an escarpment east of the Rittner Horn Massif (2260 m) underlain by volcanic bedrock of the South Alpine basement-cover transition. Extensive Pleistocene surficial deposits form embankments along the upper part of the torrent before it drops over the bedrock escarpment at an elevation of 1400 m to the bottom of the Eisack Valley (480 m).

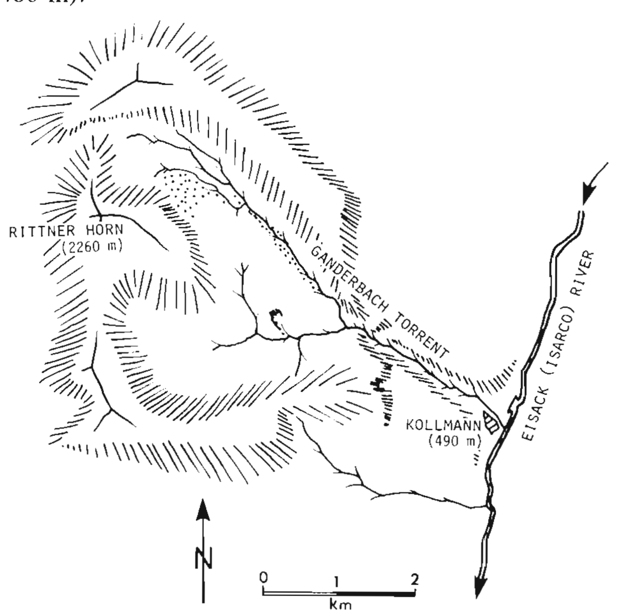


Figure 226: Sketch map of the Gonderbach Torrent which flows through a steep narrow gorge into the Eisack (Isarco) River.

During the night of 17-18 August 1891, a local cloudburst over the east slope of the Rittner Horn set off several debris slides and avalanches which choked the channel of the swollen torrent. Gaining momentum along the bedrock ravine above Kollmann, a major debris flow demolished 20 buildings and killed 39 people in the village. A conical debris blanket, about $0.5 \times 10^6 m^3$ in volume, impounded the Eisack River and created a lake 1.2 km long. About 1000 people worked for several weeks before the river channel was lowered to the point where the Eisack could reoccupy its old course (Strele, 1936, p. 136; Stacul, 1979, p. 90). After the catastrophe Kollmann was rebuilt in the same place. Check dams in the upland and a protective dam above the community have been constructed in recent years.

Blisadona (A100)

Location: Arlberg, Vorarlberg, Austria (F1)
Date(s): 9 July 1892

The Arlberg is a major east-west mountain pass connecting Vorarlberg with the rest of Austria; the region is a popular centre of Alpine skiing. In the Arlberg area Austroalpine metamorphic basement rocks on the south and intensely deformed carbonate cover rocks on the the north form a south-dipping fault zone. Roads, railroad, and settlement development have followed the valley created by erosion along this fault zone. Large debris cones spread south across the valley floor from the mouths of steep ravines cut into carbonate rocks. The cones not only receive sporadic debris flows and rockfalls, but are also regularly swept by snow avalanches.

The high Blisadona ridge (2200 m) north of the Alfenz Torrent (1100 m) consists of a northward-overturned panel of well-bedded carbonate rocks dipping 85° to the south (Fig. 227). Because of the steep inclination (40 to 50°) of the bare bedrock slope, snow avalanche bridges were installed below the Blisadona ridge soon after the railroad had been built. Pictures taken of these protective works in November 1891 also show sections of the bedrock cliff that were to fail less than a year later in July 1892 (Pollack, 1892). A deep east-trending furrow developed some 20 m behind a packet of carbonate strata at an elevation of 2100 m and fresh talus cones below this panel indicated recent rockfall activity.

On 9 July 1892, possibly as a result of snowmelt infiltration, the slab of carbonate strata, 240 m wide, 70 m high, and 15 m thick failed. The photographs taken before the failure indicated that failure was preceded by downward buckling ('Ausbauchen') and outward rotation ('Umkippen') of the lower part. After breaking and shearing along a ragged basal rupture zone, the panel collapsed into a downward tapering ravine filled with avalanche snow. When the rock stream burst onto the cone it spread out in two thin lobes (Fig 228a). The western lobe buried the road and railroad, then banked

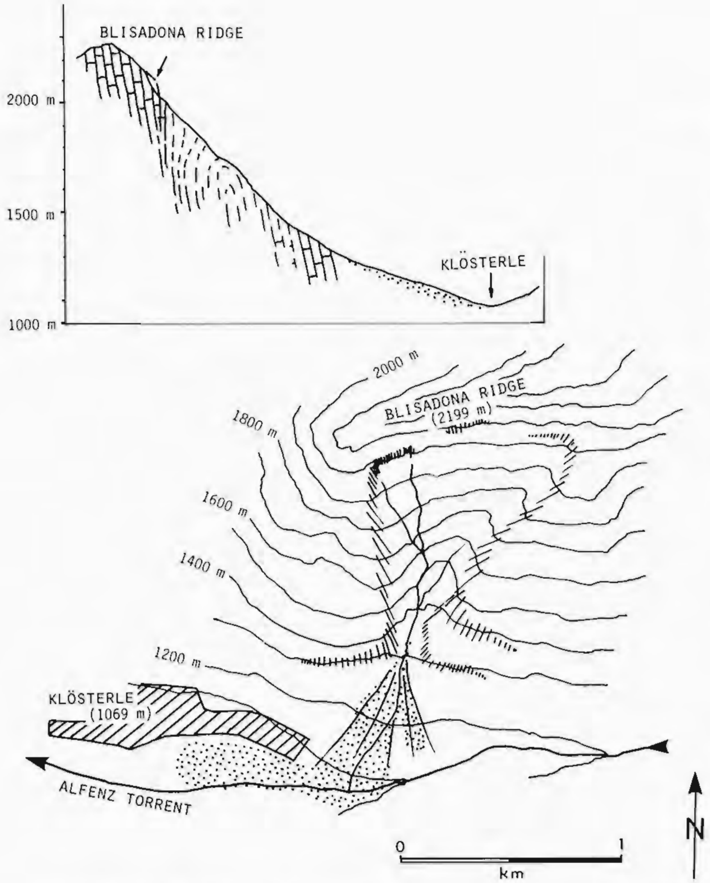


Figure 227: Sketch map and geological cross-section of the Blisadona ridge and the 1892 rock avalanche.

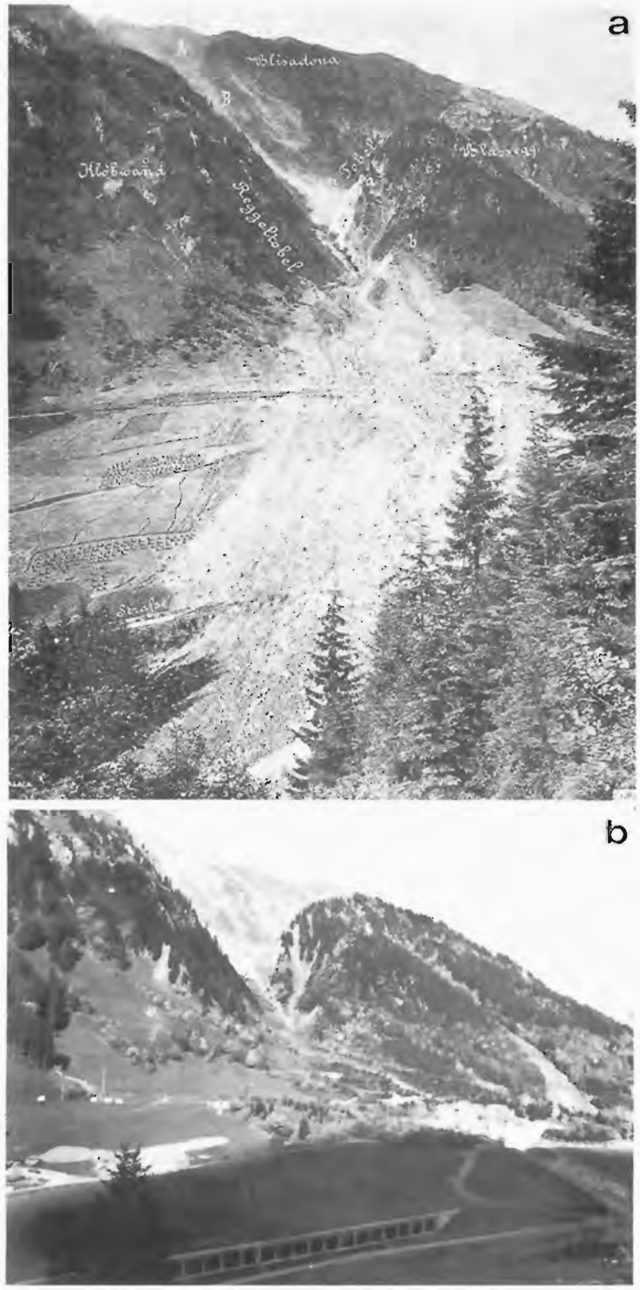


Figure 228: a) Photograph of the Blisadona rock avalanche shortly after the event (from Toula, 1893). b) The Blisadona cone today: a modern highway gallery pierces the debris cone and an armoured channel facilitates the passage of runoff, snow avalanches, and rare debris flows (GSC 204169-E).

30 m up the southern valley wall, turned 90° to the west, and came to rest among the easternmost buildings of Klösterle (1069 m), where it demolished a few houses and killed 2 persons. The small eastern lobe only buried the railroad tracks. By scouring snow and saturated soil the rock avalanche attained a volume of 0.3 to 0.5 × $10^6 m^3$, almost twice that of the initial mass of the failed slab. An airblast preceded the debris stream and toppled trees in its path; trembling of the ground was felt in communities several kilometres away. Linear furrows, up to 20 m deep were carved into the lower cone by the largest carbonate blocks. For hours after the event a tremendous dust cloud hung in the air and the flow of the Alfenz Torrent was stemmed by debris (Pollack, 1892; Toula, 1893; Tiefenthaler, 1973, p. 71-73).

To protect the railroad from similar future events a tunnel was bored beneath the apex of the debris cone. More recently, a modern highway also has been routed via a concrete gallery along the periphery of the cone (Fig. 228b). Klösterle in recent years has grown into a prosperous tourist village. Today the green surface of the Blisadona cone hardly betrays its violent recent history!

St. Gervais — Tête Rousse (A101)

Location: Arve Valley, Haute Savoie, France (B3)
Date(s): 12 July 1892 (13th century)

The Tête Rousse Glacier (3100 m) is a small cirque glacier perched against steep granitic walls above the Bionassay Glacier on the northwestern flank of the Mont Blanc Massif (Dôme du Goûter). It is connected by a glacial torrent with the gorge of Bon Nant (Fig. 229). The Bon Nant Torrent skirts the thermal spa of St. Gervais (700 m) and joins the Arve River on the plains of Le Fayet (581 m). In the 13th century Le Fayet apparently was struck by a sudden debris flood, possibly of glacial origin (Mougin, 1914, p. 372).

During the late spring and early summer of 1892 a large pocket of water accumulated in an ice cavity behind the steep snout of the Tête Rousse Glacier unbeknownst to anybody in the valley. In July 1892 hot weather triggered ice floods throughout the Alps (e.g. on 9 July 1892, one of the largest bursts of Märjelen Ice Lake was registered along the Aletsch Glacier in Switzerland). On 12 July 1892, the front of the Tête Rousse Glacier collapsed, releasing some $0.2 \times 10^6 m^3$ of water, which burst forth onto the boulder-strewn Desert de Pierre Ronde (2700 m). The water then rushed towards the moraine along the right side of the Bionassay Glacier, eroding an estimated $0.9 \times 10^6 m^3$ of blocky granitic debris. Thus an initial spurt of clean water and ice fragments changed into a bouldery debris flow which plunged into the narrow gorge above Bionnassay gaining coherence and momentum. Rising to a height of 30 m above the torrent channel, and glancing off the fringes of Bionnay, the debris flow erased several houses and killed 24 people. With a velocity of 14 m/s the bouldery mass then converged into the channel of the Bon Nant Torrent, rising locally to 60 m above its bottom. This wall of debris struck the baths of St. Gervais with unbridled force. Every building fringing the gorge was buried to the second floor and 78 people died. Farther down, the debris was diverted by a stone bridge towards the village of Le Fayet (581 m) where another 22 houses were demolished before the mass spread out over the fields on the south bank of the Arve River. The descent of the ice-debris flow from Tête Rousse to the Arve River had taken about 30 minutes. It claimed a total of 177 lives (Toula, 1893; Mougin, 1914, p. 372-373).

In 1894 and 1896 there were smaller bursts of water from the Tête Rousse Glacier. To avoid a repetition of the catastrophe of 1892 a 200 m tunnel was driven through rock and ice towards the glacial cavity. When the tunnel reached the cavity no water was encountered. However, the tunnel was maintained and in 1904 approximately 18 000 m^3 of water were diverted harmlessly onto the surface of Bionassay Glacier. With the retreat of the terminus of the Tête Rousse Glacier, the former water-filled cavity became visible and eventually disappeared (Bachmann, 1978, p. 96-97).

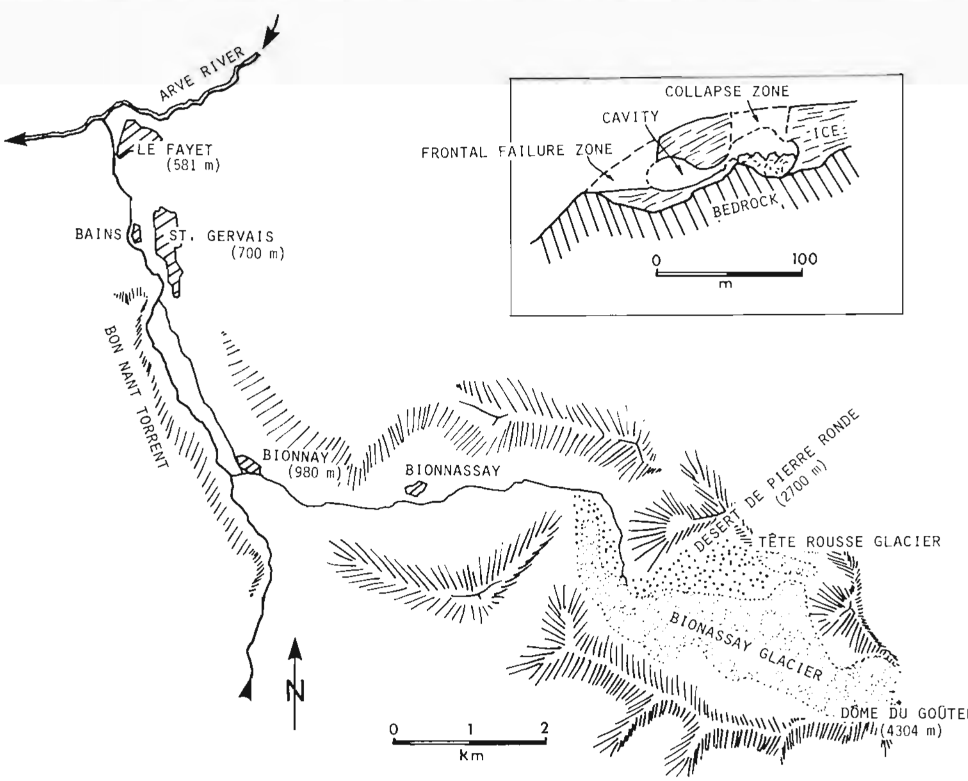


Figure 229: Index map of the Bionassay basin and the small Tête Rousse Glacier. Inset (after Toula, 1893) shows a cross-section of the water-filled cavity and the zones of collapse which led to the ice flood and debris flow of 1892.

Altels (A102)

Location: Gemmi Pass, Bern, Switzerland (C2)
Date(s): 11 September 1895 (also 17 August 1782)

The pyramidal peak of Altels (3629 m) is composed of massive Malm limestone dipping 30 to 32° northwest (Fig. 230). The rugged mountains in the area are part of the Helvetic thrust complex and dip slopes tend to fail as rock avalanches. The dip slope of the Altels is partly covered by a small glacier which during the Little Ice Age attained a considerable size. The stability of the glacier on its smooth and planar bedrock substrate is precarious at best, and at least twice a large part of it slid away from its bed, generating powerful and destructive ice avalanches.

On 17 August 1782, a huge portion of the ice tongue failed and the resulting avalanche killed 4 herdsmen and their cattle on the pastures of Spittelmatte at the base of Altels. An associated air blast knocked down trees and flung some of the animals far up the opposite valley wall. By 1895, when the overhanging ice lobe failed again, memories of the earlier avalanche were dim (Heim, 1895, p. 40-45).

As in 1782, the summers between 1892 and 1895 were characterized by hot, dry weather. In 1895 high summer temperatures persisted throughout August and into the early part of September. At that time the snout of the Altels Glacier dropped over the smooth glacier bed at an elevation of 3000 to 3100 m. Intense melting even at very high altitudes probably contributed to water pressures and discrete slippage at the base of the tongue whose volume amounted to $4.5 \times 10^6 m^3$. Shortly before 11 September 1895, a single crescent-shaped and downward-concave crack reaching bedrock opened across the width of the glacier at an elevation of 3300 m. On 11 September the contact between the upper and lower ice masses was broken. In Heim's (1895) words:

> . . . 'the cleft opens across a width of 580 m, the glacier as a whole begins to slip, within a few seconds its motion accelerates, the separated piece of glacier collapses, turns first into a rough network of seracs, then into a pile of fragments, and now rushes as a stream down the steep incline of rock . . .' (p. 11).

The expanding and disintegrating stream of ice first struck a bedrock terrace where

> . . . 'the avalanche became airborne and leaping forward at a speed of more than 100 m/sec struck the bottom of the valley, compressing the trapped air beneath it' (p. 29).

A powerful airblast radiating out from the foot of the slope carried blocks of ice enveloped in a cloud of smaller ice particles as far as 1 km from the point of impact (Fig. 231). Trees were felled or thoroughly 'sanded' by the ice spray, a small lake was blown dry, a stone wall dislodged, and cattle picked up and hurled 500 to 1000 m horizontally and 250 to 350 m vertically up the Üschenen ridge opposite the Spittelmatte pastures. Similarly, a packet of timbers near a newly constructed cabin was propelled 120 m up the Üschenen ridge. Yet, the ice spray '. . . . did not scour or incorporate underlying soil, but only overrode and covered it . . .' (ibid., p. 21).

The deposits of the main ice avalanche resembled a compact 'ice conglomerate', with well-rounded pieces of ice embedded in a turbid matrix of ice powder. The largest ice fragments did not exceed 1 m^3. Heim concluded that 'the whole avalanche pile is thus a mechanical aggregate in which the process of fragmentation ground its own cementing powder' (ibid., p. 18) and within which

> . . . 'the streaming motion with rotation, sliding, and pushing of the individual parts against each other and against the substratum works like a gigantic grinding machine' (ibid., p. 18).

The main ice avalanche fringed by a rim of stony debris formed an even layer, between 3 and 7 m thick, showing distinctly fanning flow lines. Back fall of the frontal segment doubled the thickness of the lobe below the Üschenen ridge. The catastrophe claimed the lives of 6 people and a large number of cattle.

Today the meadows of Spittelmatte are protected alpine uplands, frequented by hikers who follow the historical Gemmi Pass route between the Swiss cantons of Bern and Wallis.

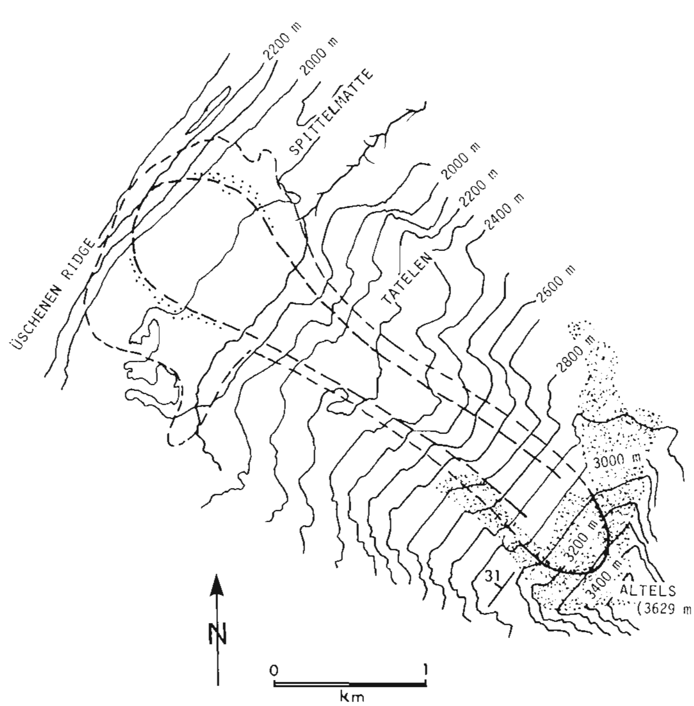


Figure 230: Sketch map of the dip slope pedestal of the Altels Glacier showing detachment zone (barbed line), spray zone (broken outer line), and main depositional area (inner broken line) of the ice avalanche of 1895.

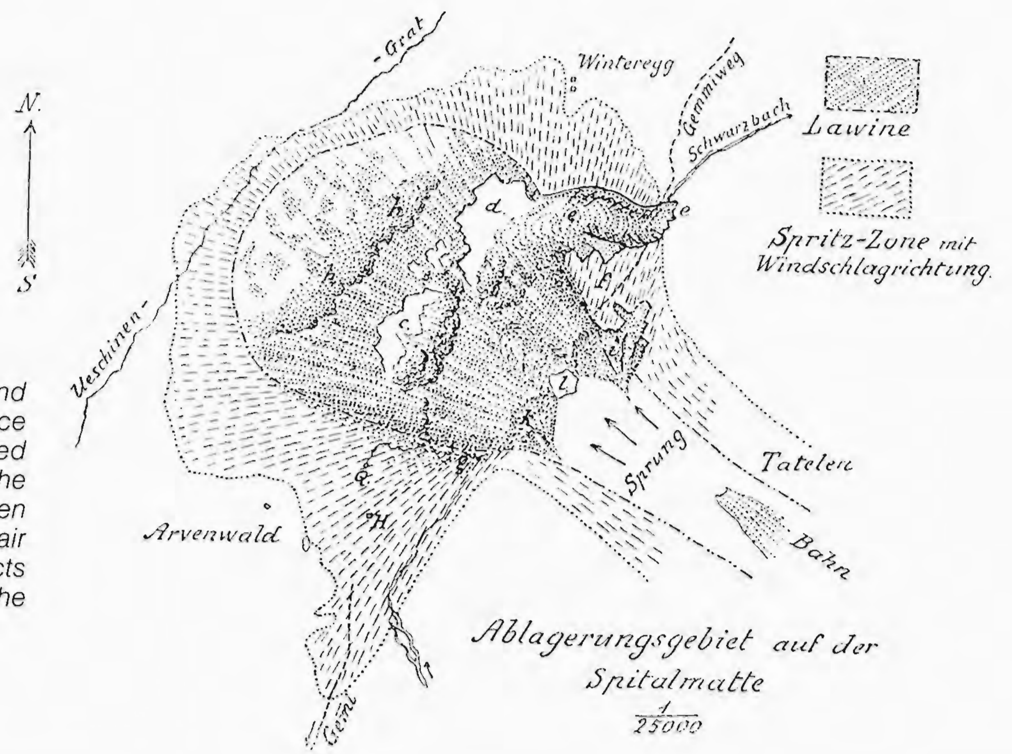


Figure 231: Deposits and spray zone of the Altels ice avalanche as documented by Heim (1895); note the radiating pattern of fallen trees produced by the air blast. Heim also depicts flow lines of the avalanche and backfall ridges (h).

