Chapter 9

Natural Hazards in Forests: Glacier and Permafrost Effects as Related to Climate Change

Wilfried Haeberli[1] and C.R. Burn[2]
[1]Department of Geography, University of Zurich, Switzerland
[2]Department of Geography and Environmental Studies, Carleton University, Ottawa, Canada

Atmospheric warming is predicted to be greater in polar regions than at lower latitudes and more pronounced at high altitudes than in lowlands. In polar regions, air and ground warming may lead to a more northerly extension of the boreal forest, as growing seasons lengthen and become warmer. Near-surface permafrost degradation will probably accompany such an evolution in environmental conditions. In some cases slope movement may be catastrophic, but in most instances the settlement is expected to be slow, and the water released by melting ground ice will evaporate. Subpolar forests in permafrost regions are primarily used for firewood and rough lumber, but not construction-grade materials. This is unlikely to change because long-term ground instability, relatively cold soil temperatures, depletion of nutrients in the active layer, and restriction of root systems to the active layer all limit tree growth. Extensive forest fires, usually initiated by lightning after a week or two of hot weather, also deplete timber stocks. There have been suggestions that wildfire may increase following climate warming. Meltwater runoff from glaciated and perennially frozen areas represents only a small portion of the annual water supply, but strongly influences stream flow in lowlands during the warm/dry season. The disappearance of perennial ice above and below the earth surface influences the seasonality of discharge by reducing meltwater production in the warm season and by increasing the permeability of frozen/thawing materials. The latter effect may have strong impacts on soil humidity and growth conditions for forest and tundra in such dry areas as Tian Shan Mountains or Tibet Plateau. In general, accelerated would cause an enlargement of the periglacial belt in high mountain areas, an upslope shifting of hazard processes and a widespread reduction in the stability of formerly glaciated or perennially-frozen slopes. In the case of accelerated future warming, the cryospheric components of high mountain environments would be expected to change at high rates and lead to pronounced disequilibria in the water cycle, in mass-wasting processes and sediment flux, as well as in growth conditions for vegetation. For those directly involved with such changes, the main challenge would be to adapt to the high rates of environment evolution. Empirical knowledge would have to be replaced more and more by improved process understanding, especially concerning runoff formation and slope stability.

Introduction

Earth's snow and ice cover is a critical component of the global climate system. The continental ice sheets of Greenland and Antarctica and the extent of sea-ice actively influence circulation of the atmosphere and the oceans (Budd *et al.*, 1998). The varying spatial extent of glaciers and ice caps through time indicates past changes in surface energy fluxes (Haeberli *et al.*, 1999a), particularly fluctuations in summer temperature and precipitation. In contrast, the duration, thickness and extent of snow cover and freshwater ice provide information about winter conditions. The climatological roles of snow cover are to nourish glaciers, reflect solar radiation reaching the ground surface and regulate ground temperatures in winter.

Permafrost, or perennially-frozen ground, can also contain large amounts of ice. The principal ecological role of permafrost is to provide a near-surface hydrologic barrier that directly influences soil moisture and runoff within the active layer, the ground above permafrost that freezes and thaws each year (French, 1996). Most soil microbial activity and plant root development occurs in the active layer. A potential positive feedback of increasing summer temperature is to alter high-latitude ecosystems from their current status as net sinks for greenhouse gases to sources of methane and carbon as the soil warms (e.g., Weller *et al.*, 1995).

Climate projections indicate that there could be pronounced reductions in the extent of seasonal snow, floating ice and glacier cover during the next century, while the permafrost active layer may deepen considerably. Such changes would probably have significant impacts on related ecosystems; both engineering and agriculture would need to adjust to changes in snow and surface and ground ice conditions (Fitzharris *et al.*, 1996). In order to document potential changes, to assess their impact on ecosystems and socio-economics, and to develop strategies for diminishing potentially-harmful effects or amplifying beneficial ones, global climate observation networks have included cryosphere components in their comprehensive, integrated view of the climate system (Cihlar *et al.*, 1997).

The design of climate-related modelling, observation, mitigation and adaptation strategies involves complex interactions among individual components of the entire climate system. This chapter reviews the main impacts on forest ecosystems of glacier and permafrost evolution during climate change. First, characteristics and the distribution of these two phenomena are outlined. Second, fundamental aspects of their response to climate change are explained. Third, the most prominent effects of such developments on forests in high-mountain and subpolar regions are discussed.

Distribution and Characteristics of Glaciers, Permafrost and Forests in Cold Regions

The distribution of glaciers, permafrost and forests is primarily a function of mean annual air temperature and annual precipitation (Fig. 9.1). In humid-maritime regions, the equilibrium line, separating accumulation from ablation areas on glaciers, is at low altitude, because of the large amount of ablation required to eliminate the deep snowfall (Shumskii, 1964). Temperate glaciers dominate these landscapes. Such ice bodies, with relatively rapid flow, exhibit a high mass

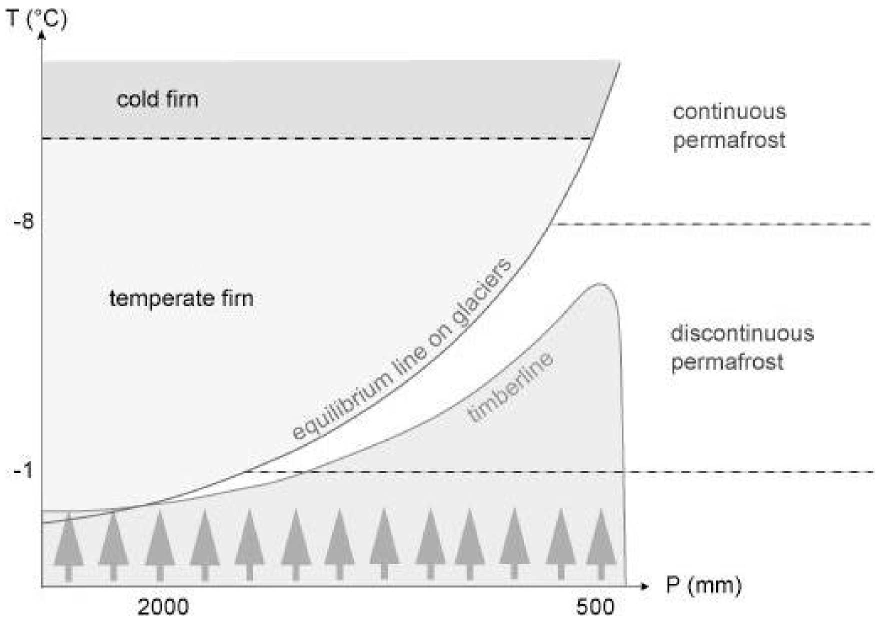

Fig. 9.1. Schematic diagram of glacier, permafrost and forest limits as a function of mean annual air temperature and average annual precipitation. Forests verge on glaciers in humid-maritime climates and grow above permafrost in dry-continental areas (after Shumskii, 1964; Haeberli *et al.*, 1989).

turnover and react strongly to atmospheric warming by enhanced melt and runoff. The ice caps and valley glaciers of Patagonia and Iceland, the western Cordillera of North America and the mountains of New Zealand and Norway are features of this type (Fig. 9.2). The lower parts of such temperate glaciers may extend into forested valleys, where summer warmth and winter snow accumulation prevent development of permafrost. In contrast, under dry-continental conditions, such as in northern Alaska, arctic Canada (Fig. 9.3), subarctic Russia, parts of the Andes near the Atacama desert, and in many central Asian mountain chains, the equilibrium line may be at high elevation. In such regions, polythermal or cold glaciers, lying far above the treeline, have a low mass turnover and are associated with periglacial conditions and permafrost (Shumskii, 1964).

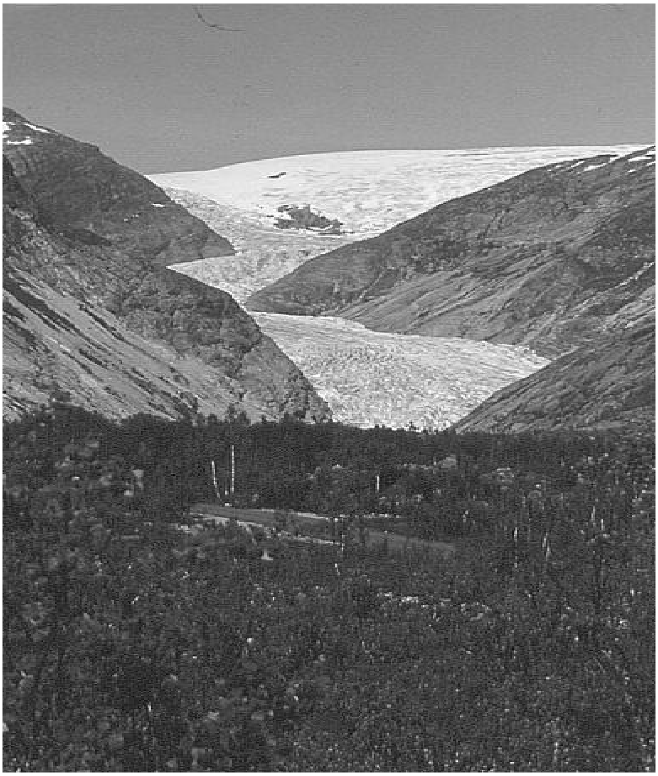

Fig. 9.2. Nigardsbreen, on the east side of Jostedalsbreen ice cap, southern Norway, where precipitation from the Atlantic Ocean is abundant. The glacier descends into permafrost-free, forested terrain. Photograph by W. Haeberli, 1999.