Airolo (A103)

Location: Valle Leventina, Ticino, Switzerland (D2)
Date(s): 28 December 1898

The important railroad and tourist centre of Airolo (1178 m) nestles against the bedrock slope of Sasso Rosso along the south side of the Helvetic Gotthard Massif. The slope is inclined approximately 40° and is underlain by gneiss dipping 50° north into the mountainside.

On 28 September 1898, a small section of the cliff below the summit ridge of Sasso Rosso failed along a spoon-shaped composite fracture surface. An avalanche of loose blocks with a total volume of $0.5 \times 10^6 m^3$ cascaded some 900 m down the mountainside, cutting a swath across the thin protective forest above the town and obliterating 10 houses. Three persons lost their lives (Heim, 1932, p. 121-122).

In this century Airolo has grown and been exposed to considerable snow avalanche hazard (e.g. 1951). Protective stone walls and dams have been built above the town to reduce snow avalanche and rockfall hazards.

Simplon (A104)

Location: Simplon Pass, Wallis (Valais), Switzerland (D2)
Date(s): 19 March 1901 (also 31 August 1597, 1843)

The Simplon Pass route is one of the oldest crossings of the Alps (Fig. 232). Since the earliest days this route between Valle d'Ossola (Italy) and the Wallis (Switzerland) has been exposed to rockfalls, snow avalanches, and icefalls. Almost 200 years ago the first galleries were blasted from the steep walls of gneissic bedrock of the Pennine core zone to make the approaches to the pass somewhat safer. A few settlements cling to debris cones and narrow benches on the sides of the valley.

During the dramatic glacial advances in the final decades of the 16th century the Homattu Glacier, like others in the region, reached far from its bowl-shaped upland basin onto a smooth rock slope southeast of Simplon Pass. On 31 August 1597, the frontal part of the glacier became detached and slid away from its rock pedestal. An avalanche of ice and morainal material annihilated the community of Egga claiming the lives of 80 people (Montandon, 1933, p. 299). A similar ice avalanche from the Homattu Glacier apparently occurred in 1843.

On 19 March 1901, a fractured gneissic spur on the serrated northwestern summit ridge of the Fletschhorn (3993 m) collapsed and a mixture of rock and snow, amounting to $0.8 \times 10^6 m^3$, plunged onto the Rossbode Glacier where it gained enough momentum to scrape up more ice, snow and morainal debris. Growing into a massive avalanche with a total volume of $5 \times 10^6 m^3$ (Montandon, 1933, p. 326), and dropping through a vertical distance of 2300 m, the raging stream of ice and rock engulfed 27 chalets and killed two people (Fig. 233). The front of the blocky lobe came to a halt 200 m from the village of Simplon (1478 m) which was shielded by a late Pleistocene morainal ridge.

Today the bouldery mass north of Simplon is covered by an open forest of larches (Fig. 234) and is known as 'Gletschersturz' (= glacier fall).

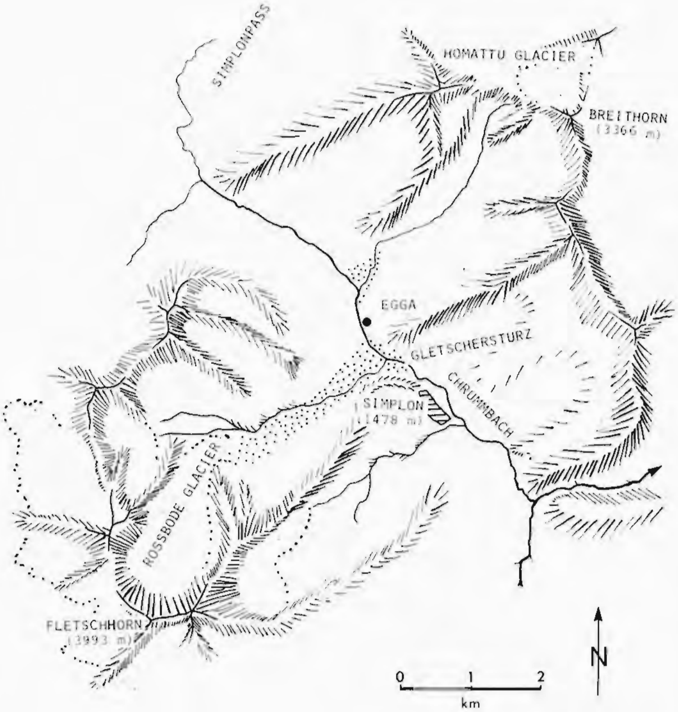


Figure 232: Index map of the Simplon Pass area showing the outline of the 1901 rock avalanche ('Gletschersturz') above the community of Simplon.

Figure 233: Photograph of the Rossbode Glacier and the 1901 rock-ice avalanche deposits shortly after failure (photo courtesy Müller).

Figure 234: a) View onto the larch-covered Gletschersturz in 1981 (GSC 204169-F). b) Gneissic blocks in the frontal part of the rock avalanche deposits (GSC 204169-G).

Bozel (A105)

Location: Isère Valley, Savoie, France (B3)
Date(s): 16 July 1904 (also 1270, 1450, 15 June 1666, 19 June 1669, 14 September 1733, 26 October 1778, 15 June 1818)

The west-flowing Doron de Bozel drains a large basin underlain by northeast-trending sedimentary and metamorphic basement rocks of the Helvetic (Dauphinois) and Pennine (Briançonnais) facies complexes. The main tributaries of the torrent collect runoff from partly glaciated catchments of the Vanoise region, where peaks attain heights between 3000 and 3800 m (Fig. 235). In its lower section the torrent flows in a gorge of highly unstable evaporitic rocks ('Faisceau de Salins') before it discharges into the Isère River at Moutiers (500 m). In the remote past the small triangular plain of Moutiers suffered repeatedly from bursts of debris which left their legacy in several levels of graves in the old graveyard of the town. Debris flows along the Doron de Bozel are generally related to mass movements which directly affected communities in the Bozel Valley (Salins-les-Thermes, Bozel, Villard-de-Plana). In the past slide-prone upland slopes deteriorated mainly due to excessive clearcuts serving medieval salterns in the area (Mougin, 1914).

The village of Bozel (840 m) is located on a large debris cone at the mouth of the Bonrieu de Bozel which drains uplands underlain by sagging slate terrain and thick colluvium. The earliest recorded damaging debris flows in Bozel occurred in 1270. The town was battered again on 15 June 1666, and on 19 June 1669, and on several occasions during the following century. However, the layout of the village remained the same. Then on 16 July 1904, a local intense cloudburst-hailstorm once again triggered a powerful debris flow down the Bonrieu de Bozel, and as in days bygone, masses of bouldery debris crashed into the buildings and bridges of Bozel; this time 11 persons lost their lives (Mougin, 1914, p. 779, 1931, p. 19-20).

The narrow gorge above Moutiers posed another problem. In 1450 a spur of evaporitic bedrock, with a volume of only 70 000 m^3, collapsed along the western wall of the Doron de Bozel gorge, across from Salins-les-Thermes. It knocked over buildings, buried the valuable salt springs, and formed a temporary debris dam across the channel of the torrent. When this dam burst a debris flood damaged large sections of Moutiers. During the great regional rainstorm of 14 September 1733, the gorge of Salins-les-Thermes again was the site of numerous embankment failures which caused severe damage to buildings.

A third threat to the valley were debris floods originating from bursts of the glacial Lac de Glière which was located above the gorge of the tributary Doron de Champagny. On 26 October 1778, rainstorm runoff, probably compounded by overflow from the glacial lake, created a debris flood along the Doron de Champagny which demolished 18 buildings at Villard-de-Plana (Mougin, 1931, p. 19-20). The threat from the glacier-impounded lake became particularly serious in the early 19th century. In 1818 ice avalanches from the steep north-facing snout of the Glacier de Lépéna built a cone of remolded ice across the channel of the torrent; the lake behind it soon attained a volume of 3.7 $\times$ 10^{6}m^3. On 15 June 1818, the ice dam failed and a wall of water, ice, and debris devastated fields and settlements all the way to Moutiers. Minor outbursts of the lake recurred until 1847 (Rabot, 1910) when the Glacier de Lépéna retreated.

In this century reforestation has improved slope stability in many parts of the basin. The town of Bozel now covers most of the debris cone of the Bonrieu de Bozel which now flows in a narrow masonry-lined channel. Slumps and rockfalls continue to threaten the principal transportation routes to new ski centres (Courchevel, Méribel) and to the uplands of Vanoise National Park (Antoine et al., 1979).

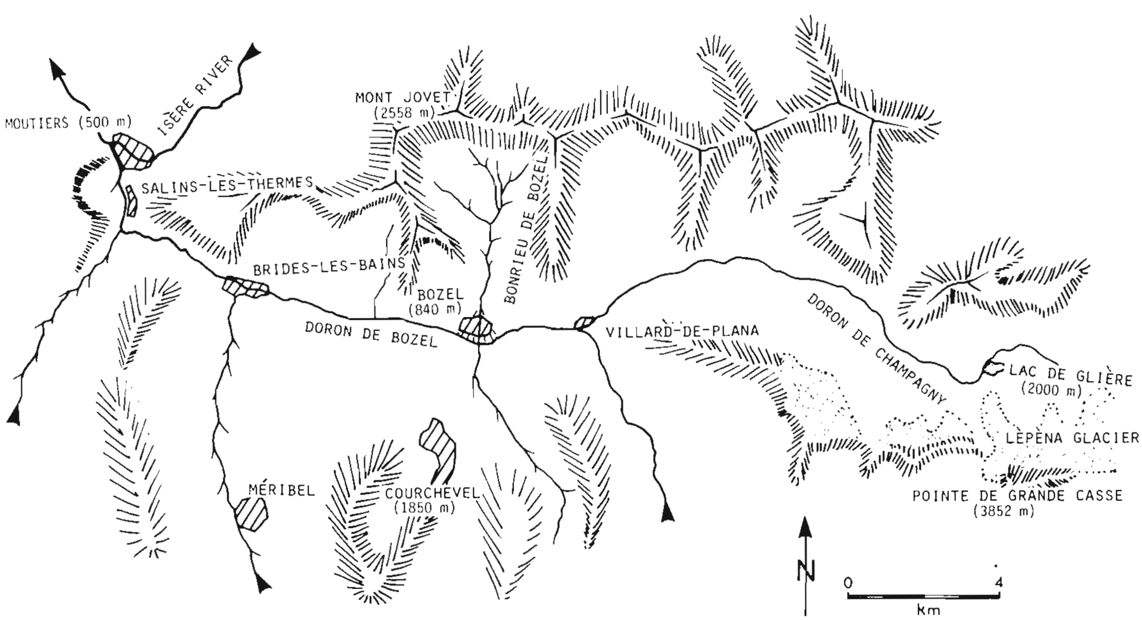


Figure 235: Index map of the Doron de Bozel drainage basin; note the completely built-over cone of Bozel.

Ziller Valley (A106)

Location: Tirol, Austria (G1)
Date(s): 29 July 1908 (also 21 July 1559, 14 July 1887, 13 July 1891, 15 July 1945, July 1946, 30 May 1968, 20 August 1974)

The northern Ziller Valley (530 m) in the Hohe Tauern Range is drained by the north-flowing Ziller River (Fig. 236). It has an extremely low longitudinal gradient (1:330), and the channel of the river is confined by large debris cones. The crests of tributary basins, between 1500 and 2200 m elevation, are underlain by north-dipping low grade metamorphic rocks of the Pennine Tauern Window and the Austroalpine basement complex. Thick deposits of late Pleistocene colluvium locally fringe the torrent channels. The debris cones and fans of the Ziller Valley host sizable communities which thus avoid the notoriously flood-prone and swampy valley bottom. Debris movements in the torrent basins are commonly triggered by midsummer thunderstorms. Dyking systems against debris flows were built as early as 1435 (Hanausek, 1975, p. 407).

On 21 July 1559, a series of four flows down the Oxlbach Torrent almost completely destroyed the community of Schlitters. On 14 July 1887, several pulses of debris totalling $0.4 \times 10^6 m^3$ and originating by extensive undercutting of colluvial embankments along the Riedbach Torrent flattened six buildings. On 13 July 1891 debris flows down the Kaltenbach Torrent burst forth in great force, covering several buildings.

On 29 July 1908, a severe thunderstorm-cloudburst deluged the eastern slope of the northern Ziller Valley. Up to 20 cm of hail and uprooted trees were washed into swelling torrents already blocked by embankment failures and debris avalanches. A dozen debris cones received destructive flows of bouldery debris, including the Niederharterbach ($0.3 \times 10^6 m^3$), the Haselbach ($0.2 \times 10^6 m^3$), and Riedbach basins. Great damage resulted from the blockage of the Ziller River by masses of timber and mud. Most of the valley bottom resembled a lake. On the cone of the Niederharterbach 11 people were killed by one of the debris flows. Cleanup work was hampered by other rainstorms that followed later in the summer. The careful observations of J. Stini relating to the mechanisms by which slumps and avalanches of debris change into devastating flows were the basis for his often-cited monograph on debris flows (Stini, 1910).

Beginning in 1931, but particularly during the twenty years after 1938, extensive clearcuts denuded the colluvial slopes of the upper Finsing basin. Snow avalanches opened the first scars in the logged-off ravines and set the stage for later debris movements which culminated in the destruction of buildings and roads during cloudbursts on 15 July 1945, and in July 1946 (Schiechtl, 1954).

On 30 May 1968, a local downpour dumped about 40 mm of rain within an hour onto the upland basin of the Oxlbach Torrent, setting off a debris flow that covered part of Schlitters with several metres of debris; six years later, on 20 August 1974, a similar cloudburst produced 60 mm within an hour and mobilized 18 000 m^3 of debris; three buildings in Schlitters were buried (Hanausek, 1975).

Today communities in the northern Ziller Valley are growing rapidly; thousands of tourists pass through the valley or spend part of the summer in it. Modern protective steelbeam-concrete dams and arrays of check dams have been built along the principal torrents. A comprehensive program of reforestation on unstable torrent embankments and sagging slopes has been carried out to protect the extremely valuable land on the debris cones. The Ziller River has been dyked along its entire length.

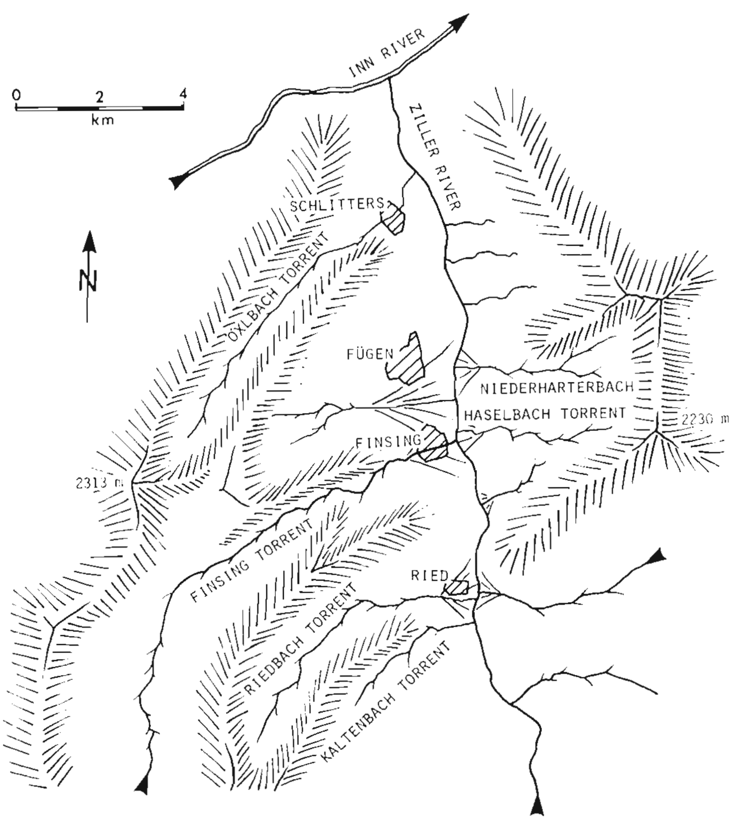


Figure 236: Index map of the northern Ziller Valley; note clusters of settlements at the apices of major debris cones.

Pra — Lagunàz (A107)

Location: Valle di San Lucano, Veneto, Italy (G2)
Date(s): 3 December 1908

Valle di San Lucano, drained by the east-flowing Tegnas Torrent, is a narrow tributary valley in the Cordèvole basin. It follows the crest of a broad anticline in carbonates of the South Alpine cover complex. Steep rockfall cones rise into sheer walls of massive dolomite dissected by funnel-shaped ravines. Along the torrent dense forest has been cleared only in a few spots, one of which used to host the twin communities of Pra and Lagunàz (800 m). Behind the narrow flats which accommodated the hamlets, the south wall of Cima d'Ambrosogn soars to a height of 2266 m (Fig. 237).

In the early 19th century one of the fractures parallel to the cliff face opened into a gaping fissure below the summit of Cima d'Ambrosogn. In 1867, concern over a possible rupture along this crack led to plans to relocate the communities of Pra and Lagunàz to a site in the Cordèvole Valley, several kilometres to the east. However, just about this time a rockfall occurred at the projected new site and relocation schemes were abandoned. During the following decades the people of Pra and Lagunàz were kept alert of the threat above their heads by sporadic rockfalls which, however, caused 'more noise than damage', at least until December 1908 (Bibolini, 1909, p. 64).

The late autumn months of 1908 were characterized by great daily temperature fluctuations. One week before the main disaster a rockfall from the Cima d'Ambrosogn came to a halt just above Lagunàz and a few days later a similar rockfall almost reached the buildings in Pra. Finally, at midnight on 3 December 1908, a wedge of about 10 000 m^3 of carbonate rock splintered away at an elevation of 2000 m and cascaded through two steep ravines onto the talus cones at Pra and Lagunàz, 1200 m below. Due to the height of fall, the steep gradient (60°), and the effect of channeling in the ravines, the carbonate blocks scattered wildly and obliterated several buildings (Fig. 237, 238). An air blast unroofed buildings and toppled trees adjacent to the rock avalanche path and a pall of dust hung over the valley for hours after the avalanche had come to rest; 28 persons lost their lives (Bibolini, 1909).

Smaller recurrent rockfalls from the same source occurred on 18 and 19 June 1911 (Schwinner, 1912, p. 196).

Like many other mountain settlements in the region, Pra and Lagunàz were abandoned. The ruins of the old buildings are crumbling beneath an encroaching forest of alders (Fig. 238). In 1959 a stone memorial to the victims of the catastrophe was erected at the former site of Pra. New, tourist-oriented buildings have since been built or are being renovated at Col di Pra, 2 km to the west.

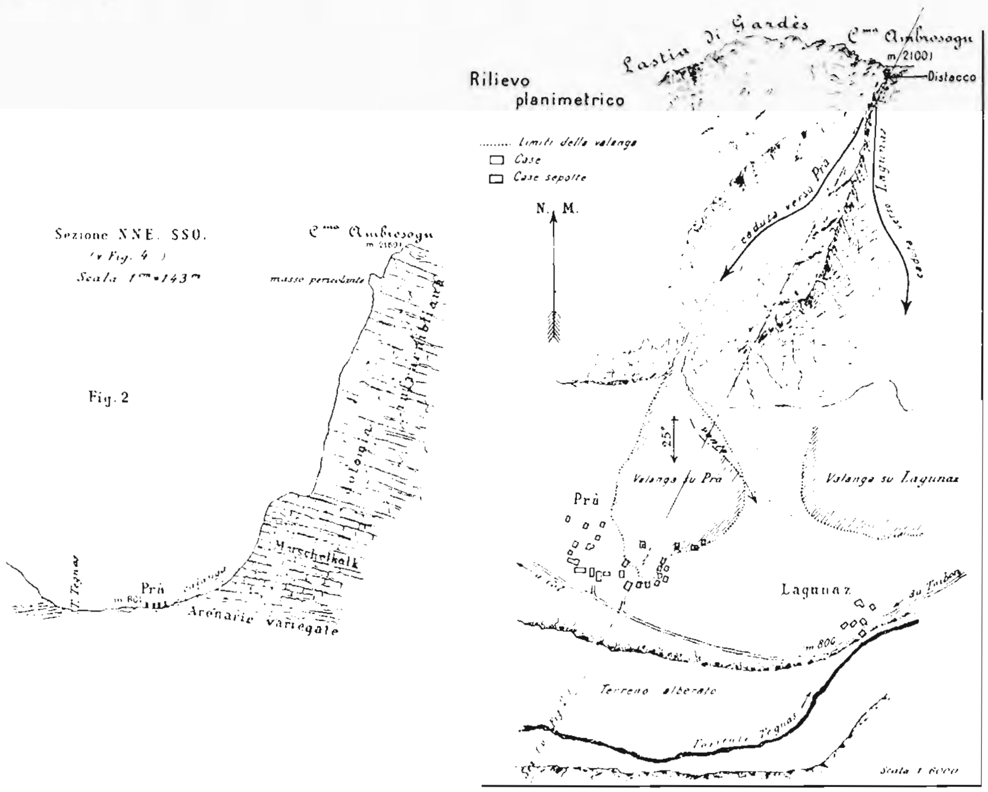


Figure 237: Map and geological cross-section of Cima d'Ambrosogn, showing the tracks and depositional lobes of the 1908 rock avalanche at Pra and Lagunàz (from Bibolini, 1909).

Figure 238: Blocks of the rock avalanche of 1908 (foreground) scattered among the ruins of Pra; the carbonate cliff of Cima d'Ambrosogn in the background (photograph taken in 1980). (GSC 204169-H)

Vandans (A108)

Location: Blundenz, Vorarlberg, Austria (E1)
Date(s): 15 June 1910 (also 1762, 1764, 29 May 1894, 12 August 1933, August 1967)

Vandans (650 m) spreads over an inclined terrace along the south bank of the Ill River which here flows northwesterly from metamorphic basement terrain through a complex of Austroalpine cover rocks (Fig. 239). The stretch of the valley between Vandans and St. Anton is dominated by coalescing debris cones of the Rellsbach, Mustrigil, and Venser torrents on the south and the cone of the Gipsbach Torrent on the north. The upland basins of most of these torrents are underlain by northeast-striking and northwest-dipping evaporitic shales overlain by dolomite cliffs rising to crests of 2400 m elevation. The Rellsbach basin contains sagging slopes of metamorphic basement mantled by thick relict colluvium.

In remote times, the dolomitic crest of the Gipsbach basin collapsed and probably demolished a settlement on the huge cone north of St. Anton; debris possibly impounded the Ill River into a sizable lake. This area has since been avoided by settlement.

In 1762 and 1764 Vandans was damaged by a series of debris flows caused by failures along the scarp face of the Vandanser Steinwand (2443 m), with the debris channeled down the gorge of the Mustrigil Torrent. On 29 May 1894, a fractured mass of carbonate about $2 \times 10^6 m^3$ in volume failed at the eastern end of the Vandanser Steinwand and, mobilized by runoff during a furious rainstorm, proceeded to stem the flow of the Ill River. A lake which became an attraction throughout the summer eventually burst across the debris wedge and created serious damage downstream.

On 14 June 1910, during a regional rainstorm that caused widespread flooding in Vorarlberg and Switzerland, about 230 mm of rain fell in the vicinity of Vandans within a period of 24 hours. Massive debris transport occurred along all the torrents above Vandans. Along the Rellsbach Torrent embankment failures along the toe zones of sagging slopes triggered debris flows that burst into Vandans, burying 30 buildings underneath a total $1.4 \times 10^6 m^3$ of rubble (Strele, 1936, p. 136-137).

On 12 August 1933, runoff from a series of thunderstorms mobilized debris that had accumulated below the scarp face of the Vandanser Steinwand; a rapid debris flow, with a volume of more than $0.1 \times 10^6 m^3$, crashed into 9 buildings in Vandans, killing 5 people (Schiermböck and Geyer, 1974). In August 1967 the Rellsbach Torrent again carried a flow with a volume of $0.1 \times 10^6 m^3$ which came to rest in a debris retention basin.

After these and other accidents a large number of check dams and protective works (transverse dams, dykes, deflection walls, and retention basins) were built in the vicinity of Vandans. On account of an extensive hydroelectic development and booming tourism, the community has greatly expanded in the last 30 years. Occasional rockfalls from the steep scarp face are obviously beyond human control, but the condition of the cliff is being closely monitored by regular inspection.

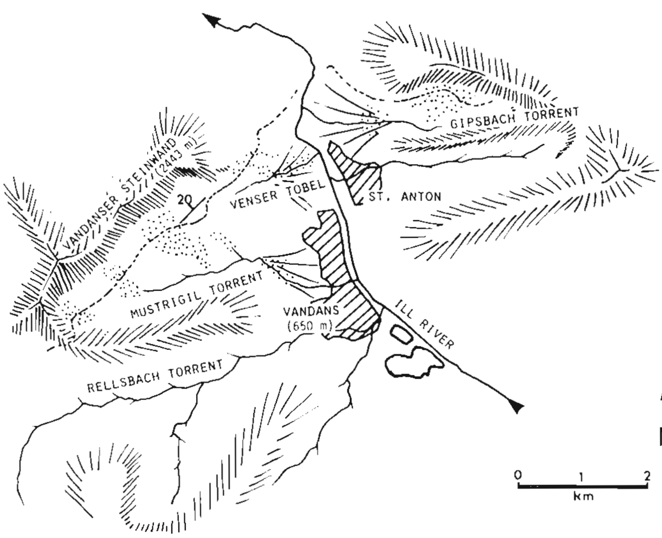


Figure 239: Sketch map of the Ill River near Vandans. Broken line indicates outcrop of evaporitic shales.

Sandling (A109)

Location: Salzkammergut, Oberösterreich (Upper Austria), Austria (II)
Date(s): 12 September 1920 (also 1560)

The Sandling Massif (1717 m) is part of the front ranges of the eastern Alps. Underlain by carbonate strata of the Austroalpine cover complex that dip about 20° to the east the southwestern scarp face of the Sandling exposes a massive upper limestone cliff and a lower more recessive calcareous flysch and evaporitic shale (Fig. 240). The escarpment is mantled by talus.

In the Middle Ages much of the unstable south flank of the mountain was dotted by adits to underground salt mines. In 1560 a slope above one of the adits failed and a number of miners were buried by the slide debris (Lehmann, 1926, p. 271).

During late August and early September 1920 a heavy rainstorm pounded the front ranges of the Alps, setting off floods in the Salzkammergut region. On 12 September one week after the rainstorm, several people who stayed in cabins on the gently sloping meadows below the southwest face of the Sandling observed that one of its isolated rock towers, the Pulverhörndl, was surrounded by a dust cloud due to almost continuous rockfalls. One person who had climbed the peak early in the day had noticed that the talus cone below the Pulverhörndl was shifting slowly towards the southwest. It soon became apparent to the onlookers that the rock spur, $0.2 \times 10^6 m^3$ in volume, was gradually moving away from the main face of the Sandling along a vertical fracture trending north-northeasterly. Accompanied by rockfalls and outward flow of shale below the tower this process of detachment continued for more than twelve hours. When trees and huts of the alpine meadows began to shift as well, the people abandoned the area. Shortly thereafter, the main part of the tower crumbled completely and a lobe of blocky carbonate spread over the meadows which were underlain by water-saturated shale and surficial deposits. The frontal segment of the overloaded area began to break up into slumps which combined into a downward pressing flow of mixed shale, moraine, and rock avalanche deposits. The initial rate of displacement at the front of this flow was 11 m/s; eventually a mass of 6 to $9 \times 10^6 m^3$ moved slowly down the tributary ravine destroying valuable forest land (Lehmann, 1926). Upon reaching the Leisling Torrent the toe of the debris lobe was reworked and transported to the cone of the torrent. During the years since then the flow slowed down and today only a small segment seems to be active. Check dams along the lower reaches of the Leisling Torrent have kept erosion of rubble along the channel within tolerable limits. Several deep cracks along the face of the Sandling indicate that not all of the Pulverhörndl was carried away in 1920 (Fig. 241).

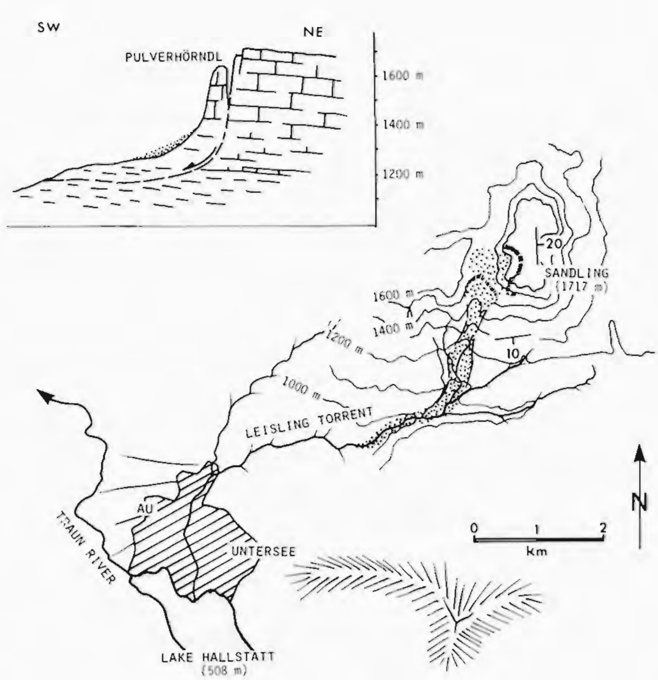


Figure 240: Sketch map and schematic cross-section of the Sandling Massif; note the overloaded shale terrace below the Pulverhörndl tower and the deposits of the slow-moving debris flow of 1920.

Figure 241: Remnants of the collapsed Pulverhörndl tower along the fracture-controlled west face of the Sandling Massif. (GSC 204169-I)

Roquebillière (A110)

Location: Moyenne Vésubie, Alpes Maritimes, France (C4)
Date(s): 25 November 1926 (also 1094, 12th century, 20 July 1564, 22 February 1742)

The Moyenne Vésubie is part of the moutainous region separating the crystalline Massif de Mercantour from the Mediterranean Sea. The ranges are underlain by sedimentary successions of the Helvetic (Dauphinois) cover complex and rise to ridges 1000 to 1500 m in elevation. The bedrock consists of shale, carbonate and flysch. Near Roquebillière the Vésubie River follows a north-trending zone of intensely deformed evaporitic shales (Fig. 242). The narrow floodplain of the river is bordered by notoriously unstable terraced slopes hosting small agricultural communities. Roquebillière is only one of many settlements in the area that have suffered from mass movements.

The earliest community of Roquebillière probably spread along the west bank of the Vésubie River; it was almost completely swept away by a debris flood in 1094. The survivors of this catastrophe moved to a new site on the east bank of the river near the debris cone of the Gordolasque Torrent. Sometime in the 12th century the Gordolasque Torrent debouched enormous amounts of debris onto its cone, forcing a relocation of the village onto a river terrace below the colluvial bench of Belvédère (800 m). On 20 July 1564, an earthquake accompanied by landsliding destroyed much of the community and killed scores of people. The village grew back to reasonable size, when, on 22 February 1742, another debris flood of the Vésubie wrecked two-thirds of the buildings (Mougin, 1931, p. 99-100).

For centuries the benches above Roquebillière were irrigated by farmers working the fields. Minor slumps along the outer rim of the terraces occurred frequently but posed problems only during major rainstorms. However, a series of rainstorms in the autumn of 1926 led to a marked deterioration of the entire shale slope behind Roquebillière. Precipitation in the Moyenne Vésubie between 21 October and 23 November was as much as 1700 mm, seven times the normal; in the five days between 19 and 23 November a slide with a volume of $3 \times 10^6 m^3$ temporarily closed the gorge of the Vésubie above Roquebillière; fortunately overflow of the river occurred before a major lake had formed. However, the most destructive landslide triggered by the rains was yet to strike Roquebillière. On 23 November 1926, a large crown crack appeared along the rim of the Belvédère terrace and on 25 November 1926, at 3 a.m., a slab of evaporitic shale, 150 m wide and $0.2 \times 10^6 m^3$ in volume slid away along a gently inclined basal shear surface. Its front ploughed into the southern section of the village destroying 20 houses and killing 19 people. The slide mass came to rest but was remobilized by more rain on 29 November 1926. Advancing at a rate of 5 m per day the reactivated flow of remolded shale demolished another 10 buildings before winter temperatures slowed its motion. However, large new cracks developed behind the head scarp and the village was partly evacuated. The remaining group of buildings — now known as Roquebillière Vieux has continued to be in a zone of recognized risk (Bénévent and Maury, 1927; Bertrand, 1927; Meneroud and Calvino, 1976). Recent development in the area has shifted to a bench above the western bank of the Vésubie River where the modern community of Roquebillière is located.

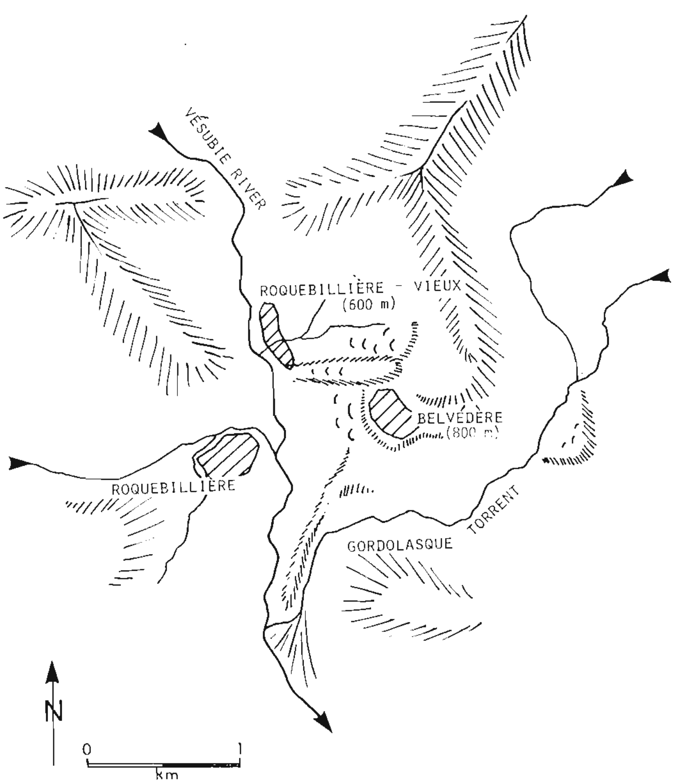


Figure 242: Index map of the surroundings of Roquebillière; note the historical slump area on the terrace of Belvédère.

Klausen-Chiusa (A111)

Location: Eisack (Isarco) Valley, Südtirol (Alto Adige), Italy (G2)
Date(s): 9 August 1921

The town of Klausen (550 m) crowds the small fan of the Tinnebach Torrent on the west side of the narrow Eisack (Isarco) Valley (Fig. 243). The catchment basin of the torrent is a large bowl-shaped plateau on which tributary ravines are incised into phyllitic bedrock of the South Alpine basement and Pleistocene surficial deposits. Midsummer thunderstorms and downpours tend to produce erosional scars in otherwise inconspicuous ravines. The steep slopes below the main escarpment are also notoriously unstable. A local legend tells of a settlement named Schönberg (=pretty mountain) which disappeared after a rainstorm, was later rebuilt in the same place, and received the fitting name Villanders (=quite different); this community still clings to the mountainside south of Klausen (Dalla Torre, 1913, p. 253).

On 9 August 1921, an intense cloudburst-hailstorm raged in the Tinnebach basin for approximately three hours. During this storm scour along numerous tributary ravines produced debris that collected in the gorge. From there a wall of boulders, mud, and trees advanced to the narrow mouth above Klausen at a rate of 2 m/s. The first pulse of debris cut a broad swath across the town and killed four people. Subsequent pulses of debris filled the channel of the Eisack River; a lake 7 m deep flooded most of the buildings that had escaped direct damage from the flows. This lake persisted for several days.

Since this disaster several protective dams, 10 m high, have been erected across the mouth of the gorge above Klausen. A large number of check dams now neutralize the most serious debris sources in the uplands; one of these debris chutes contains as many as 150 closely spaced check dams (Pitra, 1921; Stacul, 1979).

Motto d'Arbino (A112)

Location: Val d'Arbedo, Bellinzona, Ticino, Switzerland (E2)
Dates(s): 2 and 28 October 1928 (also 18 October 1915)

The broad summit ridge of Motto d'Arbino (1694 m), east of the lower Ticino Valley, is composed of gneiss and marble of the Pennine core zone of the Swiss Alps (Fig. 244). The foliation of the metamorphic rocks, which dips 80° north, controls the course of the Traversagna Torrent along the north side of the mountain. The torrent joins the Ticino River after crossing a debris cone on which the communities of Arbedo (270 m) and Molinazzo are located.

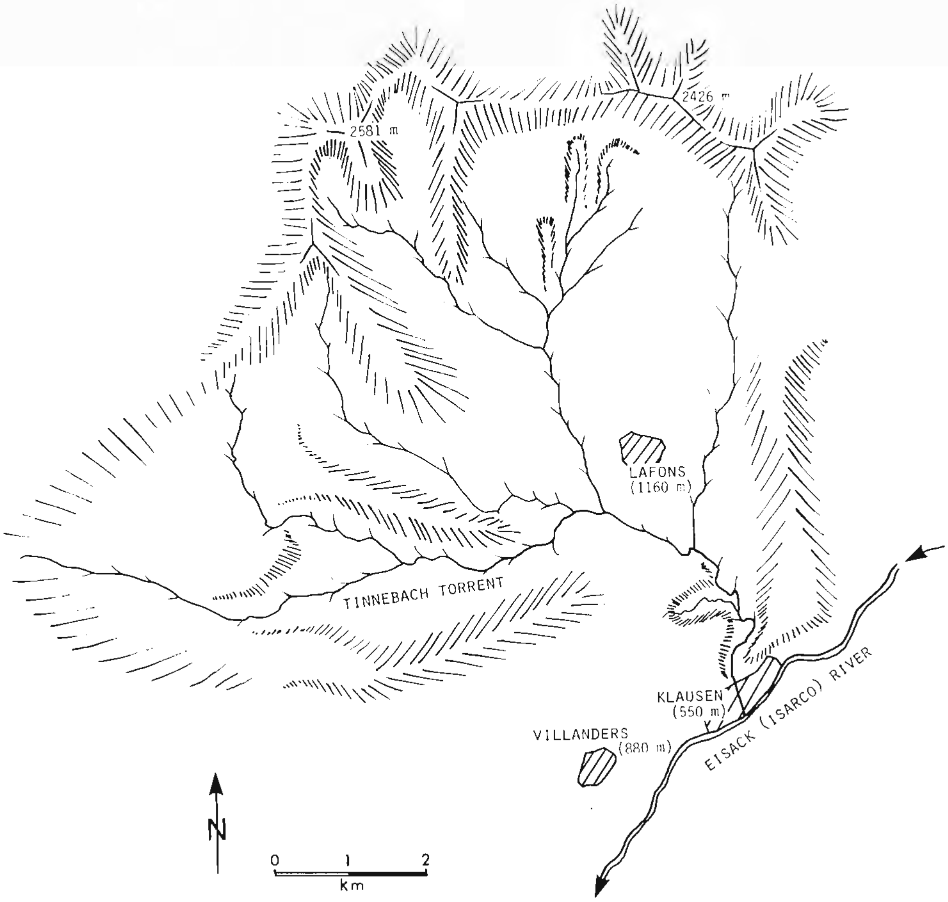


Figure 243: Sketch map of the bowl-shaped Tinnebach basin above the town of Klausen (Chiusa); most of the tributary branches of the torrent are incised into bouldery surficial deposits.

After the retreat of Pleistocene glaciers in this area much of the north side of Motto d'Arbino probably experienced slow deep-seated gravitational sagging. Material from the toe zone was carried to the cone of the Traversagna Torrent.

The first precursors of a major slope failure were rock-debris avalanches noticed as early as 1913. On 18 October 1915, a debris avalanche channeled through a bedrock ravine and buried three construction workers.

In 1925, during a routine triangulation survey in the region, monuments on the main Motto d'Arbino slope were found to be displaced (Knoblauch and Zurbuchen, 1927). The surveyor, M. Zurbuchen, issued the first warning of a possible destructive slope failure and continued to monitor the movements. During the wet summers of 1927 and 1928 a rock mass of approximately 170 $\times$ $10^6 m^3$, lodged between two fracture-controlled ravines was found to be moving at an accelerating rate of 4 to 15 mm/day. Three hamlets on the slope were evacuated. On 2 October 1928, the rate of displacement increased markedly and rockfalls broke away from the deteriorating mountainside. Soon thereafter 30 to 40 $\times$ $10^6 m^3$ of rock crashed into the gorge, blocking the Traversagna Torrent over a distance of more than 1.5 km and impounding it into a small lake; sixteen evacuated buildings were demolished (Heim, 1932, p. 156-158; Knoblauch and Reinhard, 1939, p. 82-84).

Four weeks later, on 28 October 1928, an intense rainstorm saturated the frontal part of the debris barrier, launching several debris flows which engulfed buildings in the communities of Arbedo and Molinazzo (Montandon, 1933, p. 329). In the years following the main failure parts of the sagging terrain above the head scarp of the slide mass continued to disintegrate in a series of rockfalls. Rockfall activity has been particularly intense during and after sustained periods of rain (e.g. 1936).

Between 1930 and 1950 the mouth of the Traversagna gorge was closed by a debris-retention dam 30 m high. Since then, retention space behind the dam has completely filled up. The surface of the cone is now enveloped by residential buildings; the channel of the torrent, which is wide enough to contain minor debris flows, has been provided with strong revetments.

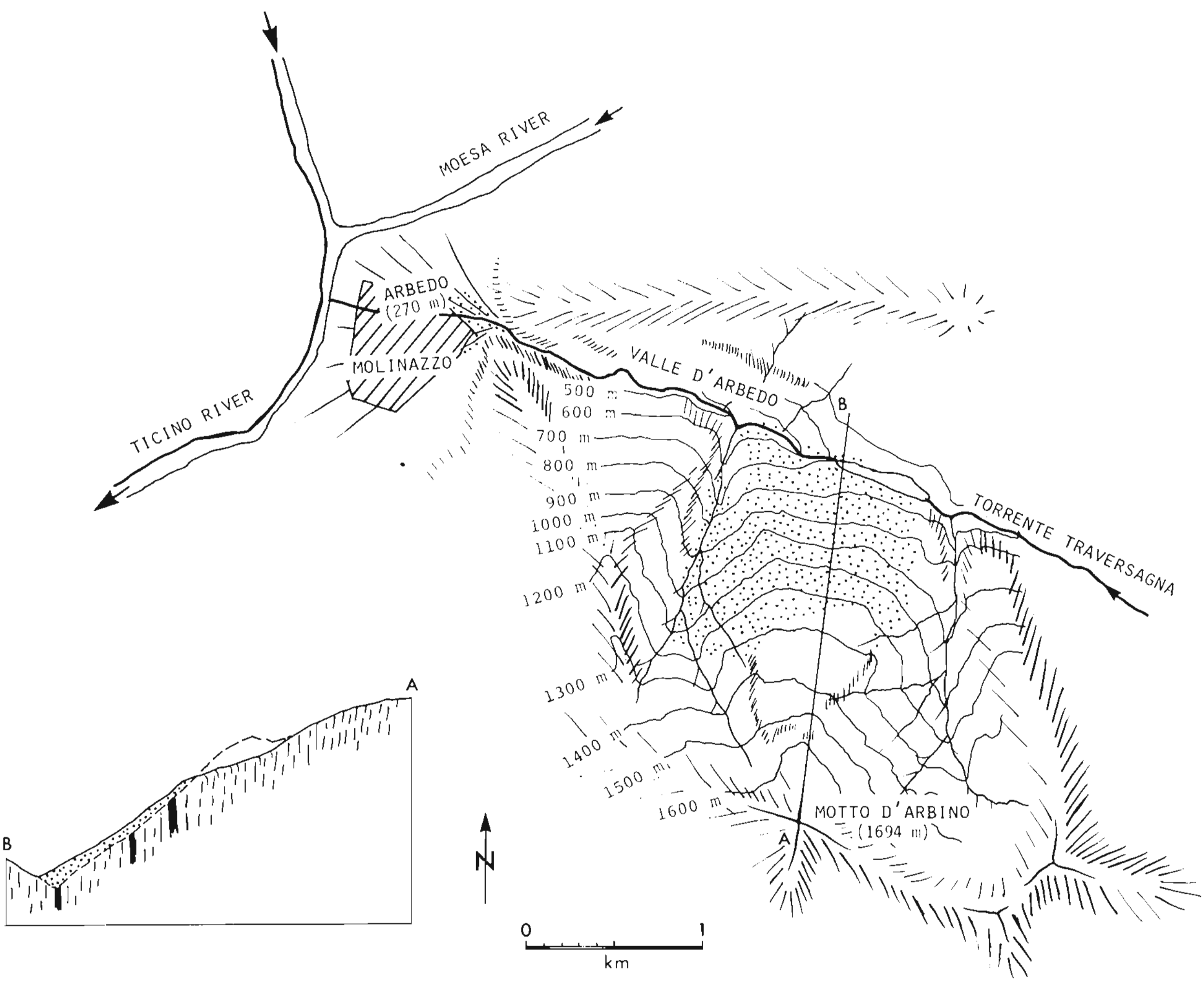


Figure 244: Sketch map and geological cross-section of the Motto d'Arbino rock-slide area and the built-over cone of Arbedo.

Linthal (A113)

Location: Glarus, Switzerland (E2)
Date(s): 6 to 7 November 1932 (also 1798, 25 August 1944)

The picturesque village of Linthal (662 m) rests at the foot of the northwestern termination of the Kilchenstock ridge (1800 m). The brow of the rising ridge behind the village has an average inclination of 38° and is underlain by gently southeast-dipping slaty sandstone of the Helvetic cover complex. Large debris cones flank the Kilchenstock (Fig. 245). The debris cone on the south side is of prehistorical age. The cone on the north of Linthal has been deposited by the Durnagel Torrent partly in historical times; major debris flows have swept down this cone in 1798 and on 25 August 1944. The main threat to Linthal, however, has been the northwestern promontory of the Kilchenstock itself.