Fig. 9.3. Cold polar glaciers on Axel Heiberg Island, Canadian High Arctic. These glaciers are surrounded by cold, thick permafrost, and lie far from the treeline in an area of polar desert. Photograph by J. Alean, 1977.

An extensive database of the topographic characteristics of glaciers exists in regional glacier inventories (IAHS(ICSI)/UNEP/UNESCO, 1989) and serves as a basis for extrapolating results from observations or modelling of individual glaciers to the regional scale (Oerlemans, 1994). Meier and Bahr (1996) estimated the total number (160,000), area (680,000 km^2), volume (180,000 km^3) and sea-level equivalent (0.5 m) of ice caps and mountain glaciers on Earth and the relative contributions from polar, subpolar, low-latitude/maritime and low-latitude/continental climatic regions (Fig. 9.4). Large areas are covered by ice in the arctic and subarctic mountains of northern Canada and Alaska (275,000 km^2) and the mountains of Eurasia (240,000 km^2), with most of the remaining glaciated area in South America, New Zealand, the subantarctic islands, Iceland and the periphery of Greenland. In general, glaciated area tends to be directly related to distance from densely populated areas, while the economic significance of glaciers is inversely related to this distance. Large maritime glaciers and ice caps provide the main meltwater contribution to present increases in sea level, while meltwater from continental glaciers often influences runoff variability in interior regions of the continents (Kotlyakov and Krenke, 1982).

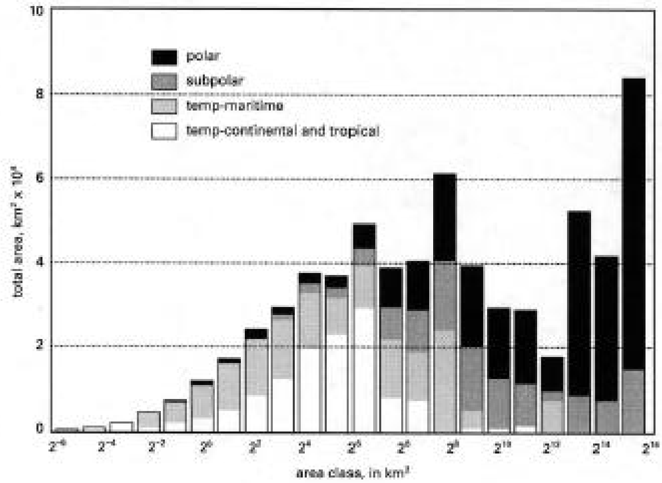

Fig. 9.4. Histogram showing the total global area of glaciers as a function of glacier size. The histogram is divided into four broad climatic regions (modified from Meier and Bahr, 1996).

More than half of Canada and Russia, most of Alaska, a quarter of China and considerable parts of cold mountain ranges worldwide are underlain by permafrost (Brown *et al.*, 1997). The terrain is especially sensitive to disturbance where permafrost is ice-rich, usually in fine-grained sediments, particularly glaciolacustrine and loessal deposits (Brown, 1997a). Coastal and river erosion accelerate in ice-rich ground, while thaw settlement occurs in association with surface disturbance. Ice-rich soil and deeper materials on hill slopes exhibit long-term down-slope movement (Foriero *et al.*, 1998), while deep-seated deformation of ice-supersaturated debris on mountain slopes leads to the development of rock glaciers (Fig. 9.5; Berthling *et al.*, 1998; Hoelzle *et al.*, 1998; Kääb *et al.*, 1998; Haeberli *et al.*, 1999b). The construction of settlements, airfields, pipelines, roads, mines and power dams in such areas requires specialized design techniques. Permafrost in forested areas (Fig. 9.6) is usually discontinuous, underlying less than 80% of the ground. Over large areas of such discontinuous permafrost, the perennially-frozen ground is warmer than -5°C and often warmer than -2°C. Such ground may be particularly sensitive to climate warming which threatens to raise its temperature above 0°C.

Fig. 9.5. Actively creeping ice-rich mountain permafrost ("rock glacier") Suvretta near St Moritz, Swiss Alps. Photograph by W. Haeberli, 1990.

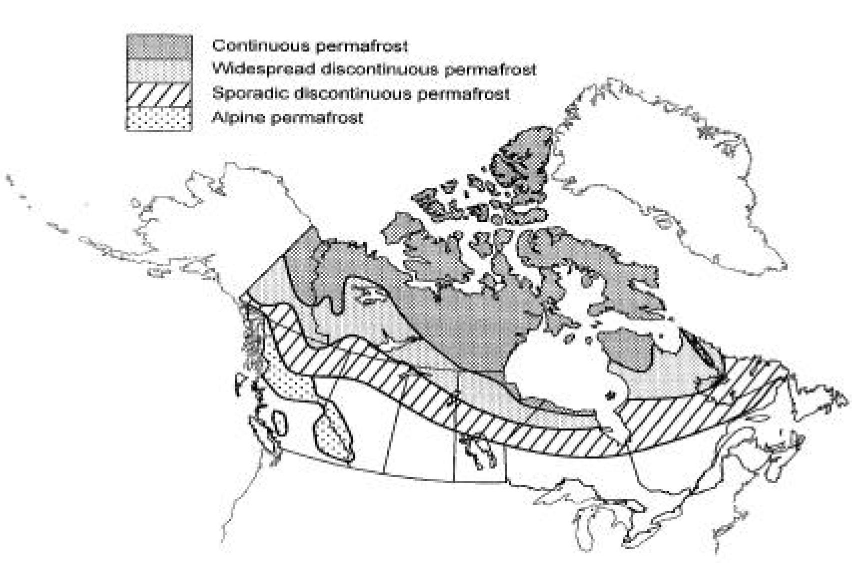

Fig. 9.6. (A) Permafrost map of Canada

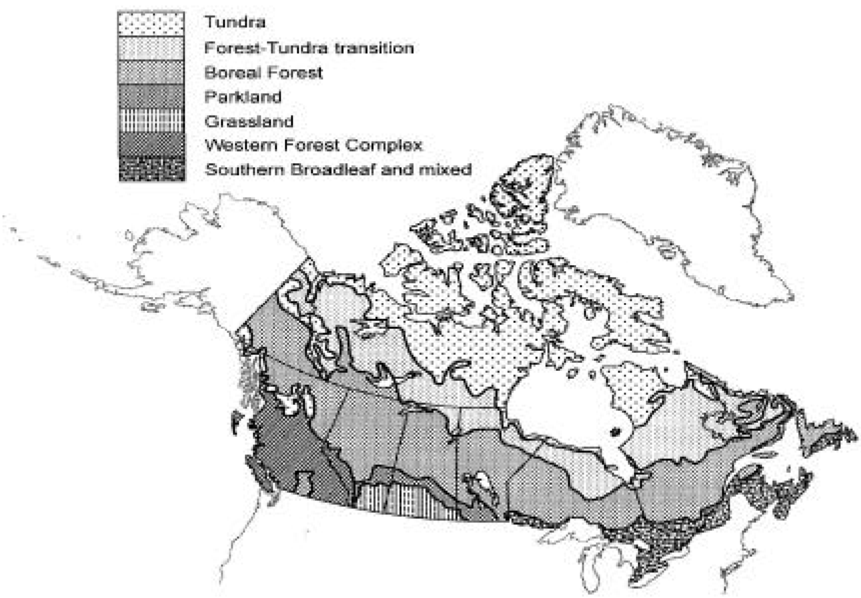

Fig. 9.6. (B) Vegetation map of Canada. Both maps are generalized from the National Atlas of Canada.

Climate Change Effects on Glaciers and Permafrost

Glaciers and discontinuous permafrost may react sensitively to atmospheric warming because of their proximity to the melting temperature. The direct response of glaciers to climate change occurs through changes to the mass balance, and ultimately through variations in glacier length and size (Johannesson *et al.*, 1989). Rates and ranges of such glacier changes can be determined quantitatively over various time intervals and expressed as corresponding energy fluxes. As a result, shifts in glacial cover in mountain areas are key indicators for the early detection of climate change (Haeberli, 1996). Internationally-coordinated long-term monitoring of glaciers started in 1894 and today involves collection and publication of standardized information on the distribution and variability of glaciers in space and time, particularly fluctuations in mass balance and length. Data on special events, such as catastrophic changes, are also available (IAHS (ICSI)/UNEP/UNESCO, 1998).

The response of permafrost to climatic warming occurs through melting at the top of permafrost over periods of years, with or without changes in active-layer thickness. Warming of the temperature profile within permafrost occurs over decades to centuries, and upward displacement of the base of permafrost to reach a new equilibrium thickness takes from centuries to millennia (Osterkamp and Gosink, 1991; Burn, 1998a). In addition to direct climatic effects, principally through changes in air temperature and snowfall, local ground temperature is also strongly influenced by factors related to climate, such as the type of vegetation and properties of the organic layer and soil. These factors can interact in various ways, making it difficult to predict the overall effect of climatic changes (Smith and Riseborough, 1983; Goodwin *et al.*, 1984). Modification of permafrost conditions and subsequent adjustment of related biotic and abiotic surface processes takes place over variable and extended time periods. First attempts are now being made to monitor active-layer thickness and permafrost temperatures within global climate-related observing systems (Cihlar *et al.*, 1997). Efforts presently concentrate on monitoring the thickness of the active layer (Brown, 1997b) and on obtaining borehole temperatures (Isaksen *et al.*, 2000).

Long-term observations of glaciers have provided convincing evidence of rapid climatic change at the global scale; the retreat of mountain glaciers during the 20th century was striking, worldwide. Characteristic average rates of glacier thinning were a few decimetres per year for temperate glaciers and centimetres to a decimetre per year for glaciers in continental areas with firn areas below melting temperature (Meier and Bahr, 1996; Haeberli *et al.*, 1999a). The total retreat of glacier termini is commonly measured in hundreds of metres to a few kilometres, and at retreating glacier termini, the total secular surface lowering is up to several hundred metres (Blair, 1994; Haeberli, 1996; Haeberli *et al.*, 1997). The apparent homogeneity of the signal at the secular timescale, however, contrasts with the large variability at local/regional scales and over time periods of years to decades (Letréguilly and Reynaud, 1990).

Intermittent periods of mass gain and glacier advance are reported from various mountain chains, especially in areas of abundant precipitation such as southern Alaska, New Zealand and Norway (IAHS(ICSI)/UNEP/UNESCO, 1998). Glaciers in the European Alps, on the other hand, have lost ≈ 30 to 40% of their surface area and ≈ 50% of their volume since the mid-19th century, the end of the "Little Ice Age" (Haeberli and Hoelzle, 1995). The recent emergence of a stone-age man from (probably) cold ice on a high-altitude ridge of the Oetztal Alps is a striking illustration which confirms that the extent of Alpine ice is probably less today than during the past 5,000 years (Haeberli *et al.*, 1999a).

Permafrost has also been affected by the recent warming, but its secular evolution is much less understood. Temperatures collected from deep boreholes in Alaska, northern Canada and Europe indicate warming during the past century (Lachenbruch and Marshall, 1986; Taylor, 1991; Isaksen *et al.,* 2000). Permafrost temperatures now appear to be increasing in the European Alps, the Kazakh and Kirghiz Tien Shan, the Tibet Plateau and most of Alaska, though not in the adjacent Yukon Territory (Canada) (Jin *et al.*, 1993; Burn, 1998a; Haeberli *et al.*, 1998; Vonder Mühll *et al.*, 1998; Osterkamp and Romanovsky, 1999).

Glaciated and perennially-frozen regions may be among the most affected if global warming accelerates, but potential changes can only be roughly estimated. Statistical relations and energy balance considerations indicate that one-third to one-half of the current mountain glacier mass could disappear over the next 100 years with a warming of 4°C (Fitzharris *et al.*, 1996; Warrick *et al.*, 1996). With an upward shift of the equilibrium line by some 200 to 300 metres, yearly thickness losses of 1 to 2 metres are expected from temperate glaciers, and many low-latitude mountain chains such as the European Alps would loose major parts of their glacier cover within decades. Large glaciers such as those around the Gulf of Alaska, in Patagonia and in the Himalayas would continue to exist – although greatly reduced – into the 22nd century. Warming of cold firn areas at high altitudes and high latitudes would be pronounced. Various scenarios for the extent of equilibrium permafrost following climate warming indicate that the lower limits of permafrost occurrence in mountain areas could rise by several hundred metres (Fitzharris *et al.*, 1996). However, the time for permafrost to reach new equilibrium thickness may be on the order of centuries, as the elevation of the permafrost base is accomplished by the relatively small geothermal flux (Osterkamp and Gosink, 1991). This time is extended by at least an order-of-magnitude for ice-rich permafrost, where the change in thickness may eventually be tens of metres. Convective effects from groundwater circulation may accelerate rates of permafrost degradation, but these are unpredictable at present and usually site-specific in scale, although some regional-scale effects have been recognized in northern Alaska (Deming *et al.,* 1992).

Impacts in Low-latitude Mountain Ranges

Glacier and permafrost belts within high-mountain ranges are characterized by extremely steep slopes, large amounts of coarse debris and intense erosion/sedimentation processes. Climate change-related impacts from glaciers and permafrost on forested areas in cold mountain chains can be direct, as in the case of natural hazards to inhabited land, or indirect via changes in landscape evolution, especially with respect to the water cycle and sediment yield. Direct hazards primarily involve the advance or retreat of glaciers, rock falls from destabilized mountain walls, ice avalanches, mudflows from glacier-clad volcanoes, floods from ice- and moraine-dammed lakes, and debris flows from steep permafrost slopes or breaching of moraines (Haeberli *et al.*, 1989, 1997; Major and Newhall, 1989; Evans and Clague, 1993; O'Connor and Costa, 1993; Walder and Costa, 1996; Richardson and Reynolds, 2000).

Glacier advance and retreat

Glacier advances into forested and agricultural land have historically been a major threat in densely-populated mountains (Tufnell, 1984; Grove, 1987). Today, glaciers are predominantly in retreat and the main significance with respect to hazards is the removal of cover from easily eroded moraine slopes, loss of buttresses from rock walls, and the formation of ice- and moraine-dammed lakes. Depending on the availability of quantitative information, predictions of glacier behaviour over a few years must be made by visual inspection in the case of unmeasured glaciers. Using continuity considerations, the behaviour of roughly parameterized glaciers may be estimated over the dynamic response time, typically periods of decades. Predictions may also be made by extrapolation of trends from measured time series. The reliability of such predictions primarily depends on the mass-balance scenario applied (Haeberli *et al.*, 1989). Satellite images have been used to document a glacier surge in connection with the formation of a lake threatening extended areas near Mendoza, Argentina (Espizua and Bengochea, 1990). In regions with maritime climatic conditions, where glacier tongues extend below treeline, recolonization of deglaciated areas by individual trees takes place after a few years, while decades are required to establish closed forests (Burga, 1999). Forest regrowth in deglaciated areas is likely to continue in the future if accelerated atmospheric warming takes place, but lack of soil, unstable slopes and harsh microclimatic conditions may prohibit fast regrowth of vegetation in some places, leaving extended areas of uncovered debris.

Rock falls

Glacier retreat and changes to permafrost may destabilize oversteepened rock walls and accelerate cliff retreat in high-mountain areas. Field and laboratory experiments and analyses of rock glaciers indicate characteristic cliff recession rates on the order of mm per year over the Holocene (Barsch, 1977; Haeberli *et al.*, 1999b). The processes involved, frost shattering and rock fall, have highly variable time and depth scales (Ødegard and Sollid, 1993; Matsuoka *et al.*, 1998). The influence of permafrost thaw on the destabilization of rock walls is thereby a virtually untouched field of research (Dramis *et al.*, 1995). The most important processes concern: (i) fracturing of rocks during seasonal and multi-annual freezing; (ii) changes in hydraulic conductivity, pore-water pressure and circulation during freezing and thawing; and (iii) changes in surface geometry from major rockfalls. Permafrost degradation may cause unfavourable changes in hydraulic conductivity leading to the onset of convective heat transfer and pore-water pressure variations that escape detection as fissures open at depth (Haeberli *et al.*, 1997). Towards the snout of valley glaciers, the pronounced lowering of ice surfaces and vertical loss in valley filling induces a change in the stress field inside the confining mountain walls.

On slopes shaded from direct solar radiation, the lowering of glacier surfaces enables the penetration of permafrost and the formation of ice in rock walls originally covered by temperate ice (Blair, 1994; Wegmann *et al.*, 1998). The penetration of the freezing front into previously unfrozen material has the potential to intensify rock destruction through ice formation in cracks and fissures. Such ice formation, in turn, reduces the near-surface permeability of the rock walls and affects hydraulic pressures inside the fissured rocks. In this way, the general lowering of pore-water pressures in lateral rock walls accompanying the disappearance of temperate glaciers may be counteracted and the rock-wall stability altered. In the absence of permafrost, at low altitudes and on sunny slopes, the stress field within lateral valley walls evolves as water pressure is reduced with decreasing ice thickness. Slope stability may become critical in many instances (McSaveney *et al.*, 1992; Evans and Clague, 1993; McSaveney, 1993; Blair, 1994). The lowest reported values for the overall slope of the run-out path (Evans and Clague, 1988, 1990) indicate that the mobility of rock avalanches in glacial settings may be enhanced by travel on low-friction snow and ice surfaces, generation of basal pore pressure by frictional heating, fluidization through snow- and ice-melt, and channelling or air launching by moraines; forested and even inhabited areas could be affected in the future (Fig. 9.7). While such events remain local in significance, warming is expected to increase the number and frequency of such incidents.

Ice avalanches

A special case of slope instability relates to hanging glaciers. The ablation of many glaciers on steep high-altitude mountain slopes is by ice avalanching, and large blocks of ice may sometimes become detached (Alean, 1985; Dutto *et al.*, 1991). Both glacier and permafrost conditions must be considered when analysing the stability of such steep hanging glaciers. At high altitudes and in the shadow of mountain peaks, firn and ice temperatures are usually well below 0ºC. At lower altitudes, however, and on slopes more exposed to the sun, the firn may be at the melting point due to percolating meltwater. In such a situation, only marginal parts of hanging glaciers, especially the vertical front, consist of impermeable, cold ice frozen to the underlying bedrock. This basal temperature pattern probably introduces a thermal anomaly within the underlying permafrost and strong lateral heat flow through the base of the hanging glacier front. The geometry and thermal condition of this ice front, where shear stresses can reach values close to the strength of ice, appears to constitute a key factor controlling the stability of entire ice bodies (Lüthi, 1994). Climate changes may introduce highly complex feedback mechanisms involving surface geometry and firn accumulation, but also temperatures and stress distribution within and at the base of such ice bodies. Long-term monitoring of ice geometries using aerial photography may help to detect unfavourable developments. Independent of attempts to forecast the timing of instability, the run-out distance of potential ice avalanches can be quite realistically assessed (Alean, 1985; Haeberli *et al.*, 1989; Margreth and Funk, 1999). Ice avalanches in connection with earthquake-induced rock avalanches are rare but can be devastating, as in the case of the 1970 mudflow triggered by a rock/ice avalanche at Huascaran, Peru, killing approximately 18,000 people (Körner, 1983; Sidle *et al.*, 1985).