In the past the ravines above Linthal channeled sporadic rockfalls to the base of the pointed ridge without doing much damage. During the rainy autumns of 1926 and 1927 rockfall activity from the Kilchenstock increased markedly and a crown crack appeared at an elevation of approximately 1600 m, indicating that a major segment of the ridge was in motion. In 1927 survey markers were installed at points below the head scarp of the suspected slide mass. Resurveys indicated that a rock mass of approximately $0.5 \times 10^6 m^3$ was in state of creep which accelerated during the summer of 1927 to 1.2 mm/day, during the summer and autumn of 1928 to 1.6 mm/day, and during October 1930 to 10 mm/day, at which time the village was evacuated. In July 1932 the rate of movement was 30 mm/day and in October 1932, 200 mm/day. On 6 and 7 November 1932, 20 000 m^3 of broken bedrock came loose in a series of rockfalls that did little damage (Heim, 1932, p. 190-197 and 214-217; Oberholzer, 1933, p. 566-567).

Today the village of Linthal is thriving in the shadow of the Kilchenstock (Fig. 246). A dam 650 m long and several metres high has been built behind the village to protect it

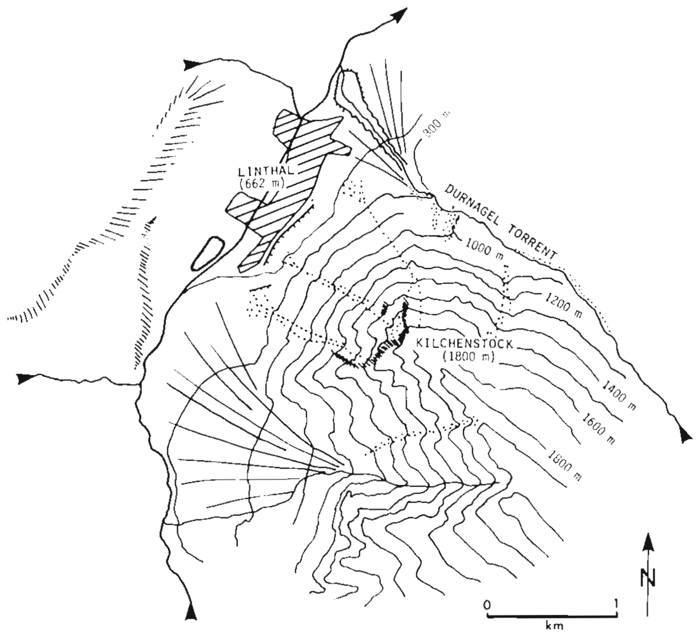


Figure 245: Sketch map of Linthal; note the protective rockfall dam below the Kilchenstock and the debris retention dam at the mouth of the Durnagel Torrent.

Figure 246: View of the Kilchenstock slope southeast of Linthal; note the two active rockfall chutes flanking the main unstable rock promontory. (GSC 204169-J)

from small rockfalls. This structure, of course, could not prevent destruction of the village if the entire slope failed. The significance of continued monitoring of creep in the unstable rock mass has been recognized since Heim's thoughtful investigation which he concluded by stating that

'. . . great rock slides get going slowly and make themselves known weeks or at least days in advance.' (Heim, 1932, p. 197).

This time interval can hopefully be used to evacuate the threatened community.

Le Châtelard (A114)

Location: Bauges, Savoie, France (B3)
Date(s): 12 to 15 March 1931 (also 11th century and 1625)

The region of Bauges is a part of the front ranges of the French Alps. Mountain ridges parallel the north-northeasterly trending folds in carbonate and argillaceous rocks of the Helvetic (Dauphinois) cover complex. Steep slopes underlain by limestone, calcareous shale, and surficial deposits are commonly unstable.

Le Châtelard (650 m) is located in the valley of the Chéran River (Fig. 247). The ridges of Mont Julioz (1600 m) to the northeast and Dent de Rossane (1900 m) to the southwest are composed of slate and intercalated ribs of limestone, in most places mantled by surficial deposits.

In the 11th century about $3 \times 10^6 m^3$ of rock failed along the northeast face of the Dent de Rossane. This slide obliterated a community in the valley and blocked the westerly flow of the Chéran River, creating a temporary lake. The survivors in the mired community founded the settlement of Les Granges (650 m) on the debris fan of the normally inconspicuous Torrent Mellesine (= bad torrent) which drains the west-facing slope of Mont Julioz. In 1625 Les Granges apparently experienced a destructive debris flow triggered by a slope failure in the uplands of the Mellesine basin.

In October 1930 large cracks appeared on the lower slope of Mont Julioz near the hamlets of Monts and Michauds (1100 m), defining an area of incipient instability 800 m long and 500 m wide. On 7 January 1931, the ground cracks captured runoff from the upper slope, and local slumping raised apprehension among the people living nearby. However, a subsequent period of cold weather and snowfall eased the fears to some degree. Then, between 28 February and 10 March 1931, about 300 mm of warm rain fell on the snow causing numerous slumps and debris avalanches along the steep mountainsides. At Le Châtelard snowmelt and runoff entered along the wide-open cracks of the failing slope. On 12 March 1931, large blocks 10 to 30 m across began to move away from a retrogressing scarp that soon involved the two hamlets. One by one buildings toppled and rotated under the strain of the expanding slide. The front of the slide transformed into a debris flow that moved down the channel of the Mellesine towards Les Granges. A total of 15

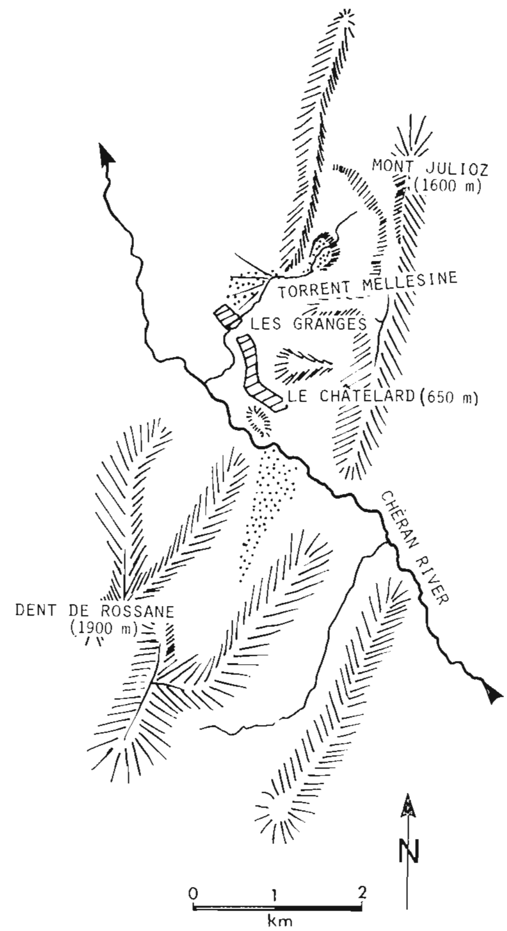


Figure 247: Index map of the bedrock ridges in the vicinity of Le Châtelard and principal area of historical slope failures.

buildings in Les Granges were demolished or seriously damaged by the debris. About $0.3 \times 10^6 m^3$ of debris spread across the surface of the cone (Gex, 1931; Montandon, 1933, p. 282 and 329). The hamlets of Mont and Michauds were abandoned.

Serrières-en-Chautagne (A115)

Location: Lac du Bourget, Rhone Valley, Savoie, France (A3)
Date(s): 17 January 1936

Serrières-en-Chautagne (260 m) consists of several hamlets strung out along the toe of the west-facing Montagne du Gros Foug (1000 m) in the foothills of the French Alps. This ridge is the core of a north-trending anticline of carbonate and calcareous shale of the Helvetic zone (Dauphinois), piercing the calcareous sandstone and shale of the Molasse Zone. The western limb of the anticline above Serrières-en-Chautagne is overlain by thick deposits of calcareous tufa. The tufa deposits weather into unstable blocks that tend to disintegrate by creep in the underlying shale.

The months of November and December 1935 were characterized by sustained rain and snowfall reaching locally a total of 900 mm water equivalent. In January 1936 all of the snow melted on the Gros Foug ridge. This resulted in a high rate of infiltration into the slope below and caused the overflow of a small reservoir. On 17 January 1936 a veneer of tufa blocks, soil and trees began to accelerate toward the hamlets of Serrières-en-Chautagne. Upon reaching the valley this mass split into several lobes, one of which destroyed 15 buildings and severely damaged several others; all inhabitants managed to escape from the doomed buildings (Martin and Messines du Sourbier, 1936).

Fidaz — Flims (A116)

Location: Rhein (Rhine) Valley, Graubünden (Grisons), Switzerland (E2)
Date(s): 10 April 1939 (also 6 August 1578, 16 May 1687, 16 March 1868

The east-trending Rhein (Rhine) Valley between Ilanz and Chur follows a major south-dipping fault zone that separates Pennine slates on the south from Helvetic carbonates on the north (Fig. 248). In this zone, characterized by a cluster of historical seismicity (Pavoni, 1977), the Helvetic basement plunges steeply to the east. The bottom of the Rhine Valley is filled almost entirely with prehistorical rockslide deposits known as Flimser Bergsturz (Heim, 1932; Trümpy, 1980, p. 225-231). These deposits originated by the failure and of some 10 000 $\times$ $10^6 m^3$ of Malm limestone formation on the Flimserstein Massif. The breakaway zone is the terrace of Flims, a dip slope inclined 15 to 25° and rimmed by a fracture-controlled limestone head scarp several hundred metres high. The vertical backwall of the terrace of Flims has been the detachment zone of several prehistorical and historical rockfalls and avalanches.

On 6 August 1578, a slab of limestone collapsed onto the talus slope above Flims; the resulting debris stream demolished several buildings and killed 10 people in the village (Niederer, 1941, p. 6). It may be significant that several strong earthquakes were felt in the region in the same year (Schorn, 1902, p. 36).

On 16 May 1687, a massive rockfall from the cliff destroyed a strip of the protective forest above Flims, and on 16 March 1868, following a winter of heavy snows, a rock-snow avalanche obliterated three cabins below the cliff (Niederer, 1941, p. 6).

On 10 April 1939, part of the nearly vertical cliff northeast of Fidaz failed. Some precursory rockfalls had been noticed prior to the main collapse, but this was not perceived as being unusual: most of the trees in the protective forest on the talus below the cliff are scarred by impact of blocks tumbling from the face above. However, on 10 April 1939, a slender rock slab, 200 m high and 0.1 $\times$ $10^6 m^3$ in volume, came loose along an easterly trending fracture surface at an

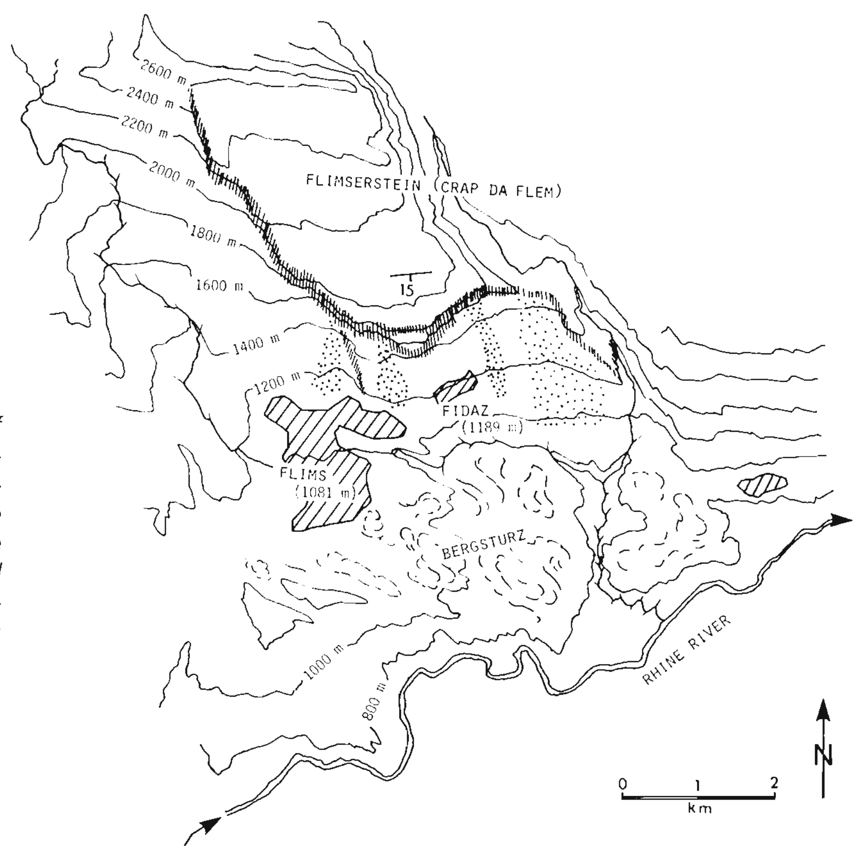


Figure 248: Sketch map of Flims and the huge Pleistocene rockslide that resulted from a failure of the Flimserstein dip slope; note historical rockfall lobes and talus derived from the carbonate cliff behind Flims and Fidaz.

Figure 249: View westward along the steep carbonate cliff of the Flimserstein above Fidaz. (GSC 204169-K)

elevation of 1600 to 1800 m. Eyewitnesses reported that at the moment of failure two cracks opened rapidly 'from the bottom up' emitting a 'thundering noise'. Then the disintegrating slab descended like a 'plowshare' over the talus and slid like a 'toboggan' through the forest. The mass then sped in a 'scouring-rolling' motion towards the children's home of Sunnehüsli (1250 m), where, preceded by an airblast, it overrode the building and killed 18 occupants. By scouring soil and colluvium, the rock avalanche attained a volume of $0.4 \times 10^6 m^3$; it carried individual blocks of limestone with volumes in excess of 500 m^3. The entire event from initial detachment of the rock slab to the sudden halt of the debris lobe took less than two minutes. An enormous dust cloud rose over Fidaz (Niederer, 1941).

On 10 November 1939, a small nondestructive rockfall broke away from the scar of the earlier slide. A barrage of artillery fire then was directed towards the cliff to bring down remaining pieces of loose rock (Niederer, 1941).

In recent years the thriving tourist centre of Flims has expanded over extensive tracts of land below the high limestone cliff (Fig. 249).

Werfen (A117)

Location: Salzach Valley, Salzburg, Austria (I1)
Date(s): 4 July 1947 (also 1683, 4 August 1974)

North of the town and castle of Werfen (550 m) the Salzach River flows in a deep gorge across the rugged front ranges of the Austroalpine carbonate complex (Fig. 250). The narrow floodplain of the river is bordered by precipitous spires, crags, walls, and funnel-shaped ravines rising above steep debris cones. The cones are, in general, uninhabited. Several bedrock spurs confine the Salzach River to a narrow channel.

The summer of 1947 was one of the hottest ever in the Alps and thunderstorms were frequent. On 4 July 1947, a series of cloudbursts deluged the southwest flank of the Raucheck Massif (2340 m). Runoff from the bare carbonate cliffs mobilized 0.1 to $0.2 \times 10^6 m^3$ of angular-blocky debris in ravines below the main mountain crest. The carbonate rubble, augmented by logs that had been deposited by snow avalanches in previous years, temporarily blocked the Kammerloch gorge. Eventually the mass burst forth onto the cone, across the railroad tracks, and into the Salzach River. A debris lobe, at least 15 m high and containing limestone blocks several metres in diameter, dammed the river. A few hours later the barrier broke and a flood wave demolished bridges and roadworks downstream from Werfen. The catastrophe claimed two lives (Lauscher, 1973, p. 144). Historical records suggest that a similar debris flow disaster occurred near Werfen in 1683. After 1947 stone-masonry check dams were built across the main tributary ravine of the uplands to prevent future mobilization of the carbonate rubble below the Raucheck (Fig. 251). The low floodplain above the bedrock spur of Werfen, avoided in the past and submerged during the blockage of the Salzach River in 1947, now hosts an outdoor sports centre and a few homes.

In the spring of 1974 a rockfall occurred on the steep northeast ridge of the Rifflkopf summit (2254 m) on the east of the Salzach River. The calcareous rockfall rubble blocked the upper reaches of a bedrock ravine that drops steeply down

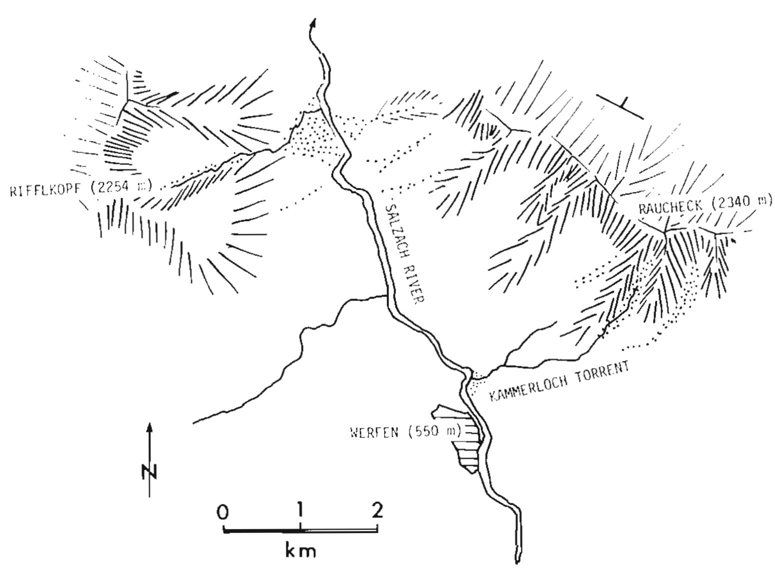


Figure 250: Sketch map of the Salzach River gorge near Werfen and the steep carbonate cliffs of the Raucheck Massif.

Figure 251: a) View of the intensely fractured carbonate cliffs near the summit of the Raucheck Massif; note the sparsely vegetated track of the 1947 debris flow in the ravine below the cliff (GSC 204169-L). b) Stone-masonry check dam across the upper Kammerloch Torrent built to prevent mobilization of carbonate rubble derived from the Raucheck Massif (GSC 204169-M).

to a cone astride the river embankment. On 4 August 1974, a heavy thunderstorm deluged the Rifflkopf Massif and runoff scoured deeply in the rockfall and talus blankets of the ravine below. About 10 000 m^3 of the rockfall rubble were mobilized into a coherent flow which swept across the highway at great speed, picking up two cars and hurling them into the river. Both drivers perished (Pippan, 1981). During the late 1970s the new highway through the Salzach gorge was placed into tunnels and galleries which thus permit safe travel along this stretch of valley exposed to rockfall, debris flow, and snow avalanche activity.

Glacier du Tour (A118)

Location: Vallee de Chamonix, Haute Savoie, France (B3)
Date(s): 14 August 1949

The Glacier du Tour is a northwest-facing cirque glacier located in a bowl of granitic gneisses below the northeastern ridges of the Mont Blanc Massif (Fig. 252). The firn basin above 2700 m feeds an intensely crevassed tongue of ice that overhangs a smooth ledge of bedrock, inclined at 25 to 35°, at an elevation of approximately 2500 m.

On 14 August 1949, without warning, approximately 2 to 3 $\times$ $10^6 m^3$ of ice broke away from the frontal lobe of the glacier at 2200 m. The slab disintegrated into a chaotic flow of ice fragments and dust, then avalanched explosively to the foot of the bedrock slope, and spread across the morainal banks in the valley, killing six hikers. Photographs of the ice avalanche in progress were taken by a tourist from across the valley (Guichonnet, 1950).

The hazard of potential ice avalanches in the runout zone below the glacier has been recognized and developments are not expanding towards the morainal ridges below the Tour Glacier.

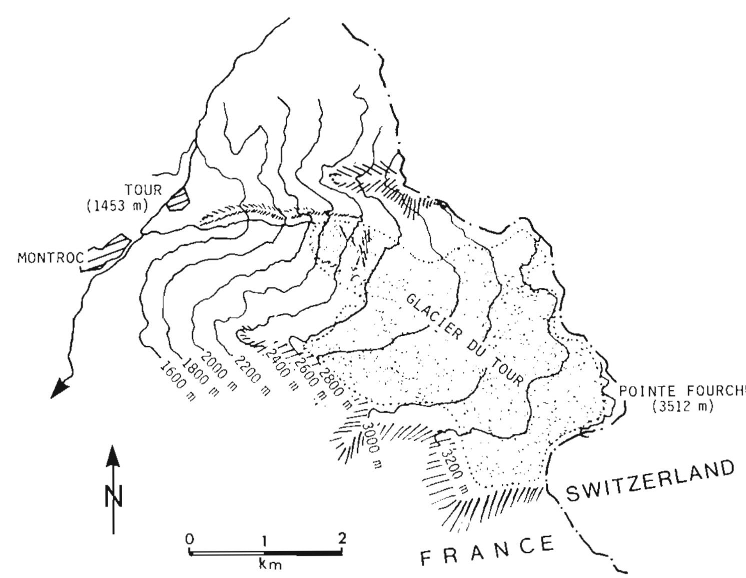


Figure 252: Sketch map of Glacier du Tour, eastern Mont Blanc Massif.

Tavernerio (A119)

Location: Como, Lombardia, Italy (E3)
Date(s): 8 November 1951

Tavernerio (420 m) straddles the lower Cosia Torrent which originates by the confluence of several tributaries flowing from bedrock basins underlain by south-dipping limestone and shale formations of the South Alpine cover complex. Dip slopes along the crest of the tributary basins below Monte Bolettone (1317 m) tend to disintegrate in rockfalls and debris slides.

Between 6 and 11 November 1951, heavy rainfalls deluged the south-central front ranges of the Italian Alps creating devastating floods in the Po Valley. In the mountains debris avalanches caused extensive damage to roads and settlements. On 8 November 1951, a slide mass of shale and carbonate blocks from the Monte Bolettone dip slope blocked the Cosia Torrent and then changed into a debris flow. The flow entered the upper part of Tavernerio, crushed ten buildings and killed 16 people (Mussio, 1974).

Today little remains to be seen of the 1951 debris flow track. Most of the torrent embankments at Tavernerio are completely built over.

Becca de Luseney (A120)

Location: Valpelline, Valle d'Aosta, Italy (C3)
Date(s): 8 June 1952

The Valpelline, drained by the Buthier Torrent extends from the high glaciated border ranges between Italy and Switzerland to the Valle d'Aosta. Bedrock ridges flanking the valley are composed of intensely fractured southwest-trending metamorphics of the South Alpine-Austroalpine basement complexes. The Comba d'Arbiere, a glacial tributary torrent, originates in a rubble strewn cirque below the ragged summit pyramid of the Becca de Luseney (3500 m) and joins the Buthier Torrent near the Hamlet of Chamen (Fig. 253).