Fig. 9.7. Combined rock/ice avalanche detached from warm permafrost and descending across Brenva Glacier on the southeastern slope of Mt Blanc, 18 January 1998 (c.f. Barla *et al.*, 2000; Deline, 2001).

Ice-covered volcanoes

Lahars and debris flows can originate on active glacier-clad volcanoes. The lateral blast during the eruption of ice-clad Mount St Helens, Washington, on 18 May 1980 destroyed more than 389 km^2 of forest and recreation area, and killed 60 people. The rock avalanche which immediately preceded the eruption, together with intermittent eruptions during the following 3 days, removed nearly 3 km^3 of rocky material from the mountain, including 40 to 45 × 10^6 m^3 of snow and ice (Brugmann and Post, 1981; Lipman and Mullinaux, 1981). Forsyth Glacier, on the northern slope of Mount St Helens, produced two simultaneous ice avalanches that were used to calibrate the timescale of the initial eruption from photographs (Voight, 1981). A series of pyroclastic flows and surges erupting from Nevado del Ruiz, Colombia, on 13 November 1985, mixed with snow and ice on the summit ice cap to trigger catastrophic lahars which killed more than 23,000 people. A total of about 9 × 10^7 m^3 of lahar slurry was transported to depositional areas that extended more than 100 km from the source area (Pierson *et al.*, 1990; Thouret, 1990). Climatic warming has the potential to reduce the ice cover on many active volcanoes and, thereby, decrease corresponding mudflow hazards to forested land.

Glacier floods

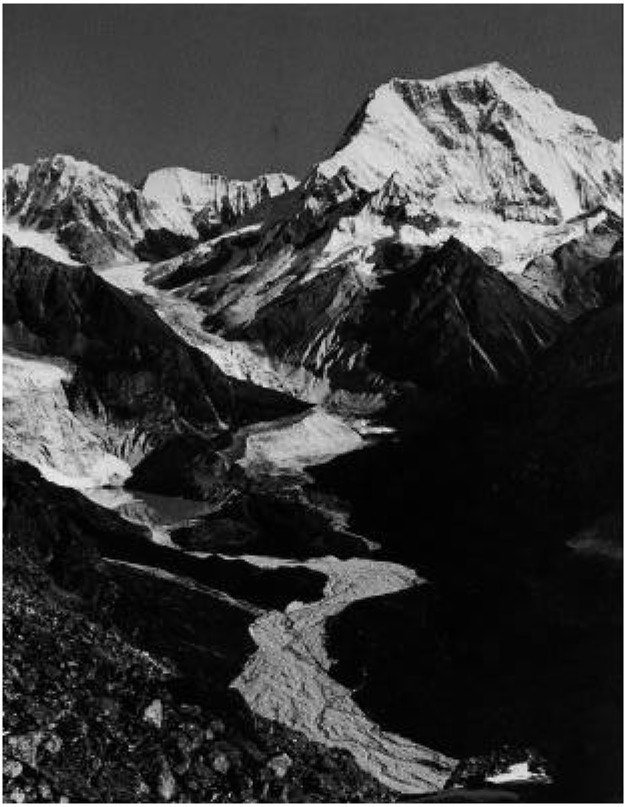

Fig. 9.8. Breach in the moraine dam of a glacier, Bhutan Himalaya. Photograph by A. Gansser, 1969.

The largest floods, of up to several km^3, with peak discharges up to several thousand $m^3 s^{-1}$ and lateral reaches of damage up to several tens of km are caused by floods from outbursts of moraine- and ice-dammed lakes or even entire fjords (Blown and Church, 1985; Vuichard and Zimmermann, 1987; Mayo, 1989). Outbursts of large ice-dammed lakes usually take place through flotation of the ice dam followed by progressive enlargement of subglacial channels (Clague and Mathews, 1973; Spring and Hutter, 1981; Clarke, 1982; Evans, 1986), but smaller events may be due to sudden-break mechanisms causing especially high ratios between peak discharge and outburst volume (Haeberli *et al.*, 1989; Walder and Costa, 1996). Breaching of moraine dams (Fig. 9.8) involves piping (progressive groundwater flow) within the morainic material, liquefied flow/slippage on steep slopes, and overtopping with retrogressive incision (Jackson *et al.*, 1989; Haeberli *et al.*, 1997). Glacier shrinkage induced by atmospheric warming may eliminate problems relating to ice-dammed lakes in some places but create new ones in others. Proglacial lakes, growing behind large terminal moraines built up by heavily debris-covered glaciers, however, constitute a recognizable and rapidly-growing hazard, especially in countries such as Nepal and Bhutan (Gansser, 1983; Yamada, 1998).

Debris flows

As a consequence of rapid snowmelt and/or heavy precipitation, debris flows of highly variable size may form in the loose moraines of glacier forefields and at marginal permafrost sites in the scree of debris cones or rock-glacier fronts (Zimmermann and Haeberli, 1992; Rebetez *et al.*, 1997). Thawing of previously frozen material leads to a loss of cohesion, with a simultaneous increase in pore-water pressure in originally ice-rich material and a reduction in resistance to erosion by running water. The transitional conditions of water-saturated fine material remaining on steeply-inclined, thawing permafrost and large caverns originating from the disappearance of massive ground ice bodies lead to extreme hydraulic heterogeneity in non-consolidated materials and are particularly dangerous. Physical and mathematical modelling with respect to debris flow dynamics (e.g., Davis, 1988) has revealed limitations in our present understanding of such important aspects as the transportation of the largest grains at the surface, the depth of erosion along the flow path, or the sudden halting of flow in the runout zone. Empirical approaches to quantifying the most important trigger conditions and flow parameters are discussed by Rickenmann and Zimmermann (1993) based on a large number of Alpine cases. Maximum volume evacuated per unit channel length, for instance, usually remains below 500 to 700 m^3 and the fact that debris flow trajectories have an overall slope greater than about 10° can be used to estimate maximum runout distances (Clague and Evans, 1994). The corresponding hazard to forested and even inhabited land in mountain areas is likely to increase with atmospheric warming but will remain local in significance.

Landscape evolution, water supply and sediments

All of the previously discussed direct hazards have a maximum reach of the order of a few to less than 100 km and are, therefore, of local to regional significance only. The disappearance of large glaciated areas, on the other hand, has the potential to influence the seasonal runoff variability in large rivers which often feed soil and groundwater in semiarid, irrigated regions such as the Ganzu corridor, China, the Argentinian Pampas, and near the Aral Sea. The relative contribution of glacier meltwater during the warm/dry season may be considerable, if not dominant, and disappearance of such summer water as a consequence of widespread deglaciation in the mountainous source areas may lead to shortages detrimental to vast areas of forested and agricultural land (Fitzharris *et al.,* 1996). Changing high-mountain landscapes with intensified erosion of deglaciated and thawed unconsolidated debris could, at the same time, enhance sediment loads in the remaining waters and reinforce accumulation trends in natural and artificial reservoirs as well as in flat riverbeds. In fact, slow trends in large-scale ecosystem functioning could represent stronger impacts than spectacular individual events of low frequency and high magnitude.

Impacts on High-latitude Lowlands

Fig. 9.9. (A) Headwall of a retrogressive thaw slump at Mayo, Yukon Territory, in late September 1986. Ice lenses in ice-rich permafrost are prominent in the lower part of the headwall. The surface vegetation is overhanging, while the distinct boundary between light and dark sediment marks the maximum depth of the active layer in the Holocene.

Fig. 9.9. (B) The ice-rich zone at base of the active layer, shown protruding out from thawing permafrost, July 1984. The ice-rich zone thaws more slowly, due to its higher latent heat content, and then supports the active layer and vegetation overhanging the thawing headwall. Photographs by C.R. Burn.

Permafrost is defined as ground that remains below 0ºC for 2 or more years (ACGR, 1988) and, where it occurs, it usually remains frozen for centuries or millennia. At the surface of permafrost terrain there is a seasonally-thawed active layer. The base of the active layer is at the permafrost table, where the maximum temperature during the year is 0°C (Muller, 1947; Burn, 1998b). In unconsolidated sediments, including peat and occasionally in bedrock, the uppermost portion of permafrost (immediately beneath the active layer) is ice-rich (Fig. 9.9; Cheng, 1983; Mackay, 1983; French *et al.*, 1986; Shur, 1988).

This ice-rich ground is the geomorphological unit primarily responsible for the sensitivity of permafrost terrain to disturbance (Mackay, 1970). Ice-rich ground has a water content when thawed that is greater than saturation (Fig. 9.10); the volume above saturation is the excess ice content. Ground with excess ice is sensitive to thawing, but if the ice content is less than or at saturation, the ground is thaw stable.

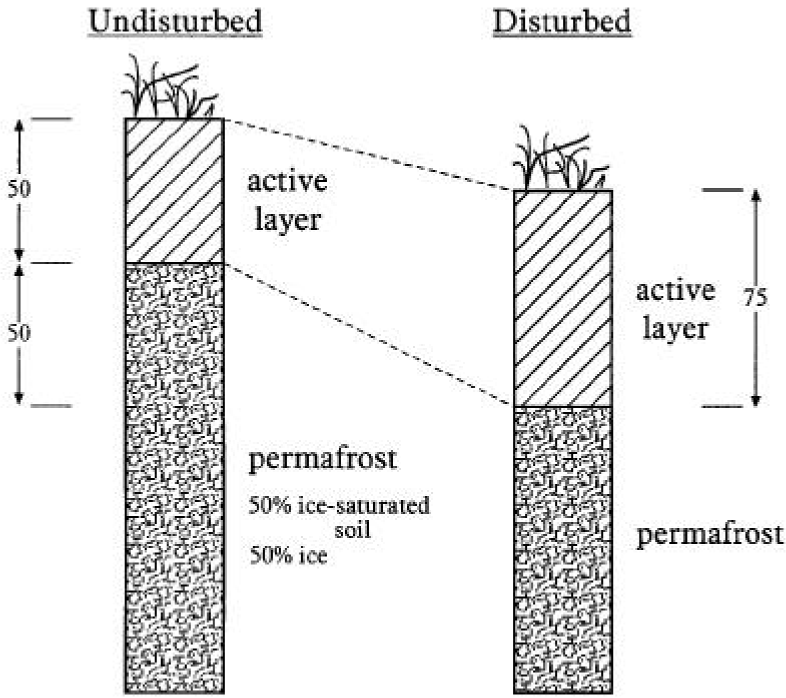

Fig. 9.10. Schematic illustration of ground subsidence with active-layer deepening. In this case, the ice-rich permafrost has an excess ice content of 50%. If the active layer is originally 50 cm thick, then by lowering the base of the active layer a further 50 cm, 25 cm of saturated soil and 25 cm excess ice are melted. Melting of the excess ice leads to 25 cm of surface subsidence (after Mackay, 1970, Fig. 9.3).

The active layer

The thickness of the active layer is a function of the ground surface temperature, the thermal properties of soil materials, and the temperature in permafrost. In general, the active layer is deepest in warm, dry soils or in bedrock, and where the mean annual temperature in permafrost is close to 0°C, while thin active layers are measured in wet peat above cooler ground (Brown, 1970). The active layer generally thickens from north to south through the permafrost regions of the Northern Hemisphere (e.g., Brown, 1970). Beneath the boreal forest of North America, the active layer varies in thickness from about 50 cm to 150 cm (Leverington, 1995; Burn, 1998a); north of the treeline the thickness may decline, but it is rarely less than 30 cm.

Changes in climate that influence active-layer development concern variations in air temperature, snow cover and precipitation. In northern regions, variations in air temperature and precipitation are not independent, most importantly in winter when snowfall is associated with relatively warm conditions. In central Yukon Territory, Canada, warm snowy winters during the 1980s led to permafrost warming by 1° to 1.5°C but subsequently the ground has cooled beneath thin snow covers (Burn, 1998c). A similar trend has been observed in eastern Alaska (Osterkamp and Romanovsky, 1999). On the North Slope of Alaska, active-layer depth increases inland, partly in response to greater trapping of snow by taller vegetation (Romanovsky and Osterkamp, 1995). Indeed, a substantial thinning of the active layer occurs across the treeline associated with the reduction in thickness and increase in density of snow cover at tundra sites, so that cooling of the ground is greater in winter and mean annual ground temperatures are lower (Smith *et al.*, 1998).

Northward migration of forests following summer climate warming may accelerate active-layer deepening through changes to snow-pack properties. This is an example of amplification in climate/permafrost relations due to ecological effects that makes prediction of the impact of climate change on terrain conditions difficult. Field evidence of such amplification has been obtained from examination of active-layer development in western arctic Canada during the early Holocene climate optimum. About 9,000 years ago, when the treeline was considerably further north, and in areas that are now tundra, the active layer was about 2.5 times its present depth (Mackay, 1992). About one half of the active-layer thinning in the last 9,000 years has been attributed to the disappearance of forest from these sites and the remainder to climate cooling (Burn, 1997).

In contrast with the winter regime, warm conditions during summer are usually dry in continental permafrost regions while rain and snowfall are associated with lower temperatures. Evaporation from moist ground also consumes energy and lowers the ground heat flux, further reducing active-layer development. A result is that changes in site wetness may alter ground temperatures and active-layer thickness more than changes in air temperature *per se* (Smith and Riseborough, 1983).

Polar regions are expected to experience greater disturbance than the mid latitudes under enhanced greenhouse warming and changes are expected to be greater in winter than in summer (Roots, 1989). Indeed, records from the last 400 years indicate that the Arctic is warming faster than lower latitudes of the Northern Hemisphere and that the rate of warming was greatest in the 20th century (Overpeck *et al.*, 1997). A wetter climatic regime is projected, with greater snowfall due to a more vigorous general circulation in the atmosphere. For permafrost terrain we should expect a thicker active layer, not primarily because of changes in summer conditions, but cumulatively due to warmer permafrost following deeper snow accumulation.

The most significant terrain hazards wrought by the presence of permafrost are due to melting of the ice-rich zone following deepening of the active layer. Changes in climate may tend to cause relatively gradual deepening of the active layer within the boreal forest because the active layer is relatively thick and climate changes tend to be gradual, superimposed on year-to-year variability. In contrast, surface disturbances, leading to removal of vegetation and organic soil, can cause a substantial, rapid impact on active-layer thickness. The most common and widespread agent of such change is forest fire (Mackay, 1977, 1995; Viereck, 1982). Anthropogenic disturbance, usually from construction or mining, may have a similar effect (Mackay, 1970; Hayley, 1988).

The ice-rich zone at the base of the active layer

The ice-rich zone commonly extends more than a metre below the base of the active layer. Individual ice lenses in this zone are up to several cm thick and ice masses 1 dm thick can occur but are not common. Ice wedges, V-shaped bodies

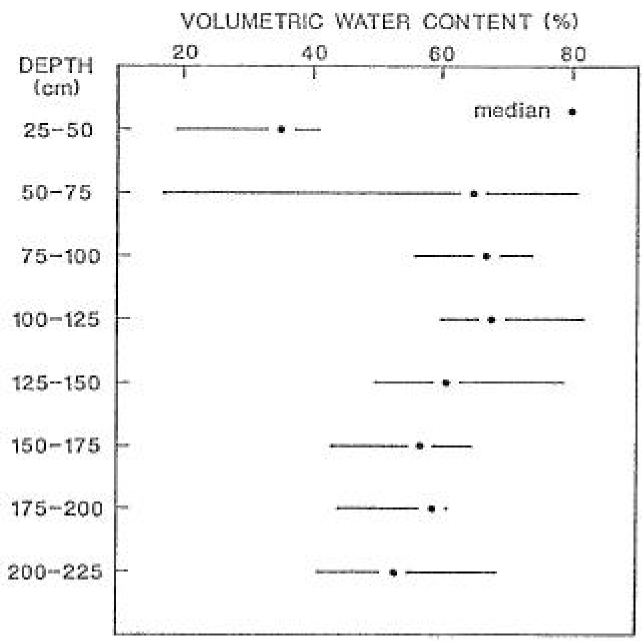

Fig. 9.11. Volumetric water content profile from glaciolacustrine sediments near Mayo, Yukon Territory. The profile shows minimum, median and maximum values of water content in core samples collected from six holes, continuously sampled over 25 cm intervals.

of massive ice, also extend downwards from the base of the active layer (Mackay, 1972; Pollard and French, 1980). Ice wedges are ubiquitous in continuous permafrost, forming polygons 10–15 m in diameter, but they are less evident in discontinuous permafrost, which underlies most of the boreal forest.

A summary profile of the ground-ice content in glaciolacustrine sediments at Mayo, central Yukon Territory, where the dominant vegetation is spruce forest, is presented in Fig. 9.11. The profile indicates that the ice content is characteristically maximal at the top of permafrost. The volumetric water content of these sediments is 40–45% when saturated, so the profile illustrates the high excess ice content in this permafrost.

Data from an experimental disturbance near the sites where the ice contents were derived illustrate ground subsidence following disturbance. Two square plots, 5 m on a side, were established in July 1994, with one left as a control area, while the vegetation and organic mat were removed from the central 9 m^2 of the other. The plots were surveyed over a grid established at 1 m intervals. The survey referenced the ground surface and bottom of the active layer (frost line) in the squares to a nearby benchmark anchored >15 m deep in permafrost. The depth of the active layer was estimated by probing with a pointed steel rod.