On 8 June 1952, following two years of exceptional snowpacks and meltwater infiltration, a slide mass consisting of $0.1 \times 10^6 m^3$ of bedrock, superincumbent moraine, ice and snow came loose along the west-facing slope of Becca de Luseney at an elevation of 2900 m. Picking up blocks of ice, morainal debris, and water, it changed into a rapid debris flow 250 m wide. Cascading down the straight gorge of the Comba d'Arbiere Torrent, the mass struck the bottom of the Valpelline and then climbed approximately 100 m up the opposite valley wall. The debris obliterated huts in the hamlet of Chamen (1700 m) and killed 4 people. A small lake formed upstream from the debris lobe and four days later launched a debris flood carrying blocks of ice and rock down the Valpelline (Stragiotti and Peretti, 1953; Abele, 1974, p. 141).

The upper Valpelline, although accessible by road, is still relatively remote and only limited development related to tourism has invaded this rugged region.

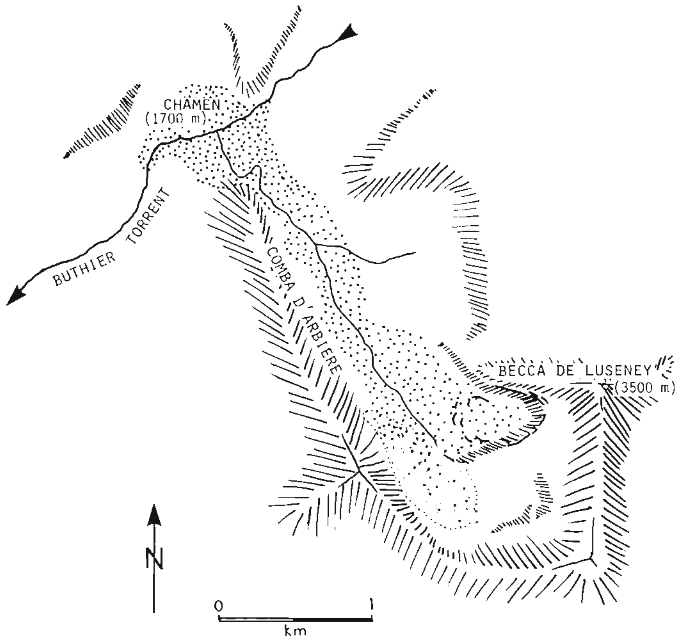


Figure 253: Sketch map of the rock-ice-debris stream of 1952 which originated on the western flank of Becca de Luseney.

Zambana (A121)

Location: Adige Valley, Trento, Trentino, Italy (G2)
Date(s): 25 November 1955 (also 1811, 17 April 1956, autumn 1959)

The community of Zambana (215 m) used to be located at the apex of a gently sloping debris fan that rises only about 25 m above the wide and intensely cultivated floodplains of the Noce and Adige rivers. Directly behind the village rises the Paganella wall, a bedrock face composed mainly of thick-bedded to massive dolomite formations of the South Alpine cover complex dipping 10 to 20° west-northwest (Fig. 254). Controlled by near-vertical faults and fractures, this wall is nearly 2000 m high. Sections of the Paganella wall tend to topple or collapse along vertical fractures. In 1811 a slice of the wall peeled off along a composite fracture just north of the community of Mezzolombardo and spread across the valley floor as an 'enormous sea of rubble' (Schaubach, 1867, v. 4, p. 385). This debris lobe has since been completely covered by alluvium of the Noce River and now supports extensive orchards.

Near Zambana the Paganella wall is breached by the fault-controlled Val Manara ravine. Although dry during most of the year the ravine can carry substantial runoff after autumn rains and springtime snow. Since its earliest days the settlement faced sporadic bursts of debris and regular floods of the Adige River. Mesolithic cultural traces near Zambana and Roman coins unearthed from the cone demonstrate that in spite of these threats settlers have been attracted to this site for more than 6000 years (Gorfer, 1977, p. 382-384).

The destructive events of 1955-1956 began on 7 August 1955, when a rockfall from the Paganella wall blocked the Val Manara gorge; on 7 September 1955 during a period of rain the mobilized debris invaded the upper section of Zambana. Other slabs fell from the cliff soon thereafter and apparently prepared the slow detachment of a downward tapering wedge of limestone with a volume of 0.2 to 0.3 $\times$ $10^6 m^3$ along two vertical fractures near the 1000 m contour. Simple pegs were installed across the main fracture near its crown to monitor the outward displacement of the rock slab. Evacuation of parts of Zambana was also ordered soon thereafter. On the night of 24 November accelerated subsidence and rotation of the rock mass indicated impending failure. In the early morning hours of 25 November 1955, the slender rock column collapsed internally and toppled forward, filling the gorge of Val Manara. The gruesome spectacle was illuminated by floodlights set up by the army beforehand. It soon became clear that the annual spring rains might mobilize the mass of debris; therefore construction of a protective stone dam above Zambana was initiated immediately. Indeed, on 17 April 1956 the first heavy spring rains set in motion most of the debris which unfortunately overtopped the protective works. The church and other buildings were covered by several lobes of carbonate rubble (Andreatta, 1956). Autumn rains in 1959 set off another series of flows flattening a second protective dyke and invading another row of houses in Zambana (Fig. 255).

After these disasters a new community, Zambana Nuova, was established 2 km east of Zambana Vecchia on the floodplain of the trained Adige River. However, the old community, Zambana Vecchia, was not abandoned; the church has been excavated completely from the rubble and some of the seriously damaged buildings have been repaired. The most threatening rock promontories of the Paganella wall are now inspected regularly because their collapse could be locally disastrous to such communities as Mezzolombardo (Largaiolli and Vaia, 1975).

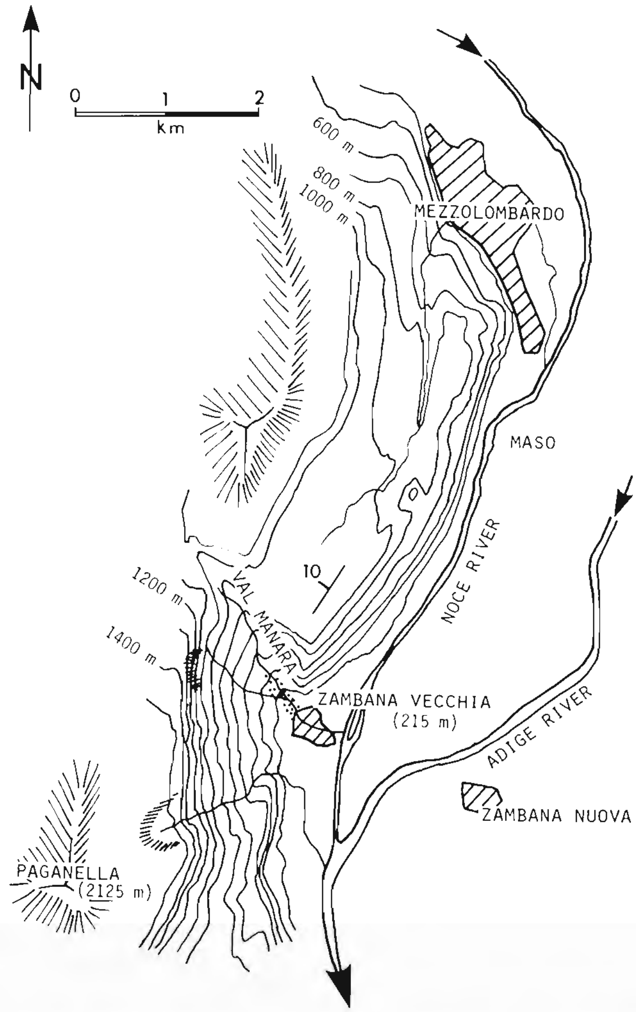


Figure 254: Sketch map of the Paganella carbonate wall at the confluence of the Adige and Noce rivers.

Figure 255: Debris flow deposits that engulfed Zambana in 1955 as seen from the mouth of Val Manara (from Andreatta, 1956).

Millstatt (A122)

Location: Lake Millstatt (Millstätter See), Kärnten (Carinthia), Austria (I2)
Date(s): 31 July 1958

Lake Millstatt (588 m) is one of several lakes in the densely populated tourist region of central Kärnten (Carinthia). Mountain ranges in the region rise to approximately 2000 m (Fig. 256) and are underlain by schists and gneisses of the Austroalpine basement complex. Many torrent basins of the region are flanked by sagging bedrock slopes and thick deposits of Pleistocene colluvium (Clar and Weiss, 1965). Debris cones along lake shores and in river valleys host sizable towns and villages. Although blessed with a mild climate, the generally hot summers tend to be interrupted by violent thunderstorms accompanied by intense downpours. Legends relating to ancient debris flow disasters speak of furious dragons ('Lindwurm') and enraged giants.

On 31 July 1958, a cloudburst dumped approximately 100 mm of rain onto the Millstätter Alpe Massif (2086 m). During this downpour an intensity of 36 mm/h was registered. Trees, knocked over by strong winds, mixed with debris eroded from torrent embankments, generating debris flows which erupted over the Millstatt and Pesenthein cones, among others. An estimated total volume of $0.3 \times 10^6 m^3$ of rubble reached the lower parts of the torrents in the area. The outer part of the overloaded cone at Millstatt slumped into the lake. About 80 buildings were severely damaged and six people were killed, four of them at a campground on the Pesenthein cone. The destruction could have been even more severe, but was held in bounds by check dams and other control works already in place along the torrents (Richter, 1959).

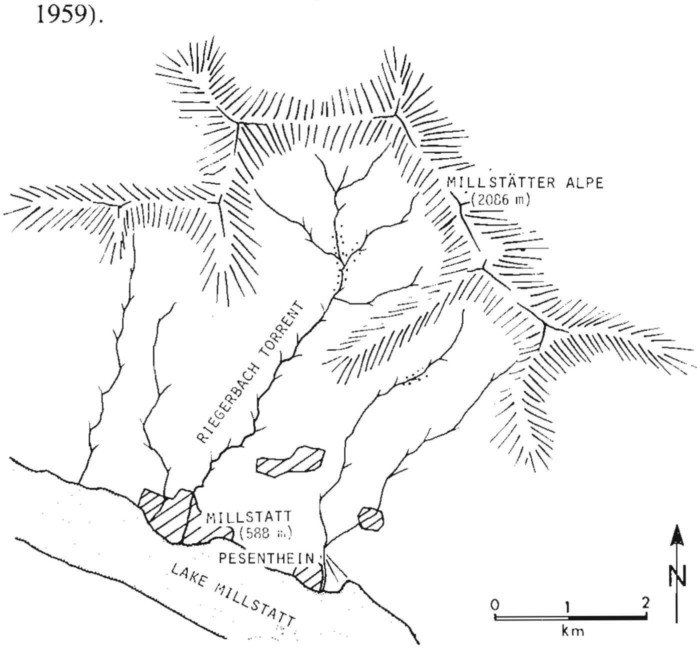


Figure 256: Sketch map of the torrent basins on the southwest slope of the Millstätter Alpe; note completely built-over fan deltas.

Figure 257: View of the Pesenthein cone and Lake Millstatt. Left cone sector is entirely occupied by campground facilities; right cone sector hosts several recently built hotels. (GSC 204169-N)

In recent years most of the cones in the area have been completely built over. Additional control works have been erected on some torrents, but, since the recurrence interval of thunderstorms like the one in 1958 is about 300 to 500 years (Richter, 1959) the debris flow risk on many small cones has been accepted (Fig. 257).

Mur Valley (A123)

Location: Steiermark (Styria), Austria (K1)
Date(s): 12 to 14 August 1958 (also 22 July 1740, 8 June 1827, 6 to 7 September 1833)

The upper Mur and Mürz valleys follow a seismically active northeast-trending Neogene fault zone (Fig. 258). The two major towns along this industrial corridor are Bruck and Leoben (500 m). Most of the mountain ranges along the Mur-Mürz lineament rise to elevations about 1500 m and are underlain by schists, phyllites and carbonates of the Austroalpine basement complex. The area is outside the Pleistocene glaciation limit of the Alps and exhibits deep weathering. Violent hailstorms and thunderstorms are frequent in this zone bordering the Pannonian Basin and produce flash floods in the lower foothills, debris flows in the mountainous basins, and debris slides along deeply weathered foliation dip slopes in metamorphic bedrock terrain. Localities with names containing the root 'Göss', 'Güss' and 'Giess' (= debris flood) reflect this physiographic situation.

The winter of 1740 in eastern Austria was long and severe. After a very short spring there were powerful thunderstorms and downpours in July. On 22 July 1740, a cloudburst centred in the vicinity of Leoben triggered massive debris movements in the Lainsachgraben basin. Debris flows demolished several homesteads there and claimed 32 casualties (Stini, 1938, p. 18).

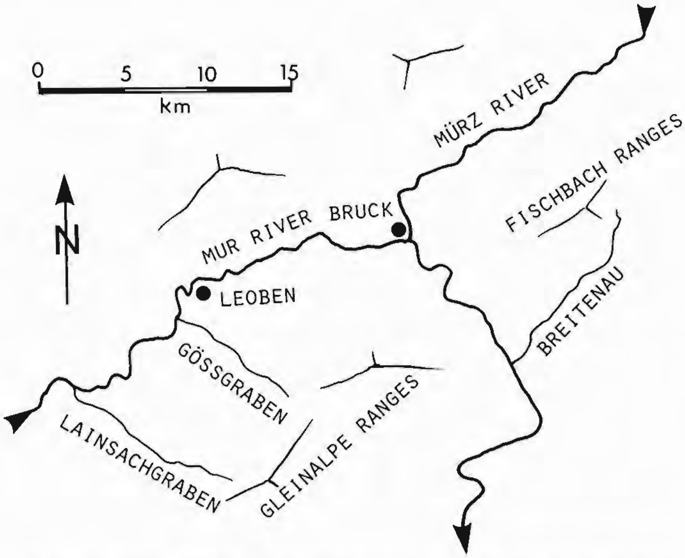


Figure 258: Index map of the Mur and Mürz valleys and adjacent mountain ranges.

On 8 June 1827, a regional rainstorm in the eastern Alps set off numerous debris avalanches and flows in the Mur-Mürz basin. Transportation routes were blocked for many weeks, dozens of buildings were demolished and many people lost their lives (Stini, 1938, p. 23).

On 6 and 7 September 1833, a regional rainstorm accompanied by violent local squalls deluged the slopes south of Leoben, triggering debris avalanches and floods in the area. A large debris slide blocked the Gössgraben Torrent, and a massive flow overwhelmed buildings along the embankments, claiming the lives of several people (Stini, 1938, p. 24).

The summer of 1958 was very hot throughout the Alps. These conditions favoured local cloudbursts of considerable intensity particularly along the eastern fringes of the mountains. Between 12 and 14 August 1958, a pulsating shower cell was centred in the Fischbach Ranges (Breitenau), southeast of Bruck. A total of 400 to 500 mm of precipitation fell within 8 hours. One squall lasting for 30 minutes dumped 100 mm of rain onto the uplands. Gullies and slides in weathered bedrock filled the torrents with debris and dislodged trees. Debris-log jams rose up to 30 m above the torrent channels and when the barriers failed violent bursts of debris engulfed roads, buildings and claimed the lives of 12 persons (Clar, 1959; Richter, 1959). It has been estimated that within the 100 km^2 of the storm cell about $10 \times 10^6 m^3$ of debris were in motion.

Since 1958, many communities in the tributary valleys of the Mur-Mürz corridor have doubled in size. Most of the debris cones in the region have been built over (Fig. 259). The major torrents have been provided with revetments or check dams, but the smaller ravines are largely in their natural state.

San Giovanni de Crèvola (A124)

Location: Crevola d'Ossola, Piemonte, Italy (D2)
Date(s): 20 August 1958

The village of San Giovanni de Crèvola (373 m) used to be located on a steep debris cone in Val Divedro, the narrow gorge south of the Simplon Pass. The catchment basin of the

Figure 259: a) Typical V-shaped, densely forested torrent basin in the Breitenau Valley. Although normally characterized by insignificant runoff this type of basin produced large debris flows during the catastrophe of 1958. Note recently constructed buildings on the apex of the small cone (GSC 204169-O). b) Example of a check dam-revetment combination in the Mur Valley (GSC 204169-P).

small torrent joining the Diveria River at this point is rimmed by bedrock ridges with elevations of approximately 2400 m. Gneissic bedrock of the Pennine core complex along the northern crest of the basin is unstable.

On 20 August 1958, an intense rainstorm set off a massive rock-debris avalanche from the unstable dip slope above San Giovanni de Crèvola. The avalanche crushed 12 houses, killed 13 people, and stemmed the flow of the Diveria River for a short time. When the debris dam yielded to the pressure of the impounded river, a debris flood surged through the bedrock gorge onto the gentle fan at Crevola d'Ossola, wreaking havoc there.

After the catastrophe San Giovanni was abandoned and, except for the overgrown debris cone, little remains to be seen. The railroad passes through a tunnel beneath the cone. The highway has been rerouted to the far side of the valley, placed safely in rockfall and snow avalanche galleries.

Riviera (A125)

Location: Alpes Maritimes, France (C5)
Date(s): 19 October 1959 (also 24 April 1952, 2 December 1959, 25 August 1965)

The Riviera is the densely populated coastal fringe of the Alps bordering the Mediterranean Sea. Bedrock slopes underlain by carbonate and flysch of the Helvetic (Dauphinois) cover complex are dissected by deep gorges and precipitous ravines that lead from attractive coastal cities to rugged upland basins. Intense rainstorms, generally caused by the mixing of warm Mediterranean air masses with cold mountain air trigger rockfalls, debris slides, and torrent activity of considerable regional impact.

On 24 April 1952, a swollen river undercut a built-over bedrock embankment at Mentone causing the collapse of 25 buildings and claiming the lives of 14 people.

In the autumn of 1959 about 90 mm of rain were registered within a time span of 24 hours. Numerous torrents severely eroded their embankments. Wind-driven waves undercut coastal cliffs which collapsed. Numerous buildings and four lives were lost. Continued rainfall led to the catastrophic failure of the Malpasset arch dam on 2 December 1959. Slight slippage of one of the abutments prepared the collapse of the structure. An enormous wave of water and scoured debris surged down the valley and claimed the lives of 424 people, most of them inhabitants of Frejus (Lotze, 1960, p. 183).

During the rainy summer of 1965 numerous debris slides and rockfalls occurred along the Riviera. On 25 August 1965, a massive rockfall near Noli killed 22 people.

Density of population along the shore of the Riviera brings with it an almost inevitable risk of rockfall and debris flow damage. However, since the catastrophe of Malpasset the variable runoff and debris movements in many of the ephemeral torrent channels are fully appreciated. The dam of Malpasset has not been rebuilt.

Illgraben (A126)

Location: Rhone Valley, Wallis (Valais), Switzerland (C2)
Date(s): 26 March 1961

The Illgraben is a rugged uninhabited tributary valley south of the Rhone River, across from the town of Leuk (Fig. 260). The huge debris cone at the mouth of the Illgraben Torrent has forced the Rhone River against the north side of the valley. It also forms a major climatic-cultural border between the eastern and western regions of the Wallis (Valais). The communities of Susten (624 m) and Pletschen are located on the eastern sector of the cone (Fig. 261). The Illgraben basin is underlain by northwest-dipping intensely fractured quartzite and evaporitic carbonates of the Pennine cover complex. Failures along the bare and deeply ravined scarp slope of the Gorwetschgrat (1800 m) have supplied abundant debris to the Illgraben Torrent and its cone. Minor historical debris flows (e.g. 1920, 1928, and 1934) have been discharged onto the upper segment of the cone which is mantled by an extensive protective forest.

On 26 March 1961, a failure of 3.5 × $10^6 m^3$ of evaporites and carbonate rock filled the upper Illgraben with an elongate mass of debris. Runoff from the crest of the basin

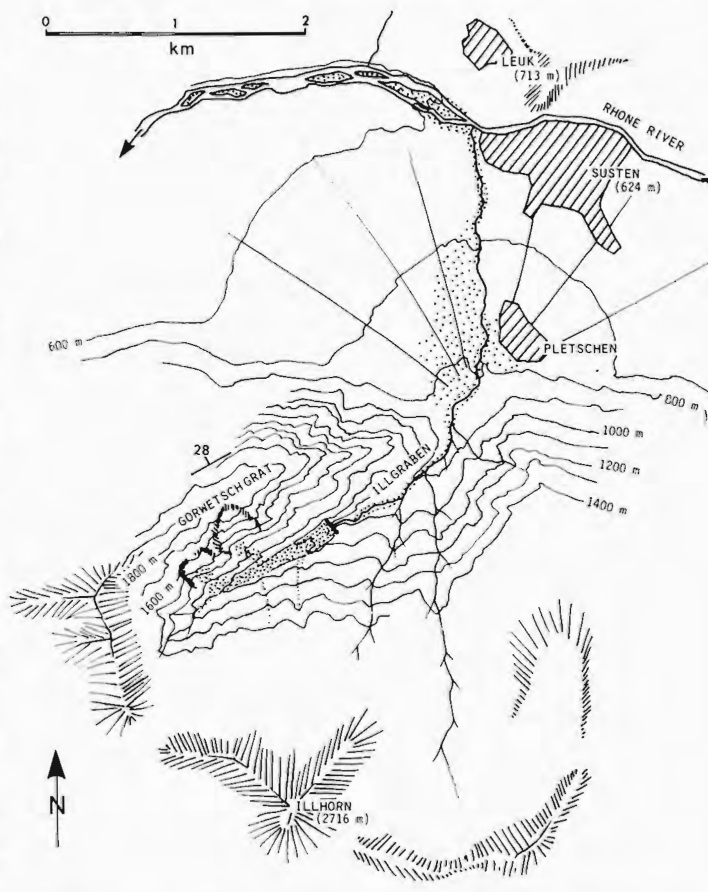


Figure 260: Sketch map of the Illgraben and the large cone at its mouth.

soon saturated the rubbly lobe and, by 6 June 1961, a small lake had formed behind the slide. It emptied rapidly and a major debris flow reached the Rhone River. There, the debris threatened to destabilize the trained river channel.

To prevent additional debris flows from the unstable slide mass, a 50 m high steel-and-concrete dam was built across the bedrock gorge immediately below the slide deposits (Lichtenhahn, 1971a). Soon debris completely filled the space between the dam and the slide; thus a volume of approximately $2 \times 10^6 m^3$ of slide material was effectively neutralized (Fig. 262). The torrent continued to undercut the part of the slide projecting above the level of the dam's crown, and check dams had to be erected along the channel below the dam to keep the torrent from shifting towards the built-over cone sector of Pletschen and Susten. These works will require careful maintenance in the future (Fig. 262).

Figure 261: View of the Illgraben cone; note protective forest along the torrent channel; community of Susten at lower right. (GSC 204169-Q)

Figure 262: a) Steel-concrete dam, approximately 50 m high, at the toe of the Illgraben rockslide of 1962 (GSC 204169-R). b) Two check dams across the channel of the Illgraben Torrent below the main retention dam; note damage caused by small debris flows (GSC 204169-S).

Vaiont (A127)

Location: Piave Valley, Friuli-Venzia Giulia and Veneto, Italy (H2)
Date(s): 9 October 1963 (also 4 November 1960)

The Vaiont Torrent is a major eastern tributary of the Piave River which flows south across the front ranges of the southern Alps (Fig. 263). Before it was impounded by the ill-fated Vaiont dam the torrent dropped from an elevated bedrock basin (700 to 1000 m) through a narrow gorge to the floodplain of the Piave River. A slightly elevated bench on the west side of the Piave River opposite the mouth of the Vaiont gorge hosts the town of Longarone (474 m). Both the Piave and Vaiont valleys are rimmed by bedrock ridges soaring to elevations of approximately 2000 m. Regional bedrock structure is dominated by east-trending folds and south-directed thrust faults involving well bedded carbonate formations of the South Alpine cover complex. The Piave Valley follows transverse high-angle faults (e.g. Linea del Col delle Tosatte) which may have been the locus of minor displacements or at least intense shaking during historical earthquakes. Schaubach (1867, v.4, p. 450), a traveller in the early 19th century, reported that

'. . . below Termine (i.e. a few kilometres north of Longarone) the road crosses a rock fissure formed by an earthquake in the 16th century..'