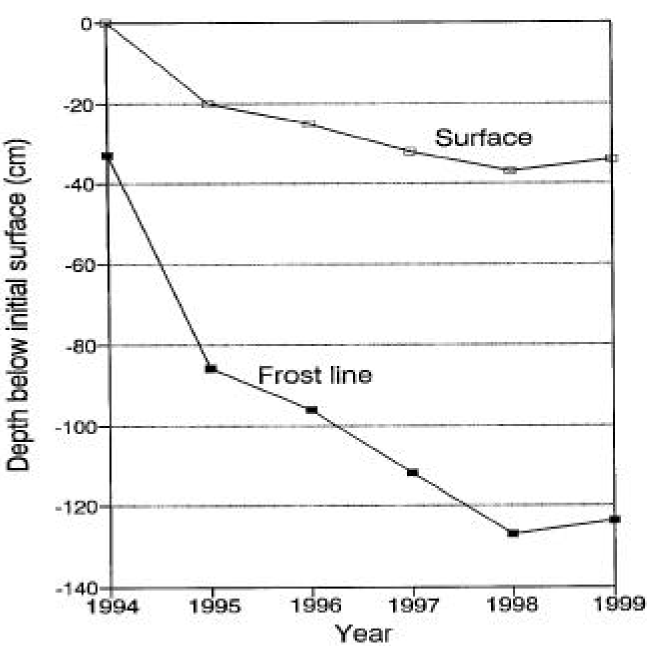

Fig. 9.12. Ground subsidence (1994–1999) at an experimental plot in ice-rich glaciolacustrine sediments near Mayo, Yukon Territory, following surface disturbance. The absolute change in ground surface elevation and the top of frozen ground in late July – early August each year is indicated.

Subsequently the plots have been surveyed each year in late July or August. Fig. 9.12 indicates the lowering of the frost line and the extent of ground subsidence that has occurred from 1994 to 1999 in the disturbed square. At the undisturbed square these variables have shown no trend. The difference between the surface subsidence and the frost line in Fig. 9.12 indicates the depth of the active layer. The rate of subsidence and of active-layer deepening has decreased with time, but the ratio of subsidence to active layer deepening has been about 0.6:1. The precision of the measurements in the irregular microrelief is ±2.5 cm, as determined by repetitive measurements in the undisturbed, control plot. The deeper active layer does not include the thickness of melted ground ice, so the amount of permafrost degradation would have been underestimated if only the active layer had been measured. During five years the ground has subsided by about 35 cm and the active layer has thickened from 33 cm to 90 cm; total permafrost degradation has been 92 cm, with 35 cm from melting of excess ice and 57 cm due to active-layer development.

The original active-layer thickness at the site was 33 cm. For a small thickening of the active layer (e.g., 25%), as may occur during climatic warming, these data indicate ground subsidence of about 5 cm in relatively ice-rich material. In other deposits less ground ice may melt. While these amounts may appear small, such deepening is sufficient to initiate landslides on ice-rich slopes, as in Canada's Mackenzie Valley, following extensive forest fires in summer 1986 (Harry and MacInnes, 1988).

Thermokarst development

Fig. 9.13. Thermokarst lake at Mayo, Yukon Territory, showing trees fallen as a result of lake expansion following thawing of permafrost. Photograph by C.R. Burn, 1988.

The relief produced by melting of ground ice is collectively called thermokarst. Many depressions act as basins for accumulation of water and snow, and once water depth is greater than about one half of the winter ice thickness, permafrost is unsustainable beneath the pool and permafrost degradation proceeds vertically downwards and horizontally (Mackay, 1992). A characteristic of the expansion of thermokarst ponds and lakes in the boreal forest is toppling of trees into the water as permafrost surrounding the lakes thaws (Fig. 9.13). Development of thermokarst lakes requires flat or rolling topography to contain the water-filled depressions.

Fig. 9.14. Aerial photograph of the Mayo permafrost research area, Yukon Territory, showing thermokarst lakes and two retrogressive thaw slumps developed in glaciolacustrine sediments. Part of aerial photograph A27482-84 © 1989 Her Majesty the Queen in Right of Canada, reproduced from the collection of the National Air Photo Library with permission of Natural Resources Canada.

During the Holocene, periods of warm climate have been associated with increased thermokarst development throughout the permafrost regions of Russia (Czudek and Demek, 1970), Alaska (Carson, 1968) and Canada (Harry and MacInnes, 1988; Rampton, 1988). However, individual lakes may be initiated by a site-specific disturbance, such as the melting of an ice wedge, or uprooting of a tree. Changes in surface conditions following forest fire may also lead to ground thawing over a sufficient area for several lakes to develop simultaneously. Although the association of thermokarst development with climate change over millennia is well established, there is no evidence that climatic amelioration since the “Little Ice Age” has led to such development throughout the discontinuous permafrost zones (Fig. 9.6A). Detailed studies at Mayo, Yukon Territory, indicate that site-specific causes are responsible for the development of a cluster of lakes during the last 600 years (Fig. 9.14; Burn and Smith, 1990).

However, at the southerly margin of permafrost in the Northern Hemisphere, there is considerable evidence that permafrost which grew during the “Little Ice Age” is currently melting, as the limits for sustainability of such frozen ground move northwards (Thie, 1974; Halsey *et al.*, 1995). Across the northern Prairie Provinces of Canada recent degradation of permafrost in peatlands is well documented. Such thawing leads to elimination of forests that developed on peat plateaus raised by ice above the surrounding wetland (Vitt *et al.*, 1994). While an association with recent climate warming is proposed for many areas, Zoltai (1993) demonstrated that such changes may also be repeated in a cyclical fashion due to the recurrence of fire at 500-year intervals.

Forest fire

Forest fires cause the most widespread disturbance to surface conditions in permafrost regions. The degree of disturbance is determined by the intensity and nature of the wildfire, which are associated with antecedent meterological conditions. Projections of the nature and frequency of forest fires in a warmer world suggest the extent of damage may increase (Flannigan and Van Wagner 1991). However, certain environments are less susceptible to disturbance by fire, particularly where there is sufficient moisture to reduce the intensity of burning at the surface (Swanson, 1996).

Fires initially raise ground surface temperatures in summer due to reduction in shading (Rouse, 1976). Destruction of vegetation leads to reduced evapo-transpiration, raising ground temperatures further as the ground surface dries out. Deepening of the active layer usually follows, so that by the end of summer the ice-rich zone has begun to degrade. On hillslopes, thawing leads to landslides, as the ice-rich zone provides a lubricated plane with low shear strength (Harry and MacInnes, 1988). Such disturbances may also follow particularly warm summers when there is substantial rainfall in early autumn (Mackay and Mathews, 1973), but are almost ubiquitous in sloping ground after fire (Viereck, 1982).

Active-layer deepening is accentuated in years following fire by an increased snow depth in burned areas because interception of snow by the forest canopy is eliminated or reduced. In combination, such changes at the ground surface lead to an increase in ground surface temperature of >2°C and may lead to permafrost eradication. Where permafrost is thin, the degradation may be completed within a decade, but where the thickness is >10 m, centuries are required to thaw the ice-rich ground completely (Burn, 1998a). Vegetation recovery is assured over such timescales, so that permafrost degradation will likely stop. While the most substantial damage occurs in ice-rich terrain, the slowest recovery is at dry sites. Mackay (1977, 1995) carefully documented the active-layer thickening (1968–1993) and recovery following a fire in 1968 at Inuvik, Northwest Territories.

The "drunken forest"

The boreal forest is comprised of a mix of species, but is dominated by white and black spruce (*Picea glauca, P. mariana*), and also contains larch (*Larix laricina*), pine (*Pinus contorta*, *P. banksiana*), birch (*Betula papyrifera*), poplar and aspen

Fig. 9.15. (A) Deciduous forest of poplar trees (*Populus balsamifera*) growing in permafrost-free ground on a point bar near Mayo, Yukon Territory, July 1991.

Fig. 9.15. (B) A stand of "drunken forest" 9 km north of Mayo, Yukon Territory, July 1994. The trees are black spruce (*Picea mariana*). The ground cover is of lichens and feather mosses. Photographs by C.R. Burn, 1994.

(*Populus balsimifera, P. tremuloides*). Inclement conditions during the growing season lead to relatively small trees, rarely >20 m tall, and often <10 m. Where the active layer is relatively deep, usually in dry or permafrost-free ground, trees are usually erect (Fig. 9.15A), but in moist, ice-rich terrain, trees are typically tilted in various directions (Fig. 9.15B). The subsequent woodland is known as a "drunken forest" and is dominated by spruce trees.

Zoltai (1975a) suggested that growth of trees in the boreal forest occurs in cycles punctuated by wildfire. Trees that grow shortly after fire are erect, but as the active layer thins with vegetation succession, tilting occurs. Zoltai (1975a) found that tilted trees were prevalent in stands >100 years old. Growth of ground ice at the base of the active layer may be responsible for the tilting (Mackay, 1983). Separately, Zoltai (1975b) examined the reaction wood in tilted trees to investigate specific conditions causing the tilting. From two locations in the Mackenzie Valley, northwest Canada, he obtained evidence suggesting that tilting is associated with increased autumn temperatures and precipitation, presumably melting the ice-rich zone. This suggestion implies that tilting may become more prevalent following climate change.

Projections of the impact of climate change

The most comprehensive data on the response of permafrost to recent climate change are from Alaska, where ground temperature records are available from a number of sites (Osterkamp and Romanovsky, 1999). Analysis of these data suggests that snow cover plays a critical role on the ground thermal regime because, although air temperatures have not increased in the 1990s, average snow depth has increased and ground temperatures have warmed over large areas. However, in eastern Alaska and in adjacent Yukon Territory, Canada, the inverse has occurred, as snow depths have declined (Burn, 1998c). At present we have no published long-term records of active-layer thickness in the English literature and the analysis of this variable has remained dominantly theoretical. An international effort, the Circumpolar Active Layer Monitoring Program (CALM), has been mounted to obtain such data series (Brown, 1997b).

Nevertheless, Anisimov and Nelson (1996) and Anisimov *et al.* (1997) have taken projections of future climates and attempted to infer the changes they may induce. The equilibrium permafrost distribution in a 2×CO_2 world projected by these simulations shows permafrost disappearing from much of its present distribution and the continuous/discontinuous boundary moving several hundred km northward. The time required to reach equilibrium for such transformation is on the order of several millennia.

More useful data have been obtained from geothermal simulations of specific sites, which provide the flexibility to model local ground conditions (Osterkamp and Romanovsky, 1996). These are computationally-intensive, but an important result is that permafrost degradation can often be modelled with few variables (Burn, 1998a). Of these variables, the most critical is the increase in permafrost degradation with the square root of time. This relation implies that for sites with permafrost thickness >10 m degradation will normally take more than a century.

The effects of climatic variability on permafrost stability have also been investigated in the context of climate change. Riseborough and Smith (1993) imposed a variable climate, with similar statistical properties to that of the last 30 years, on the ground thermal regime in warm, thin permafrost characteristic of the boreal forest near Fort Simpson, Northwest Territories. They ran ground temperature simulations for centuries with both stable and warming climate. The work showed that permafrost may become established rapidly due to frost penetration in winter, but the rate of summer thawing is less due to insulation provided by the thawed active layer. In consequence, permafrost has a persistence that is due to the latent heat held in ground ice and the seasonally-changing thermal properties of the surficial layers.

To conclude discussion of permafrost conditions, we reiterate that permafrost is a persistent phenomenon and its future presence is assured within the boreal forest. The terrain hazards presented by permafrost relate to the presence of ground ice. Given that near-surface layers of permafrost terrain are characteristically ice-rich, ground subsidence and slope instability are likely the principal hazards to be expected under climatic change.

Conclusions and Perspectives

Atmospheric warming is predicted to be greater in polar regions than at lower latitudes and more pronounced at high altitudes than in lowlands (Fitzharris *et al.*, 1996; Beniston, 2000). In polar regions, air and ground warming may lead to a more northerly extension of the boreal forest as growing seasons lengthen and become warmer. Increases in snowfall will benefit forest growth by maintaining warmer soil temperatures and increasing soil moisture early in the growing season. Near-surface permafrost degradation is expected to accompany such an evolution in environmental conditions, so that substrate stability is not assured. In some cases slope movement may be catastrophic, but in most instances the settlement may be slow, and the water released by melting ground ice will evaporate. The potentially increasing rate of climatic change, when combined with the slow speed of adjustment in ground-ice conditions and of forest migration over extended distances, is likely to induce long-lasting disequilibria.

Subpolar permafrost regions do not support major commercial forest activities. The forests are used for firewood and rough lumber, but construction-grade materials are produced elsewhere. This is unlikely to change because long-term ground instability, relatively cold soil temperatures, depletion of nutrients in the active layer, and restriction of root systems to the active layer all limit tree growth. Extensive forest fires, usually initiated by lightning after a week or two of hot weather, also deplete timber stocks. There have been suggestions that wildfires may become more common following climate warming. In combination, these effects will probably preclude large-scale forestry operations in permafrost regions, but local enterprises may flourish at sites in the discontinuous zone where there are extensive tracts of unfrozen ground.

In mountain areas, meltwater runoff from glaciated and perennially frozen areas only represents a small contribution to the annual water supply but can have strong influence on streamflow in adjacent lowlands during the warm/dry season. The disappearance of perennial ice above and below the earth surface influences the seasonality of discharge by reducing meltwater production in the warm season and by increasing the permeability of frozen/thawing materials. The latter effect may have strong impacts on soil humidity and growth conditions for forest and tundra in such dry areas as Tian Shan Mountains or Tibet Plateau. Revegetation of terrain following deglaciation is slow under conditions of high-mountain climates and, therefore, deglaciated morainic deposits are exposed to erosion for decades to centuries. In general, accelerated future warming would cause an enlargement of the periglacial belt in high mountain areas, an upslope shifting of hazard processes, and a widespread reduction in the stability of formerly glaciated or perennially-frozen slopes.

Besides densely populated high-mountain areas such as the Alps or parts of the Andes, the described changes would predominantly affect remote areas and have little impact on human settlements. A widespread problem may emerge, however, if the empirical basis for hazard forecasting and protection, i.e., historical documents, statistics of measured time series or traces in nature of past events with long recurrence intervals, lost its applicability as snow and ice conditions evolved beyond the range of Holocene variability. Retreat of glaciers and corresponding changes in landscape and scenery could represent the most directly visible and most easily understood signals of global warming.

In the case of accelerated future warming, the cryospheric components of high mountain environments would likely change at high rates and lead to pronounced disequilibria in the water cycle, in mass-wasting processes and sediment flux, as well as in growth conditions for vegetation. For those directly involved with such changes, the main challenge would be to adapt to the high rates of environment evolution. Empirical knowledge would have to be replaced more and more by improved process understanding, especially concerning runoff formation and slope stability.

Acknowledgements

Scientific activities of WH on glaciers, permafrost and climate change are funded by the University of Zurich, the Swiss National Science Foundation, the EU-PACE project, the Federation of Astronomical and Geophysical Data Analysis Services of ICSU and the International Hydrological Programme of UNESCO. Research by CRB in northwest Canada is supported by the Natural Sciences and Engineering Research Council of Canada; the Polar Continental Shelf Project, Natural Resources Canada; the Inuvik Research Centre of the Aurora Research Institute, Inuvik, Northwest Territories; and the Northern Research Institute, Yukon College, Whitehorse, Yukon Territory. The cartography is by Christine Earl. Polar Continental Shelf Project Contribution #02600.

References

ACGR (Associate Committee on Geotechnical Research) (1988) Glossary of permafrost and related ground ice terms. *National Research Council of Canada, Associate Committee on Geotechnical Research, Technical Memorandum*, 142.

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

Anisimov, O.A. and Nelson, F.E. (1996) Permafrost distribution in the northern hemisphere under scenarios of climatic change. *Global and Planetary Change* 14, 59-72.

Anisimov, O.A., Shiklomanov, N.I. and Nelson, F.E. (1997) Global warming and active-layer thickness: results from transient general circulation models. *Global and Planetary Change* 15, 61-77.

Barla, G., Dutto, F. and Mortara, G. (2000) Brenva Glacier rock avalanche of 18 January 1997 on the Mount Blanc range, northwest Italy. *Landslide News* 13, 2-5.
Barsch, D. (1977) Nature and importance of mass-wasting by rock glaciers in Alpine permafrost environments. *Earth Surface Processes* 2, 231-245.
Beniston, M. (2000) Environmental change in mountains and uplands. *Key Issues in Environmental Change*. Edward Arnold, London, and Oxford University Press, Oxford.
Berthling, I., Etzelmüller, B., Iken, T. and Sollid, J.L. (1998) The rock glaciers on Prins Karls Forland, Svalbard. I: Internal structure, flow velocity and morphology. *Permafrost and Periglacial Processes* 9, 135-145.
Blair, R.W. (1994) Moraine and valley wall collapse due to rapid deglaciation in Mount Cook National Park, New Zealand. *Mountain Research and Development* 14, 347-358.
Blown, A. and Church, M. (1985) Catastrophic lake drainage within the Homathko River basin, British Columbia. *Canadian Geotechnical Journal* 22, 551-563.
Brown, J. (1997a) Disturbance and recovery of permafrost terrain. In: Crawford, R.M.M. (ed.), *Disturbance and Recovery of Arctic Lands.* Kluwer Academic Publishers, Dordrecht, The Netherlands, pp. 167-178.
Brown, J. (1997b) Circumpolar active layer monitoring (CALM) network – November 1997. *Frozen Ground* 21, 22-23.
Brown, J., Ferrians, O.J.Jr, Heginbottom, J.A. and Melnikov, E.S. (1997) Circum-arctic map of permafrost and ground ice-conditions. U.S. Geological Survey, *Circum-Pacific Map Series*, CP-45.
Brown, R.J.E. (1970) *Permafrost in Canada: its influence on northern development.* University of Toronto Press, Toronto.
Brugmann, M.M. and Post, A. (1981) Effects of volcanism on the glaciers of Mount St. Helens. *US Geological Survey, Circular 850-D*.
Budd, W.F., Coutts, B. and Warner, R.C. (1998) Modelling the Antarctic and Northern Hemisphere ice-sheet changes with global climate through the glacial cycle. *Annals of Glaciology* 27, 153-160.
Burga, C.A. (1999) Vegetation development on the glacier forefield Morteratsch (Switzerland). *Applied Vegetation Science* 2, 17-24.
Burn, C.R. (1997) Cryostratigraphy, paleogeography, and climate change during the early Holocene warm interval, western Arctic coast, Canada. *Canadian Journal of Earth Sciences* 34, 912-925.
Burn, C.R. (1998a) The response (1958-1997) of permafrost and near-surface ground temperatures to forest fire, Takhini River valley, southern Yukon Territory. *Canadian Journal of Earth Sciences* 35, 184-199.
Burn, C.R. (1998b) The active layer: two contrasting definitions. *Permafrost and Periglacial Processes* 9, 411-416.
Burn, C.R. (1998c) Field investigations of permafrost and climatic change in northwest North America. In: Lewkowicz, A.G. and Allard, M. (eds), Proceedings of 7th International Conference on Permafrost, Yellowknife, N.W.T., 23-26 June 1998. Quebec City, Quebec. *Collection Nordicana* 57, 107-120.
Burn, C.R. and Smith, M.W. (1990) Development of thermokarst lakes during the Holocene at sites near Mayo, Yukon Territory. *Permafrost and Periglacial Processes* 1, 161-176.
Carson, C.E. (1968) Radiocarbon dating of lacustrine strands in arctic Alaska. *Arctic* 21, 12-26.
Cheng, G. (1983) The mechanism of repeated-segregation for the formation of thick-layered ground ice. *Cold Regions Science and Technology* 8, 57-66.