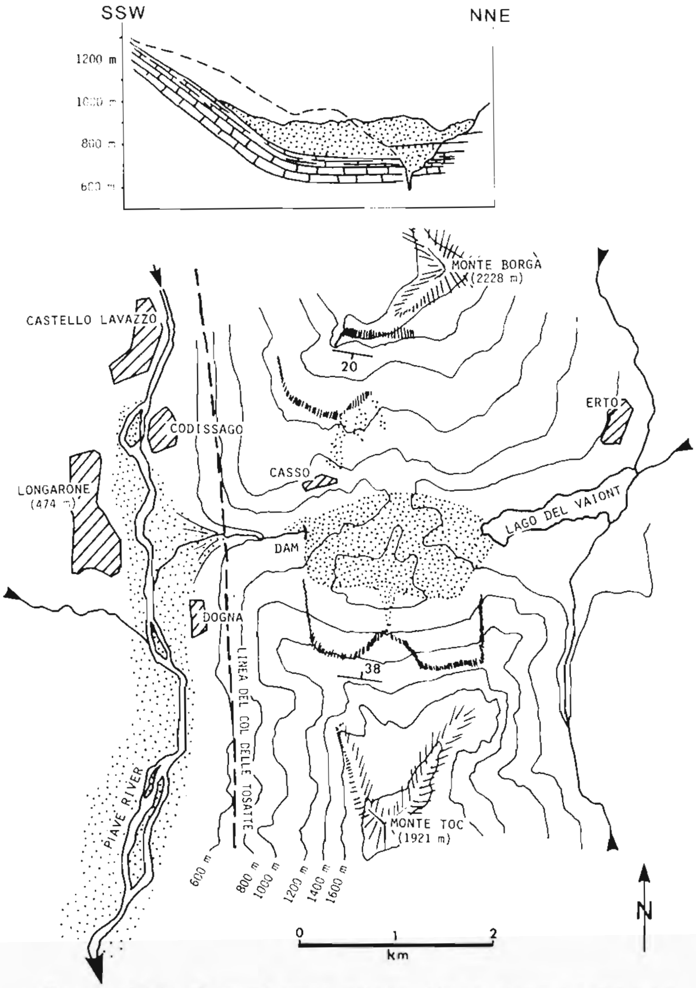


Figure 263: Sketch map of the Piave Valley near Longarone; inset shows cross-section (no vertical exaggeration) of the rockslide that filled the ill-fated Vaiont reservoir (after Broili, 1967).

The Vaiont Valley follows a gently eastward plunging syncline of Jura-Cretaceous limestone and marl. The syncline is slightly asymmetric to the south and overthrust from the north by the Monte-Borgà thrust plate. The northern wall of the valley is also partly covered by deposits of prehistorical rock avalanches. The southern valley wall is a dip slope rising to the summit ridge of Monte Toc (1921 m); bedding dips between 30 and 40° to the north.

In September 1960 a bold hydroelectric concrete dam, 261 m high, was completed across the upper part of the Vaiont gorge. Soon thereafter the valley floor above the slender concrete shell disappeared beneath the impounded waters of the reservoir. On 4 November 1960, a mass of deeply disintegrated bedrock and surficial deposits, $0.7 \times 10^6 m^3$ in volume, slumped away along the submerged toe of the southern embankment below Monte Toc, leaving behind a horseshoe-shaped scar. Subsequent changes in the level of the reservoir continued to cause deep-seated movement of the Monte Toc slope. These movements were now monitored by regular surveys. Late in 1960 en-bloc dislocation of the lower segment of the slope became evident, when a composite fracture opened high on the mountainside. The zone of cracks could be followed for about 2 km along strike. Now more instruments (e.g. piezometers) were installed on the slope and seismic sounding was carried out at the site. The results of the seismic survey apparently indicated a peculiar downward decrease in the seismic velocity indicating that the rock mass may have been dilating at depth; the seismograms disappeared after the catastrophe.

In August and September 1963 precipitation in the Piave Valley was three times higher than normal. Infiltration of rainwater into the dilating dip slope below Monte Toc probably contributed substantially to its eventual failure in October. On the day before the disastrous failure of the rock mass and its rapid descent into the reservoir creep rates of 40 cm/day were registered.

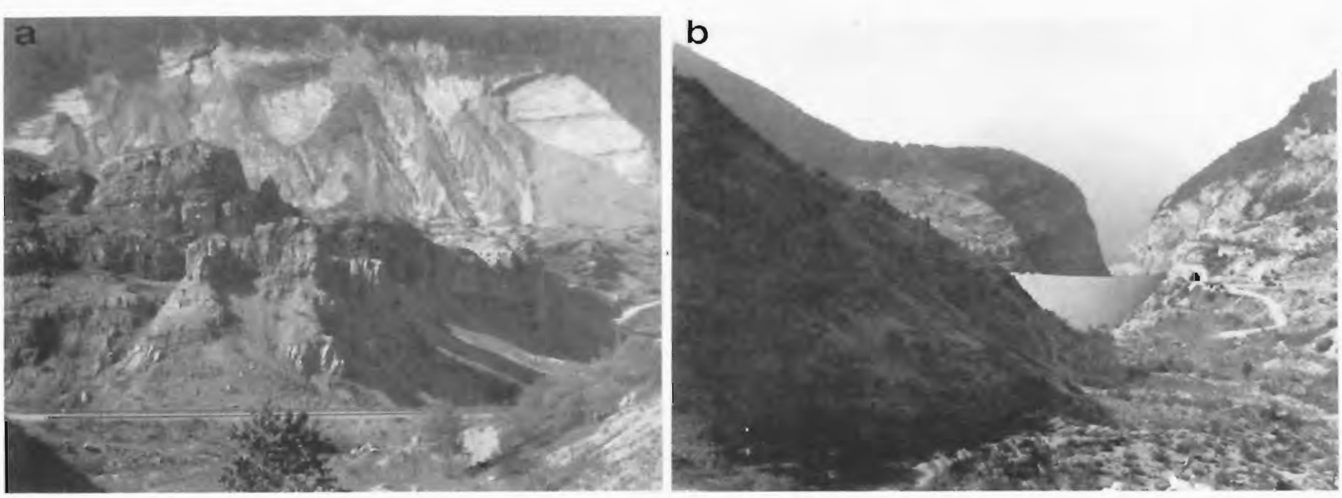


Figure 264: a) Detachment surface and toe of the Vaiont slide; note coherent strata of the slide mass (GSC 204169-T). b) View downstream towards the concrete dam which closes the Vaiont gorge (photo taken in 1980) (GSC 204169-U).

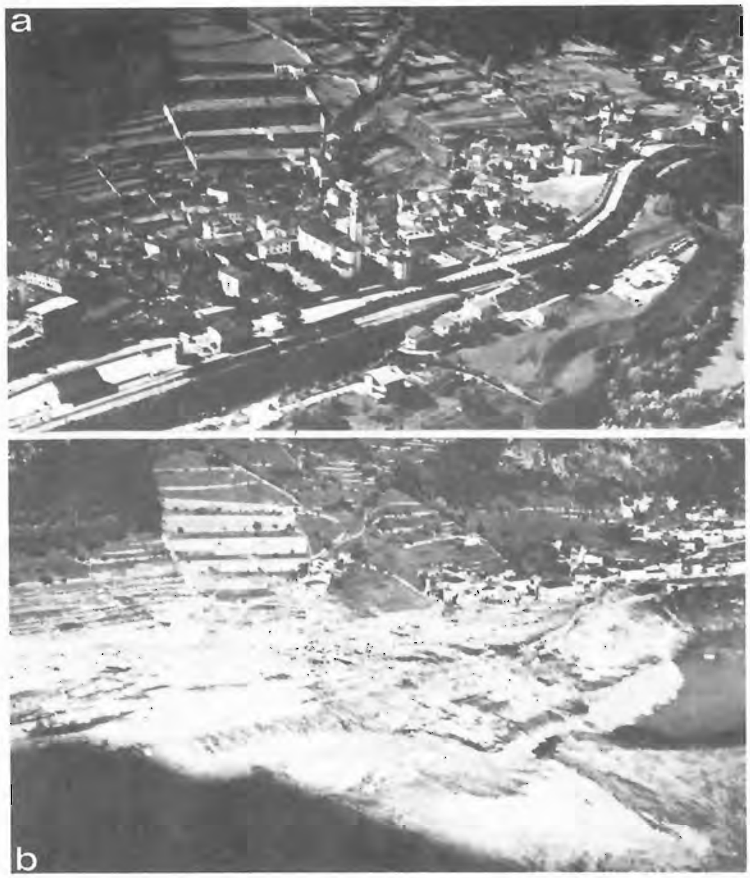


Figure 265: Before and after views of Longarone, a sizeable town destroyed by the debris flood from the Vaiont reservoir.

On the night of 9-10 October 1963, the unstable seat-shaped slab of the lower Monte Toc slope failed. Most of the surface of rupture followed a medium to thin-bedded limestone-marl unit dipping about 30° to the north (Fig. 263). Small undulations of the bedding planes and fractures across bedding resulted in a composite detachment surface with a somewhat steeper inclination. The thickness of the whole slab amounted to 250 m and its volume was calculated as approximately 250 × $10^6 m^3$. As the slab moved away from the head scarp it remained internally coherent (Fig. 264a) and pre-slide features on the surface, e.g. stone walls and trees, were rotated 10 to 30° against the direction of movement. Parts of the toe of the slide were crushed and overridden by the main mass as it plunged into the narrow reservoir. A wall of water surged almost 250 m up the north side of the valley, then veered westward, and overtopped the concrete dam which withstood this onslaught. About 30 × $10^6 m^3$ of water, cutting a sharp trim line above the dam (Fig. 264b), picked up momentum by dropping into the narrow defile below, then scoured blocky debris from the bottom of the torrent channel, and fanned across the Piave River. The wall of water erased most of Longarone and nearby smaller communities, killing at least 1900 people (Fig. 265).

In the wake of the catastrophe several scientific investigations were launched in rapid succession. The results were published in a series of papers by Weiss (1964), Broili (1967), Müller (1968), Cadisch (1970), and Jaeger (1979, p. 402-423). Several engineers and a geologist involved in the construction of the Vaiont power project were charged with causing a landslide and flood, manslaughter and injury. The long trial included eyewitness accounts of 2500 people and presentations by 60 lawyers; three engineers were found guilty of manslaughter, because they had been somewhat lax in their responsibilities during the time leading up to the disaster.

The site of the Vaiont dam has been left as it was after the catastrophe. The ramifications of this slope failure were broadened by the international interest in the trial. Vaiont has remained a monument to human skill (the dam held) and misjudgment (signs of impending failure were misread). It has led to greater attention towards slope stability problems along hydroelectric reservoirs in steep-sided valleys. Longarone has been rebuilt and is again a thriving town spreading over the bench along the west side of the Piave River.

Pettneu (A128)

Location: Stanzer Valley, Arlberg, Tirol, Austria (F1)
Date(s): 29 to 30 June 1965

The Stanzer Valley, drained by the east-flowing Rosanna Torrent, separates metamorphic bedrock of the Austroalpine basement complex on the south from intensely deformed Austroalpine carbonate rocks on the north. Most of the agricultural-tourist communities in the valley, including Pettneu (1200 m), spread over south-facing debris cones below steep ravines flanked by bare dolomite cliffs (Fig. 266). The ravines are the tracks of regular snow avalanches, sporadic rockfalls and debris flows.

The ravine north of Pettneu, called Gridlontobel, rises steeply towards a ridge of fractured dolomite. A prehistorical slide mass of broken dolomite and shale is lodged in a funnel-shaped re-entrant at 2100 to 2300 m elevation below the serrated summit ridge. The debris cone of Pettneu has received sporadic flows from the toe zone of this unstable rock mass since 1789.

The spring of 1965 was cold and melting of the thick snowpack in the uplands was long overdue when in mid-June temperatures suddenly rose. This delayed arrival of summer was accompanied by thunderstorms and heavy rain. Due to infiltration of meltwater and rain the frontal section of the slide mass in the Gridlontobel basin became saturated and parts began to slump into the main ravine which was still filled with avalanche snow.

On 29 June 1965, masses of bouldery debris continued to break away along a concave scarp; the torrent issuing from the Gridlontobel onto the cone turned increasingly turbid. These warning signs prompted an evacuation of buildings along the channel of the normally inconspicuous creek bed. On 30 June 1965, several pulses of slushy debris overwhelmed the eastern sector of the community destroying half

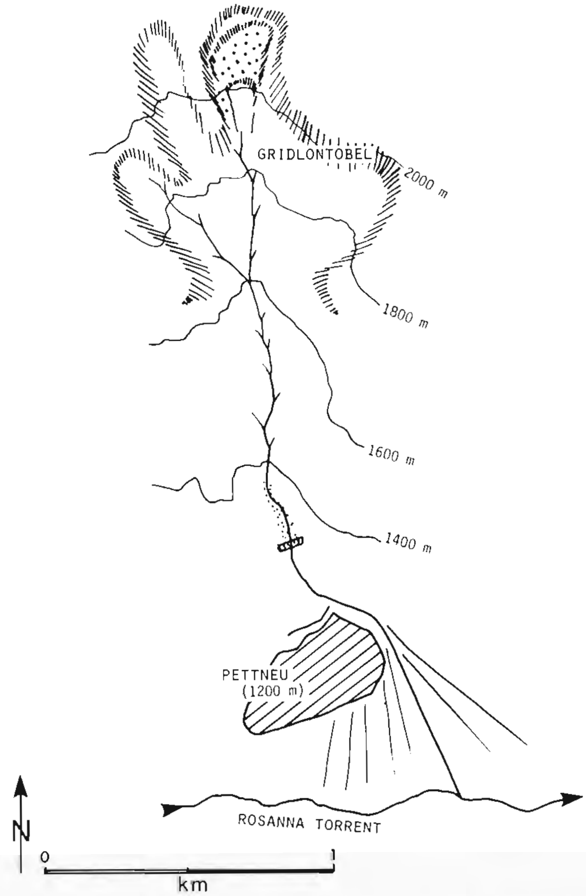


Figure 266: Sketch map of the Pettneu cone and the main debris source in the Gridlontobel basin.

Figure 267: a) View of Pettneu from across the Rosanna Torrent; note protective forest at the apex of the cone (GSC 204169-V). b) Discharge section of the protective dam above the Pettneu cone; source area of the 1968 debris flow in the upper right (GSC 204169-W).

a dozen abandoned buildings. The total volume of the flows was estimated to have been $0.3 \times 10^6 m^3$. A protective forest above the town prevented the debris from taking a more destructive path. After the catastrophe a large protective dam with a selective discharge section of steel beams was erected above the cone (Fig. 267). Its design capacity of $0.3 \times 10^6 m^3$ corresponds roughly to the volume of potentially unstable debris sources remaining in the upland (Moll, 1976). The people of Pettneu contributed much free labour to the restoration of their town and also planted 15 000 extra trees in the protective forest zone above the community.

Mattmark (A129)

Location: Saas Valley, Wallis (Valais), Switzerland (C2)
Date(s): 30 August 1965 (also 21 August 1633, 25 June 1829, 20 July 1868, 2 July 1968, 8 July 1970)

The Saas Valley is drained by the north-flowing Saaser Vispa, which joins the Rhone (Rotten) River at the town of Visp (Fig. 268). The lower part of the valley is a narrow bedrock gorge. In contrast, the upper section opens along several flats hosting the villages and hamlets of Saas. The glacially sculptured valley is rimmed by peaks and ridges of high grade metamorphic rocks of the Pennine core zone. They rise to elevations above 4000 m and are partly mantled by glaciers.

The Allalin Glacier overhangs the west side of the upper Saas Valley (Fig. 269a). It originates in an ice-filled cirque between 2800 and 3100 m and descends eastward over a bedrock slope inclined about 30° towards the Saas Valley. During historical advances the Allalin Glacier frequently crossed the bottom of the valley (2000 m) and blocked the torrential Saaser Vispa. The dam of ice and moraine which impounded the Mattmarksee (Lake Mattmark) intermittently posed a hazard to the communities of the lower valley. Between 1589 and 1808 the ice dam failed at least 20 times. The most serious flood occurred on 21 August 1633, when a wave of debris devastated forests and fields on the valley bottom and demolished 18 buildings in Visp. In 1834 a potential flood threat from the Mattmarksee was forestalled by the construction of an ice tunnel across the snout of the Allalin Glacier. This work was carried out by Ignatz Venetz, an engineer who had gained considerable experience during rock tunneling work along the Simplon Pass and at the ice lake of Mauvoisin (Richter, 1889b, 1891).

During the time interval between 1917 and 1924 the front of the Allalin Glacier again stemmed the flow of the Vispa and this time a spillway tunnel was blasted through the rock wall along the east side of the valley. Shortly thereafter the glacier retreated up the bedrock slope on the west. In the mid-1960s the snout of the Allalin Glacier once again advanced.

In 1965 work was in progress at Mattmark to impound the Vispa by means of a rock-and-earthfill dam, 120 m high, for a major hydroelectric reservoir. The two bulky lateral moraines of the Allalin Glacier, deposited during the Little Ice Age, supplied most of the fill for the projected dam. A construction camp was established halfway betwen the two morainal ridges — directly below the icy ramp of the advancing Allalin. This site had been chosen because it avoided the almost ubiquitous snow avalanche hazard of the upper Saas Valley; however, small ice avalanches triggered by slippage along the ice-rock interface of the glacier snout had been registered at the site in 1954, 1961, and 1963.

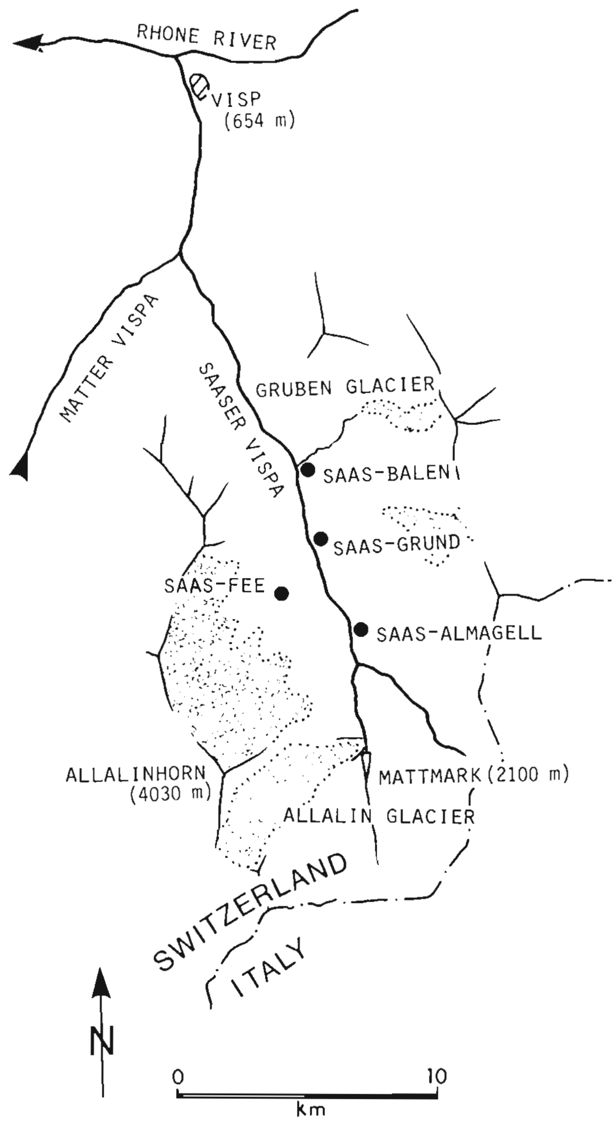


Figure 268: Index map of the Saas Valley and bordering glacier-covered mountain ridges.

During the summer of 1965 snowmelt was somewhat delayed but eventually proceeded rapidly. Then between 22 and 24 August 1965, a regional rainstorm deluged the upper Wallis region. Finally, on the evening of 30 August 0.8 to

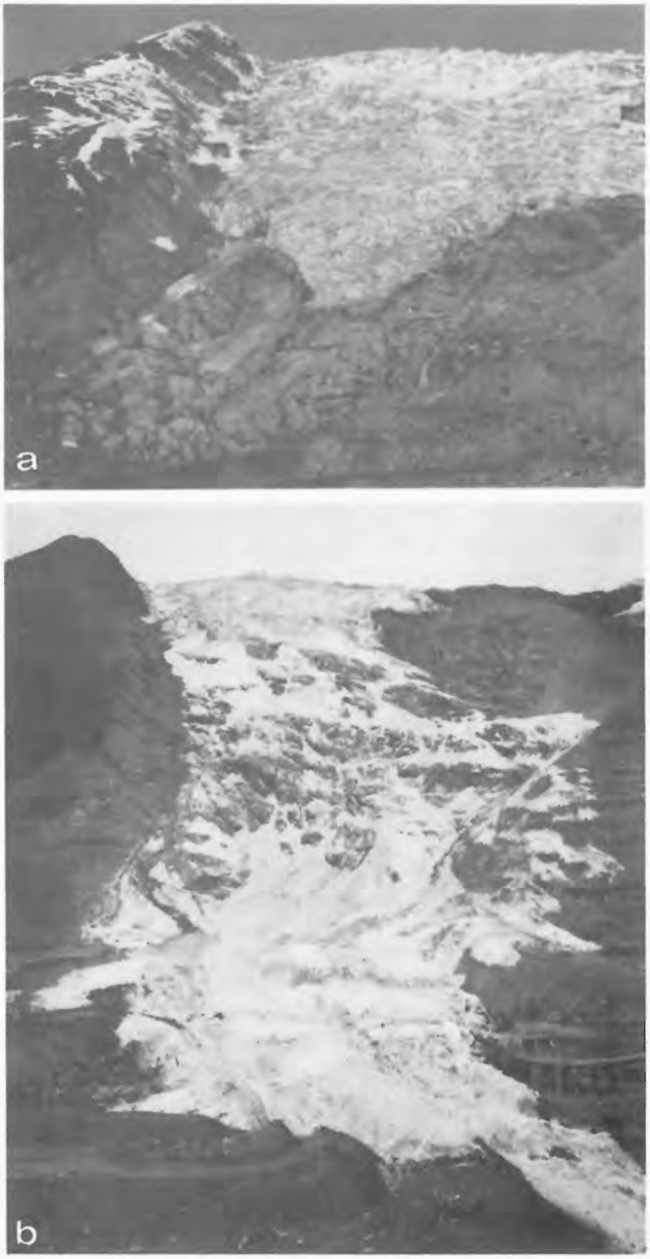


Figure 269: a) Allalin Glacier in 1981 (GSC 204169-X). b) The destructive Allalin ice avalanche of 1965 (Photo: ElectroWatt Zurich).

$1.0 \times 10^6 m^3$ of ice at the heavily crevassed front of the hanging glacier slid away. Leaving behind a downward concave fracture 40 m high at an elevation of 2500 m, the avalanche of ice, 100 to 200 m across, surged down the smooth rock slope and overwhelmed the construction camp located in its track. A powerful air blast flattened barracks and other timber structures before the camp disappeared underneath a blanket of pulverized ice and morainal debris (Fig. 269b). A total of 88 construction workers lost their lives in the ice lobe whose thickness reached 10 to 20 m (Vivian, 1966). The question as to whether or not the camp had been appropriately located led to a trial. After six years of litigation and hearings the accused parties were acquitted of the charge of negligence.

Careful studies since the catastrophe have indicated that failure occurred during a phase of enhanced slippage along the ice-rock interface. Such minor surge-like advances do not always lead to a failure of the glacier front. However, the catastrophic failure was the extreme manifestation of recurrent 'normal' slippage during warm summer periods (Röthlisberger, 1974).

In recent years another area of the Saas Valley has been threatened by a different kind of glacial process. The Gruben Glacier rests in a cirque at an elevation of 2800 to 3100 m east of the Saas Valley (Fig. 268). The high-gradient meltwater torrent from the glacier descends to the small debris cone hosting the community of Saas-Balen (1450 m). Along the front of the glacier three small ice-margin lakes drain subglacially. Blockage of the natural ice tunnels that connect the lakes set off violent bursts of water through new exits. Debris flows triggered by such sudden outflows damaged buildings on the cone on 25 June 1829, 20 July 1868 and 2 July 1968. The last of these attained a total volume of $0.4 \times 10^6 m^3$ and devastated a great part of Saas-Balen. Although construction of a drainage tunnel in the glacier was begun almost immediately after the 1968 disaster, another burst of water on 8 July 1970, almost thwarted the construction effort. A simple system of drainage tunnels and pipes has since prevented sudden discharge of large quantities of water into the glacial torrent (Röthlisberger, 1979; Lichtenhahn, 1979).

Zell am See (A130)

Location: Lake Zell (Zeller See), Salzburg, Austria (H1)
Date(s): 12 June 1966 (also 29 October 1567, 1598, 1632, 3 and 21 July 1737, 1817, 1834, 1879, 17 July 1884, 15 to 19 August 1966).

Zell am See (750 m) covers the fan delta of the Schmittenbach Torrent along the western shore of Lake Zell (Zeller See). From the fan delta, which has been inhabited for at least 3000 years, a relatively straight torrent channel leads into a bowl-shaped catchment area of converging ravines in intensely deformed phyllite of the Austroalpine basement complex and relict colluvium (Fig. 270). The semicircular crest of the basin culminates in the Schmittenhöhe (1900 m).

During the Middle Ages demand for timber in nearby mining districts and extensive overgrazing of the uplands contributed to an ever deteriorating process of bedrock slumping and erosional scour along the denuded slopes of the Schmittenbach basin. Erosional scars swept by powerful snow avalanches and deepened by storm runoff developed into notorious sources of debris referred to as 'Blaiken'.

The first destructive flows that demolished houses and public buildings in Zell am See occurred during regional rainstorms on 29 October 1567, in 1598, and 1632 (Hölzl, 1975).

On 3 July 1737, a violent cloudburst, centred on the Schmittenhöhe, set off lobes of blocky debris along the tributary ravines of the Schmittenbach Torrent. These lobes coalesced into pulsating flows that invaded Zell am See, demolishing many buildings and killing four people. On 21 July 1737, a second downpour brought additional devastation to the delta cone. After these disasters and a similar one in 1817 the townspeople began to protect themselves from further invasions of debris by constructing stone-masonry walls up to 250 m long along the torrent. Nevertheless, debris spilled over these structures during storms in 1834, 1879, and during a particularly devastating cloudburst on 17 July 1884. Deposits of these debris flows raised the level of the lower part of the town by 1.5 m and from then on the church had to be entered by steps leading down from street level rather than up as before (Fig. 271a). Aggradation in front of the delta suggests that on the average an annual discharge of approximately 20 000 m^3 of solids reached the lake during the 19th century. Beginning in 1886 the Schmittenbach basin was brought increasingly under control by systematic reforestation, check dams, and debris retention walls. As a result, the average annual debris load of the torrent decreased to 1200 m^3 (Hartwagner, 1954). During the last fifty years the town has become a major tourist centre and residential construction activity has expanded westward into the Schmittenbach basin (Fig. 271b).

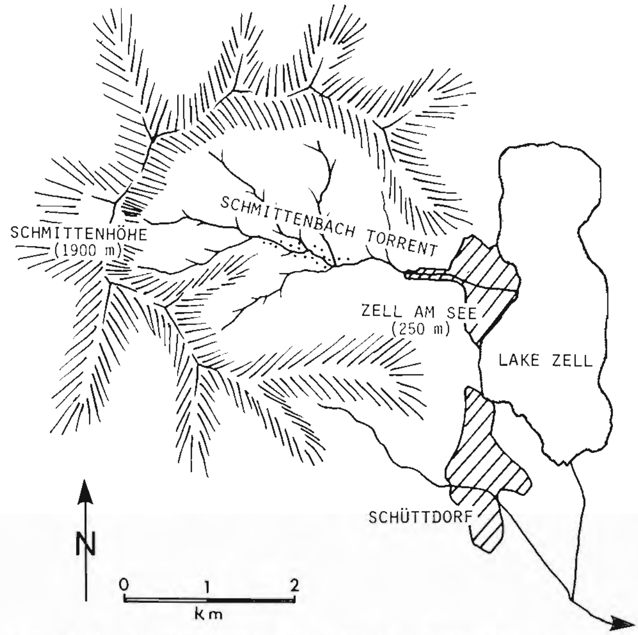


Figure 270: Sketch map of Lake Zell and the built-over fan delta of the Schmittenbach Torrent.

On 12 June 1966, a local hailstorm-downpour dumped about 65 mm of precipitation in 90 minutes on the converging slopes of the Schmittenhöhe amphitheatre. Intense scouring of debris chutes and shallow slope failures soon blocked the main channel of the torrent. Compounded by several log jams, massive pulses of debris invaded Zell am See. Some 70 buildings suffered heavy damage and one of the debris avalanches along the steep fringes of the town overran a building killing six inhabitants. A period of intense rainfall between 15 and 19 August 1966 triggered another series of debris flows and slope failures in the vicinity of Zell am See (Hölzl, 1975).

Today, development activities along the banks of the Schmittenbach Torrent and on steep slopes near the town continue unabated. New protective structures and extensive revetments have been added to the channel. Although the forest cover in the Schmittenbach basin has increased from 58% (in 1884) to 75% (in 1975), recent demand for new ski runs in the uplands has again reduced this value and thus increased the potential for erosional scour by storms and snow slides.

Figure 271: a) Entrance to the church of Zell am See below street level, an indication of aggradation by historical debris flows on the Schmittenbach cone (GSC 204169-Y). b) View into the intensely developed zone above the apex of the Schmittenbach cone; ski slopes and snow avalanche tracks in the background (GSC 204169-Z).

1966 Rainstorms (A131)

Location: Northern Italy and southern Austria
Date(s): 15 to 18 August 1966, and 3 to 5 November 1966

In the southeastern Alps the precipitation pattern of the summer of 1966 was characterized by local thunderstorms in June, followed by a succession of rainy days throughout July and August. Towards the middle of August the situation deteriorated markedly when cold air moved into the region from the northwest and collided with warm humid air masses that approached the mountains from the Adriatic Sea. Between 15 and 18 August some 200 to 300 mm of rain fell in southeastern Austria. Locally as much as 100 mm were registered within periods of 24 hours. Flooding along major rivers, debris flows from tributary basins, and numerous thin-skinned slope failures and debris avalanches resulted in great material damage and loss of lives (Kravogel and Wurzer, 1967; Schreiber and Zettl, 1967; Fuxjäger, 1975).

Then, between 3 and 5 November 1966, strong winds carried warm moist air from the Mediterranean against stationary cold air in northern Italy. All of northern Italy suffered from severe floods. The southeastern Alps again suffered from debris avalanches, debris flows, embankment failures, and torrential runoff. Precipitation that had begun as snowfall continued as warm rain, and locally reached between 200 and 700 mm (Fig. 272). Exhaustive documentation of the damage caused by mass movements during this storm resulted in considerable advances in the application of remedial measures against future rare events of this type (Guntschl, 1966; Martinis and Salvini, 1972; Venzo and Vaia, 1972; Castiglioni et al., 1974; Croce et al., 1974; Fuxjäger, 1975).

Siror (A132)

Location: Fiera di Primero, Val Cismon, Trentino, Italy (G2)
Date(s): 4 November 1966 (also 12th century)

The confluence of the south-flowing Cismon Torrent and the east-flowing Canale Torrent defines a gently sloping triangular surface whose apices are taken up by the communities of Fiera di Primero (710 m), Siror, and Tonadico (Fig. 273). Other villages in the area cling to the flanks of debris cones that spread from the mouths of steep gullies towards the Cismon Torrent. The walls of the Val Cismon are underlain by unstable phyllitic bedrock of the South Alpine basement complex and late Pleistocene ice-contact deposits. Ephemeral tributary torrents have carved ravines and gullies into these materials. The inclined surface between Siror and Tonadico probably owes its origin to slope failures and debris flows from the east side of the valley, one of which, according to tradition, annihilated a settlement sometime in the 12th century (Gorfer, 1977, p. 1026). For centuries the area between Siror and Tonadico was restricted to agricultural use.

On 4 November 1966, during the great rainstorm in the region, debris flows descended most of the seemingly harmless ravines and gullies of Val Cismon, and invaded the built-over fringes of Siror, Tonadico, and Mezzano. Scores of buildings were demolished.

Today, areas below ravines still are used mainly for agriculture, but recently vacation homes have expanded on terrain formerly avoided. Check dams have been erected across most of the gullies that show signs of slumping and scour (Fig. 274). Dykes and diversion dams also have been built at the apices of several debris cones.

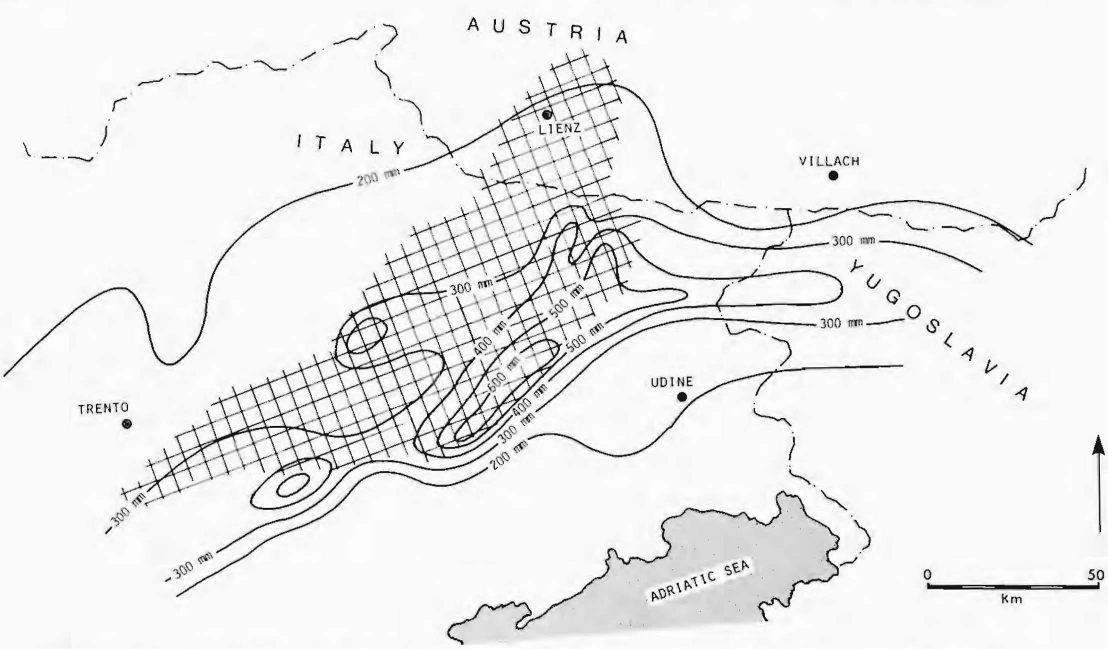


Figure 272: Isohyets of the 3-5 November 1966 rainstorm in the southeastern Alps; criss-cross pattern outlines zone with major destructive mass movements (see references in text for data sources).

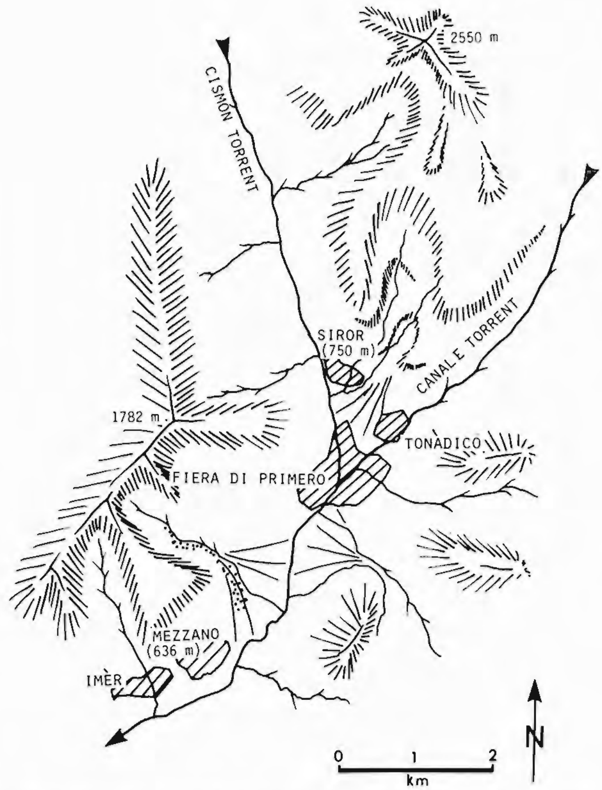


Figure 273: Sketch map of the Fiera di Primero – Siror area.

Figure 274: Recently completed array of check dams across a ravine above Siror which is carved into slump-prone relict colluvium and phyllitic bedrock. (GSC 204170-A)

Strigno (A133)

Location: Val Sugana, Trentino, Italy (G2)
Date(s): 4 November 1966 (also 1564, 31 August 1757, 3 August 1851)

Strigno (460 m) and Samone (680 m) are two of several communities dotting vineyard-covered terraces and benches on the north side of the fertile Val Sugana (Fig. 275). From the floodplain of the east-flowing Brenta River (420 m) debris fans and cones rise into tributary basins underlain by phyllite, granite, and carbonates of the South Alpine basement and cover complexes. Coarse late Pleistocene ice-margin deposits mantle the flanks of most upland gorges.

Samone rests on a south-facing terrace overlapped laterally by the small debris cone of the Cinaga Torrent. The generally dry bed of this torrent merges downward with the fan of the Chieppena Torrent. Devastations on the fan of the Chieppena Torrent are known to have occurred as early as 1564. Most of these flows seem to have been released by regional rainstorms, accompanied by rockfalls from carbonate ravines, debris avalanches from unstable terrace rims, and embankment failures along the toes of sagging slopes of phyllitic bedrock.

On 30 to 31 August 1757, a series of severe thunderstorms, following an extended period of rain, created one of the most memorable historical flood events in the front ranges of the southern Alps (Sonklar, 1883, p. 82-84). In Val Sugana, damage from debris floods was extensive. One of the squalls during this storm triggered a debris flow from a zone of fractured schistose bedrock below the granitic cliffs above Samone. The bouldery lobe of debris crashed into Samone demolishing 20 houses and killing four persons (Strele, 1936, p. 130). On 3 August 1851, during a major rainstorm, a debris flow from the same general source area apparently flattened five buildings in Strigno. The largest historical mass movements occurred during the great regional storm on 4 November 1966 (Fig. 276). Blocky material, mobilized mainly along tributary ravines of the Chieppena and Cinaga torrents, created two major debris flows that broke across the protective dykes on the Chieppena fan, destroying several houses in Villa and killing three people (Venzo and Largaiolli, 1968, 1969; Gorfer, 1977, p. 915-916).

Today many of the erosional scars created during the 1966 rainstorm are overgrown by thin stands of alder. The lower channels of the torrents have been provided with sturdy dykes and masonry revetments (Fig. 277).

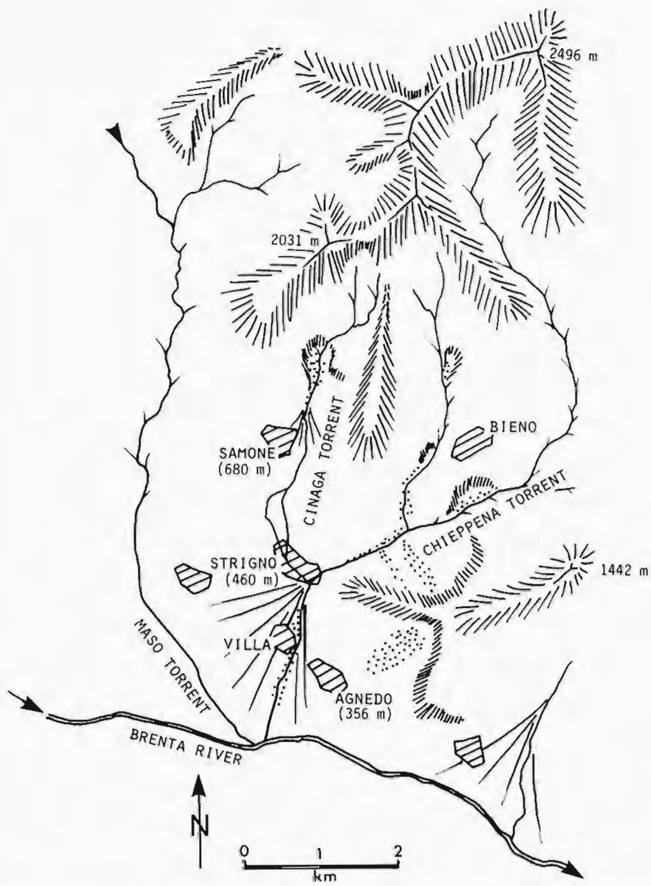


Figure 275: Sketch map of the densely populated wine-growing reach of Val Sugana near Strigno; note the composite debris fan of the Chieppena Torrent.

Figure 276: Aerial view of debris damage on the Chieppena fan in the autumn of 1966; note breached dykes at points of curvature in the channel (from Dragogna, 1975).

Figure 277: Reconstructed masonry-lined channel of the Chieppena Torrent below Strigno; note armoured sills defining the normal discharge section of the torrent. (GSC 204170-B)

Putschall (A134)

Location: Möll Valley, Kärnten (Carinthia), Austria (H1)
Date(s): 6 November 1966 (also 12 October 1889, 31 July 1900, 14 April 1901, 9 August 1902, 29 May 1917, autumn 1965)

The narrow upper Möll Valley is the southern approach route to one of the most daring road crossings of the Alps. Most of the area is underlain by metamorphic basement and cover rocks of the Pennine Tauern Window. The schistose metamorpic rocks are exposed along south-facing foliation dip slopes and steep north-facing escarpments. Debris cones and fans constrict the narrow floodplain of the torrential Möll River. Villages and hamlets in the valley are perched against south-facing dip slopes, on terraces of late Pleistocene morainal debris, and on debris cones. Many of the slopes underlain by phyllitic bedrock and surficial deposits suffer from sporadic snow-debris avalanches and bedrock slumps. Thus heavy rainstorm-snowmelt on 12 October 1889, and abrupt springtime runoff on 14 April 1901, and 29 May 1917, caused extensive damage to villages and transportation routes. During a violent rain-hailstorm on 31 July 1900 and a thunderstorm on 9 August 1901, large debris flows descended tributary ravines and blocked the Möll River, thus endangering several of the downstream communities (Stini, 1938).

Mass movements were widespread in the upper Möll basin during the autumn storms of 1965 and 1966. Devastating debris flows issued from the gorge of the Gradenbach Torrent, a western tributary northwest of Döllach (Fig. 278). The Gradenbach Torrent enters the main valley over a debris cone that hosts the hamlet of Putschall (1053 m). The source of most of the debris transported by the torrent is the toe of a sagging slope on the north side of the Gradenbach gorge, involving a total volume of more than $150 \times 10^6 m^3$ of phyllite and calcareous schist, dipping 50° to the south-southeast. The surface of the sagging slope is inclined 26°. Creep of bedrock reaches to a depth of 100 to 200 m and has led to the development of a diffuse head scarp below the Eggerwiesenkopf (2268 m). The toe of the creeping rock mass infringes on the channel of the torrent for a length of approximately 900 m.

Movement of the sagging slope along the Gradenbach Torrent was first recognized in the spring of 1917 when an abnormal snowmelt set off debris flows that covered the upper uninhabited part of the cone. In 1962 a subsidiary slump from the toe of the moving slope forced the evacuation of a farmstead located on it.

During the intense rainstorms of 1966 movement of the whole slide mass accelerated and much of the Gradenbach channel was blocked by masses of broken rock. Bursts of debris repeatedly fanned over the cone; by 6 November 1966, some $1.3 \times 10^6 m^3$ of bouldery material blanketed the

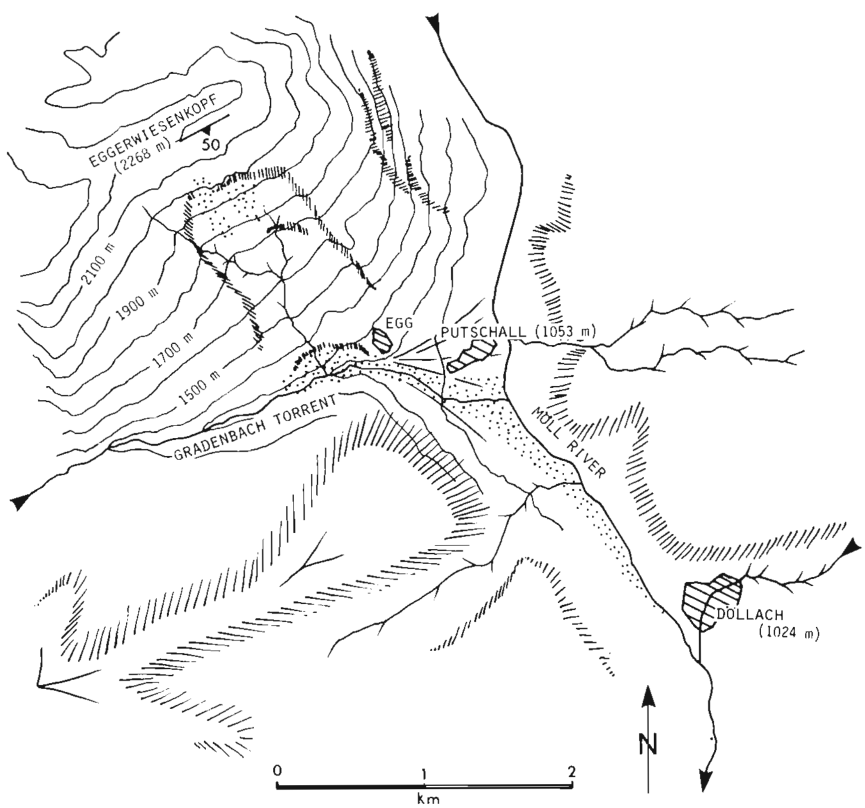


Figure 278: Sketch map of the sagging foliation dip slope along the north side of the Gradenbach Torrent and debris cone of Putschall.

Figure 279: Debris lobes on the Gradenbach cone engulfing the community of Putschall in the autumn of 1966 (from Kronfellner-Kraus, 1974).

Figure 280: Check dams across the Gradenbach Torrent; note damage to the abutments of the check dams caused by lateral pressure from the toe zone of the sagging Eggerwiesenkopf slope. (GSC 204170-C)

Figure 281: Putschall in 1981; note dense alder fringing the controlled channel of the Gradenbach Torrent. (GSC 204170-D)

cone, forcing the evacuation of Putschall (Fig. 279). Huge masses of bedload also choked the channel of the Möll River. After the catastrophe an extensive research project was initiated along the Gradenbach Torrent and on the sagging slope above (Kronfellner-Kraus, 1974, 1980; Horninger and Weiss, 1980; Moser and Glumac, 1983). Detailed surveys defined the principal zones of displacement on the sagging slope. Installation of piezometers coupled with local meteorological observations established that creep rates are directly related to infiltration of surface waters from rainstorms and snowmelt. Check dams in the Gradenbach gorge were found to be strained to the point of failure by the lateral pressure exerted by the slope (Fig. 280); their reconstruction and maintenance has been found to be extremely difficult (Kronfellner-Kraus, 1980). The hamlet of Putschall continues to exist on the cone (Fig. 281).

Inzing (A135)

Location: Inn Valley, Tirol, Austria (G1)
Date(s): 26 July 1969 (also 1701 , 31 August 1807, 1855, 1879, 24 July 1929)

The communities of Inzing (616 m) and Flaurling (675 m) are located on debris cones projecting northward in the wide Inn Valley west of Innsbruck (Fig. 282). The cones are drained respectively by the Enterbach and Kanzingbach torrents which originate along north-facing cirques of the Rosskogel Massif (2649 m). The bedrock underlying this massif is composed of well-foliated schists of the Austroalpine basement complex. Walls and floors of the cirques (e.g. Inzinger Alm) are sparsely vegetated and underlain by late Pleistocene morainal and ice-contact deposits locally more than 50 m thick. The two torrents are deeply incised into these deposits and, above their bedrock gorges, are flanked by steep colluvial embankments. Climatically, the Rosskogel Massif is part of a zone along the Inn Valley noted for its frequent summer thunderstorms (Aulitzky, 1970).

In 1701, slumps and debris avalanches in the surficial deposits of the Flaurlinger Alm triggered debris flows that buried the community of Flaurling, killing 12 people (Hanausek, 1975, p. 108). On 31 August 1807, an exceptional rainstorm, accompanied by warm winds, swept through the northern Alps, causing numerous debris flows. One of these flows erupted over the cone at Inzing, demolishing about 80 buildings. This catastrophe and others in the area inspired Aretin (1808) to a vigorous proposal for improved land and forest practices as a first step to prevent debris flow damage.

Following another series of debris flows along the Enterbach Torrent in 1855 and 1879, a large protective rock dam was built across the apex of its cone. Finally, after a massive flow on 24 July 1929, check dams were also installed in the debris source area below the Inzinger Alm.

On 26 July 1969, an extremely localized cloudburst-hailstorm deluged the Rosskogel Massif. Within 30 minutes this downpour produced 80 mm of precipitation. Rapid and concentrated runoff resulted in deep scour and debris avalanching in the ravines feeding the upper Enterbach basin. Debris brought in by tributary torrents combined with that derived from the high embankments of the main torrent to knock out several critical check dams. Two major pulses of debris burst forth onto the cone, depositing a total of $0.4 \times 10^6 m^3$ of bouldery material. The flows reached velocities of 14 to 17 m/s, erased 12 buildings and a public swimming pool in Inzing, and killed three people (Fig. 283). Total damage to the community was still small compared to what it might have been had expert advice not been heeded some time before the catastrophe when westward expansion of the town had been proposed in the wake of development pressures (Aulitzky, 1970).

Figure 282: Sketch map of the Enterbach and Kanzingbach basins near Inzing showing the debris sources consisting of late Pleistocene morainal and ice-contact deposits and relict colluvium.

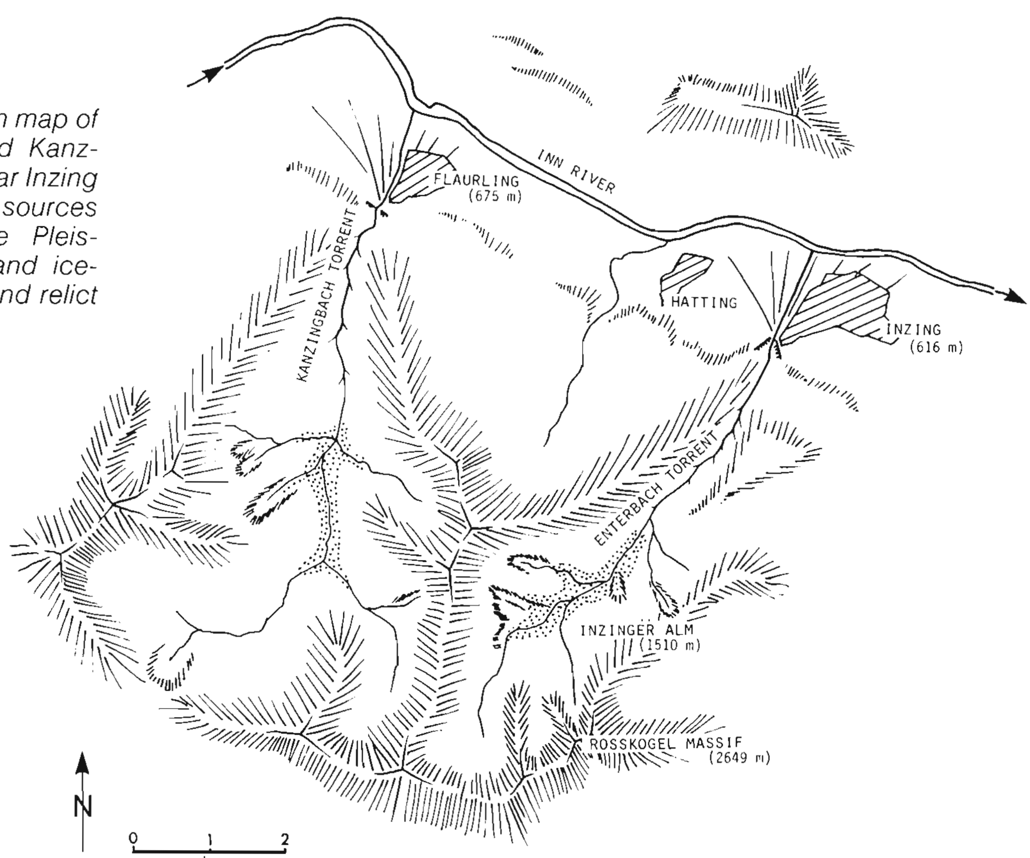


Figure 283: Impact of the 1969 debris flow on the Enterbach cone (photograph Alpine Luftbild); note the traditional land use pattern which assigned residential use to one flank of the cone and agriculture to the rest.

Figure 284: Protective dam above Inzing; note selective discharge section of steel beams ('filter dam') which allows passage of normal bedload. (GSC 204170-E)

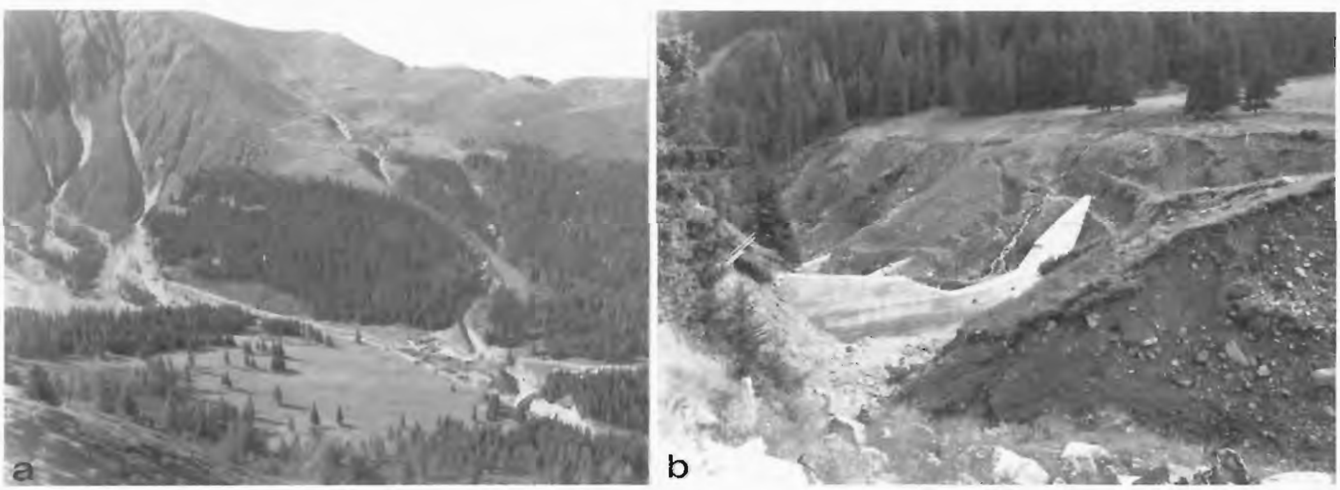


Figure 285: a) View of the crest of the Enterbach basin; note converging erosional ravines (GSC 204170-F). b) Steel-concrete check dam under construction on the Enterbach Torrent; note gently rising wings, deeply founded abutments, and drainage bores (GSC 204170-G).

After the 1969 disaster, the torrent channel on the cone was moved 150 m to the west away from Inzing. The protective dam above the community was converted into a steel-concrete-masonry structure with a retention capacity of more than $0.3 \times 10^6 m^3$ (Fig. 284). New arrays of check dams were built in the source area below the Inzinger Alm (Fig. 285), and about 10 000 m^3 of rock (used in the construction of the protective dam) were blasted away from a constriction in the lower gorge to avoid blockage of the channel during future flows (Hopf and Wanner, 1975).

Lecco (A136)

Location: Lago di Lecco, Lombardia, Italy (E3)
Date(s): February 1969

The city of Lecco (210 m) spreads over the eastern shore of Lago di Lecco astride a gently sloping fan delta. In recent years the city has expanded to the foot of the south-facing scarp slope of Monte San Martino (1157 m). This mountain is composed of intensely fractured north-dipping carbonate strata of the South Alpine cover complex. The lower part of the unstable cliff is mantled by extensive rockfall material.

In February 1969 a slab of dolomite 10 000 to 12 000 m^3 in volume, broke away from the joint-controlled face of Monte San Martino and scattering blocks claimed eight casualties in Lecco.

Subsequent model experiments and in situ rockfall tests (Broili, 1974; Fumagalli and Camponuovo, 1975) provided insights which were applied in the construction of protective walls at the foot of the slope.

Tagliamento Valley (A137)

Location: Friuli-Venezia Giulia, Italy (I2)
Date(s): 6 May 1976 and 15 to 17 September (also 15 August 1692, 2 November 1851, 28 October 1882, 22 to 23 August 1891, 13 to 14 September 1903, 27 March 1928, 5 November 1966)

The drainage basin of the south-flowing Tagliamento River in northeastern Italy is one of the most remarkable geomorphic-tectonic regions of the Alps. Over a distance of only 20 km the southern Alps rise from their foreland near the Adriatic Sea to mountain ridges 1000 to 2000 m high. The westerly trending ranges are composed mainly of carbonate and shale units of the South Alpine cover complex. In detail, the topography is controlled by west-trending thrust faults which separate north-dipping rock panels above an active zone of subduction and high seismicity (Fig. 286). Large earthquakes have occurred here in 1348, 1511, 1700, 1790, 1928, and 1976. The proximity of the mountains to the Mediterranean Sea also tends to cause blockage of warm maritime air masses by the cold mountain air, thus releasing intense rainstorms. On one hand slopes and ravines, shattered by seismic shaking, may experience slope failures during a violent rainstorm; on the other, cliffs weakened by excessive infiltration of water may fail during sporadic earthquakes. To make matters worse, many torrent basins in the region have suffered from overgrazing and deforestation. The Tagliamento River itself flows in a wide braided channel.

On 15 August 1692, following a period of heavy rainfall that apparently triggered widespread mass movements in the Tagliamento basin, the north-northwesterly facing dip slope below Monte Auda (1700 m) failed along a composite bedding-thrust surface dipping 30° towards the Tagliamento River (500 m). About $30 \times 10^6 m^3$ of dolomite surged down the mountain, crossed the river, and climbed 100 m up the northern flank of the valley (Fig. 287). The rock avalanche annihilated the village of Borta and killed at least 53 inhabitants (Fig. 288). It also stemmed the flow of the Tagliamento River, creating a lake almost 7 km long. After the lake had risen to more than 80 m above the bottom of the valley it overflowed in two major bursts on 4 and 20 October 1692. Remnants of the lake persisted for some time, and approximately 30 m of lacustrine silts were deposited in the former valley before vigorous downcutting by the Tagliamento River created a new channel across the rockslide and the accumulated alluvial-lacustrine deposits upstream from it (Cavallin and Martinis, 1974). A rock mass of approximately $10 \times 10^6 m^3$ lodged to the east of the failure surface below Monte Auda still represents a potential hazard, but at present does not directly threaten any permanent settlement.

The month of October 1851 was very rainy and many slopes became saturated with runoff. On 2 November a

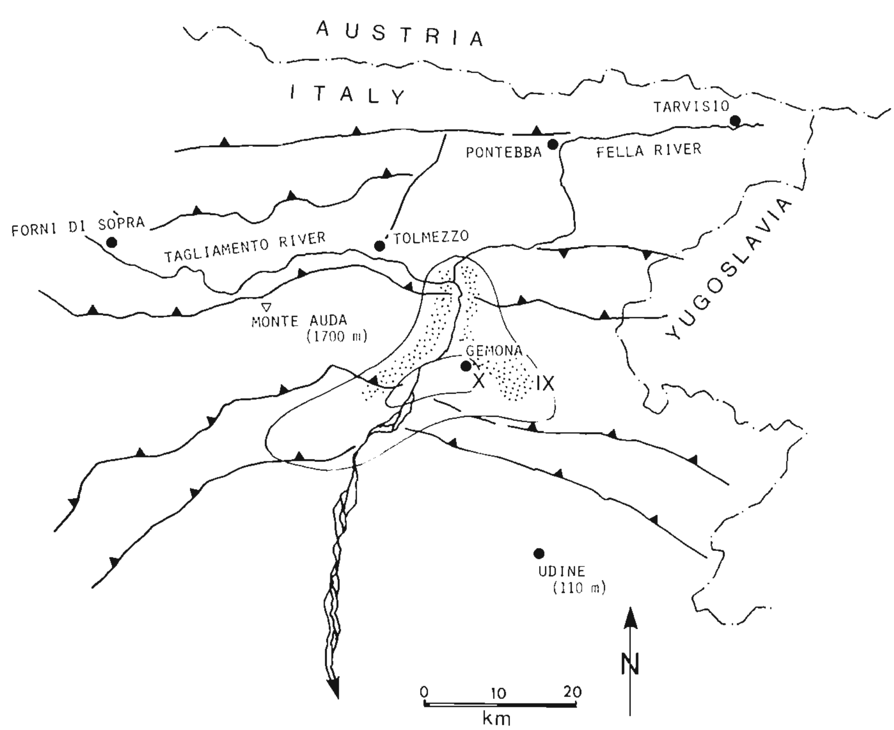


Figure 286: Index map of the part of the Tagliamento basin most seriously affected by the earthquakes of 1976 (Mercalli intensity zones 9 and 10). Major north-dipping thrust faults are shown as barbed lines. Area of greatest rockfall density related to the 1976 earthquake is stippled. Note location of Monte Auda.

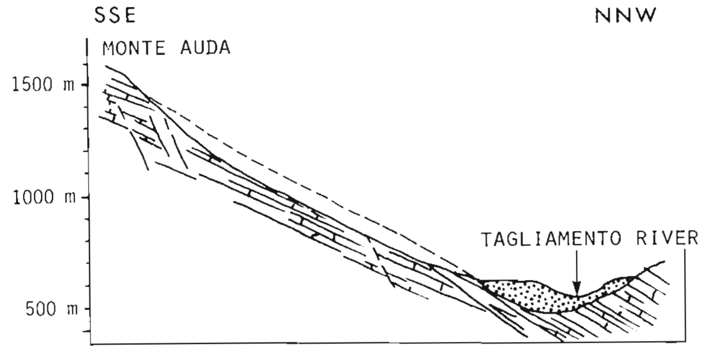


Figure 287: Cross-section of the north-facing dip slope of Monte Auda and the rock avalanche deposits of 1692 (after Cavallin and Martinis, 1974).

particularly intense storm deluged the tributary basins of the Tagliamento River. The storm possibly coincided with a minor earthquake. Numerous slopes failed. The bedrock terrace of Cazzaso, carved into a dip slope of argillaceous limestone 300 m above the town of Tolmezzo, began to show signs of creep. Although movement did not develop into catastrophic avalanching the crumbling community of Cazzaso shifted by more than 10 m and the sagging slope has remained a problem ever since. On the same day the Fella River was blocked by several debris flows from steep tributary basins; these flows claimed a total of more than 20 casualties. Near the village of Pontebba the channel of the torrential Fella rose 24 m due to aggradation behind embankment slumps (Stini, 1938, p. 26; Tosolini, 1974, p. 61-62).

On 28 October 1882, during the second pulse of a regional rainstorm there were many debris flows and debris avalanches in the Tagliamento region. One of them struck the outskirts of the village of Forni di Sòpra, killing three people. On 22 to 23 August 1891, a catastrophic downpour, releasing 300 mm of rain in 24 hours, again triggered massive flows from small catchment areas in the upper Tagliamento basin, threatening the existence of several hamlets. At Tarvisio six houses were toppled. A similar series of thunderstorms between 13 and 14 September 1903, caused severe debris floods; along the Fella River a hamlet disappeared underneath a blanket of carbonate rubble (Stini, 1938, p. 33 and 36; Tosolini, 1974, p. 62).

On 27 March 1928, an earthquake shook the mountains south of Tolmezzo, releasing some hundred rockfalls and minor rock avalanches in the epicentral area. The rockfalls were mapped by Gortani (1928) in what can be considered a pioneering study of its kind. The type of ravelling along the ravines of high scarp faces depicted in his photographs and shown on his map are remarkably similar to those published after the earthquakes of 1976. Some of the slabs that splintered away from the cliffs disintegrated into dry rock avalanche lobes.

On 5 November 1966, the great rainstorm that deluged the southern Alps also affected the upper Tagliamento basin.

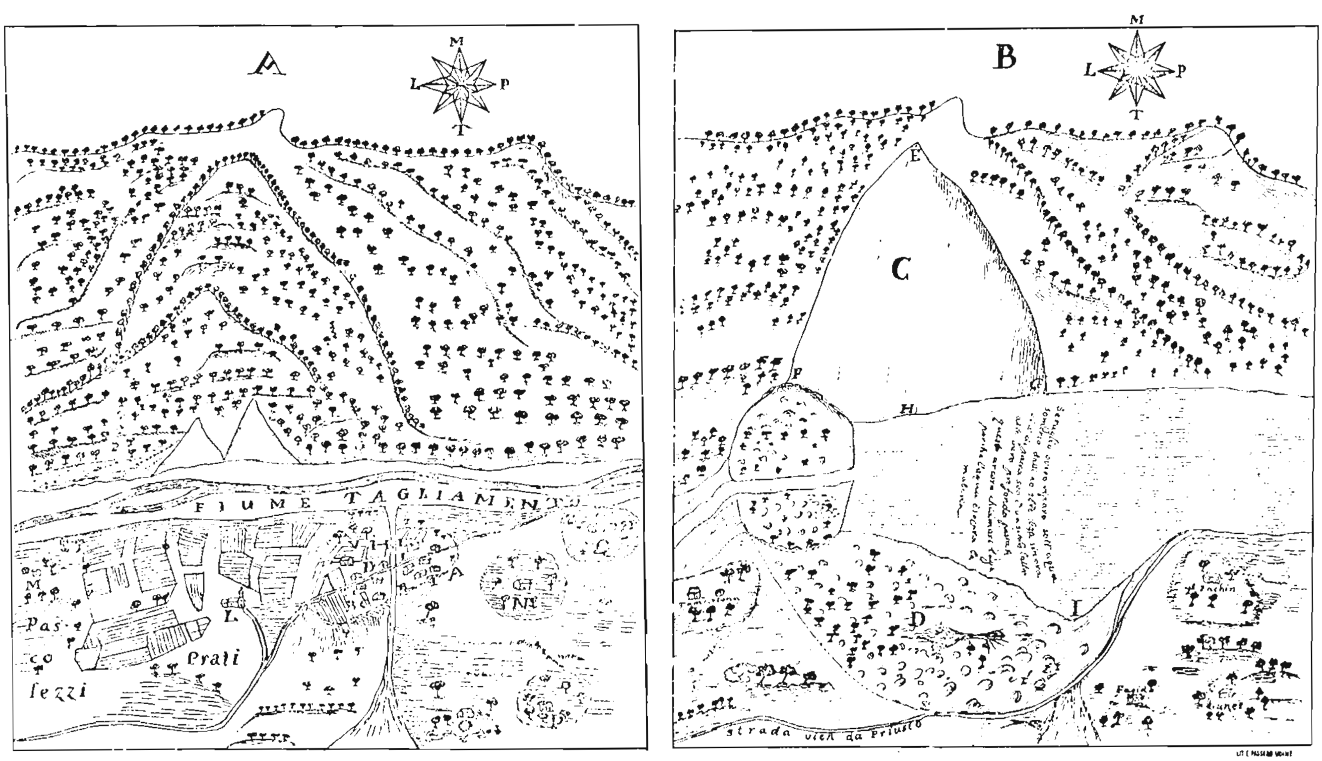


Figure 288: Before a) and after b) sketches of the 'Frane di Borta' by Pascolo Pascoli made one month after the catastrophe (from Cavallin and Martinis, 1974).

Numerous transportation routes and several buildings were completely covered with blocky debris (Martinis and Salvini, 1972).

On 6 May, 15 September, and 17 September 1976, the Gemona area experienced three fierce earthquakes with magnitudes ranging between 6 and 7 ('Friuli Earthquake'). In the epicentral area, straddling the lower reaches of the Tagliamento, numerous rock avalanches and rockfalls from scarp faces of intensely fractured carbonate bedrock descended into steep bedrock ravines. Some of the rock avalanches exceeded a volume of $0.1 \times 10^6 m^3$. Near Gemona, scarplets and cracks, some in excess of 1 km in length, were observed along several mountainsides. Some of the rockfall deposits which accumulated as steep cones during the first earthquake were remobilized into spreading lobes during the second series of earthquakes between 15 and 17 September 1976 (Govi and Sorzana, 1977).

Rebuilding of the towns and transportation routes after the Friuli earthquake also included work to control mobilization of unstable debris lobes and to protect buildings underneath overhanging fractured rock faces. Most of these tasks will remain difficult challenges for some time to come (Querini, 1980).

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