Cihlar, J., Barry, T.G., Ortega Gil, E., Haeberli, W., Kuma, K., Landwehr, J.M., Norse, D., Running, S., Scholes, R., Solomon, A.M. and Zhao, S. (1997) *GCOS/GTOS plan for terrestrial climate-related observation.* GCOS 32, version 2.0, WMO/ TD-796, UNEP/DEIA/TR, 97-7.

Clague, J.J. and Evans, S.G. (1994) Formation and failure of natural dams in the Canadian Cordillera. Geological Survey of Canada, *Bulletin 464*.

Clague, J.J. and Mathews, W.H. (1973) The magnitude of jökulhlaups. *Journal of Glaciology* 12, 501-504.

Clarke, G.K.C. (1982) Glacier outburst floods from 'Hazard Lake', Yukon Territory, and the problem of flood magnitude prediction. *Journal of Glaciology* 28, 3-21.

Czudek, T. and Demek, J. (1970) Thermokarst in Siberia and its influence on the development of lowland relief. *Quaternary Research* 1, 103-120.

Davis, T.R.H. (1988) Debris flow surges: a laboratory investigation. *Mitteilungen 96 der VAW/ ETHZ*, 122.

Deline, P. (2001). Recent Brenva rock avalanches (Valley of Aosta): new chapter in an old story? *Geografia Fisica e Dinamica Quaternaria*.

Deming, D., Sass, J.H., Lachenbruch, A.H. and De Rito, R.F. (1992) Heat flow and subsurface temperature as evidence for basin-scale ground-water flow, North Slope of Alaska. *Geological Society of America Bulletin* 104, 528-542.

Dramis, F., Govi, M., Guglielmin, M. and Mortara, G. (1995) Mountain permafrost and slope instability in the Italian Alps. The Val Pola landslide. *Permafrost and Periglacial Processes* 6, 73-81.

Dutto, F., Godone, F. and Mortara, G. (1991) L'écroulement du glacier supérieur de Coolidge (Paroi nord du Mont Viso, Alpes occidentales). *Revue de Géographie Alpine* 2, 7-18.

Espizua, L.E. and Bengochea, J.D. (1990) Surge of Grande del Nevado Glacier (Mendoza, Argentina) in 1984: its evolution through satellite images. *Geografiska Annaler* 72A, 255-259.

Evans, S.G. (1986) The maximum discharge of outburst floods caused by the breaching of man-made and natural dams. *Canadian Geotechnical Journal* 23, 385-387.

Evans, S.G. and Clague, J.J. (1988) Catastrophic rock avalanches in glacial enviroments. Landslides. In: *Proceedings of the Fifth International Symposium on Landslides, 10-15 July 1988, Lausanne.* Vol. 2, pp. 1153-1158.

Evans, S.G. and Clague, J.J. (1990) Reconnaissance observations on the Tim Williams Glacier rock avalanche, near Stewart, British Columbia. *Geological Survey of Canada Paper 90-1E*, 351-354.

Evans, S.G. and Clague, J.J. (1993) Glacier-related hazards and climatic change. *In* The World at Risk: Natural Hazards and Climatic Change. In: Bras, R. (ed.), *American Institute of Physics, Conference Proceedings* 277, 48-60.

Fitzharris, B.B., Allison, I., Braithwaite, R.J., Brown, J., Foehn, P.M.B., Haeberli, W., Higuchi, K., Kotlyakov, V.M., Prowse, T.D., Rinaldi, C.A., Wadhams, P., Woo, M.-K., Xie Youyou, Anisimov, O., Aristarain, A., Assel, R.A., Barry, R.G., Brown, R.D., Dramis, F., Hastenrath, S., Lewkowicz, A.G., Malagnino, E.C., Neale, S., Nelson, F.E., Robinson, D.A., Skvarca, P., Taylor, A.E. and Weidick, A. (1996) The cryosphere: changes and their impacts. Climate Change 1995. Impacts, adaptations and mitigation of climate change: scientific-technical analyses. *Contribution of Working Group II to the Second Assessment Report of the Intergovernmental Panel on Climate Change*. Cambridge University Press, Cambridge, UK, pp. 241-265.

Flannigan, M.D. and Van Wagner, C.E. (1991) Climate change and wild fire in Canada. *Canadian Journal of Forest Research* 21, 66-72.

Foriero, A., Ladanyi, B., Dallimore, S.R., Egginton, P.A. and Nixon, F.M. (1998) Modeling of deep seated hill slope creep in permafrost. *Canadian Geotechnical Journal* 35, 560-578.

French, H.M. (1996) *The periglacial environment,* 2nd edn. Longman, London.

French, H.M., Bennett, L. and Hayley, D.W. (1986) Ground ice conditions near Rea Point and on Sabine Peninsula, eastern Melville Island. *Canadian Journal of Earth Sciences* 23, 1389-1400.

Gansser, A. (1983) Geology of the Bhutan Himalaya. *Denkschriften der Schweizerischen Naturforschenden Gesellschaft 96*, Birkhäuser Verlag, Basel-Boston-Stuttgart.

Goodwin, C.W., Brown, J. and Outcalt, S.I. (1984) Potential responses of permafrost to climatic warming. In: McBeath, J.H. (ed.), *The potential effects of carbon dioxide-induced climatic changes in Alaska*. University of Alaska, Fairbanks, pp. 92-105.

Grove, J.M. (1987) Glacier fluctuations and hazards. *The Geographical Journal* 153, 351-369.

Haeberli, W. (1996) Glacier fluctuations and climate change detection. *Geografia Fisica e Dinamica Quaternaria* 18, 191-199.

Haeberli, W. and Hoelzle, M. (1995) Application of inventory data for estimating characteristics of and regional climate-change effects on mountain glaciers: a pilot study with the European Alps. *Annals of Glaciology* 21, 206-212.

Haeberli, W., Alean, J-C., Müller, P. and Funk, M. (1989) Assessing risks from glacier hazards in high mountain regions: some experiences in the Swiss Alps. *Annals of Glaciology* 13, 96-102.

Haeberli, W., Wegmann, M. and Vonder Muehll, D. (1997) Slope stability problems related to glacier shrinkage and permafrost degradation in the Alps. *Eclogae Geologicae Helvetiae* 90, 407-414.

Haeberli, W., Barsch, D., Brown, J., Guodong, C., Corte, A.E., Dramis, F., Evin, M., Gorbunov, A.P., Guglielmin, M., Harris, C., Harris, S.A., Hoelzle, M., Kääb, A., King, L., Lieb, G.K., Matsuoka, N., Ødegaard, R.S., Shroder, J.F., Sollid, J.L., Trombotto, D., Von der Mühll, D., Shaoling, W. and Xiufeng, Z. (1998) Mapping, modelling and monitoring of mountain permafrost: A review of ongoing programs. *Programme, Abstracts, and Working Group Reports, Seventh International Conference on Permafrost, 23-26 June 1998, Yellowknife, N.W.T.* pp. 233-246.

Haeberli, W., Frauenfelder, R., Hoelzle, M. and Maisch, M. (1999a) On rates and acceleration trends of global glacier mass changes. *Geografiska Annaler* 81A, 585-591.

Haeberli, W., Kääb, A., Wagner, S., Vonder Mühll, D., Geissler, P., Haas, J.N., Glatzel-Mattheier, H. and Wagenbach, D. (1999b) Pollen analysis and ^{14}C age of moss remains in a permafrost core recovered from the active rock glacier Murtèl-Corvatsch, Swiss Alps: geomorphological and glaciological implications. *Journal of Glaciology* 45, 1-8.

Halsey, L.A., Vitt, D.H. and Zoltai, S.C. (1995) Disequilibrium response of permafrost in boreal continental western Canada to climate change. *Climatic Change* 30, 57-73.

Harry, D.G. and MacInnes, K.L. (1988) The effect of forest fires on permafrost terrain stability, Little Chicago-Travaillant lake area, Mackenzie Valley, N.W.T. *Current Research, Part D. Geological Survey of Canada, Paper* 88-1D, pp. 91-94.

Hayley, D.W. (1988) Maintenance of a railway grade over permafrost in Canada. In: *Proceedings of 6th International Conference on Permafrost, 2-5 August 1988, Trondheim, Norway*. Tapir, Trondheim, 3, 43-48.

Hoelzle, M., Wagner, S., Kääb, A. and Vonder Mühll, D. (1998) Surface movement and internal deformation of ice-rock mixtures within rock glaciers at Pontresina-Schafberg, Upper Engadin, Switzerland. In: Proceedings of 7th International Conference on Permafrost, Yellowknife, N.W.T., 23-26 June 1998. Quebec City, Quebec. *Collection Nordicana 57*, 465-471.

IAHS(ICSI)/UNEP/UNESCO (1989) World glacier inventory - status 1988. Haeberli, W., Bösch, H., Scherler, K, Østrem, G. and Wallén, C.C. (eds), Nairobi.

IAHS(ICSI)/UNEP/UNESCO (1998) Fluctuations of Glaciers 1990-1995. Haeberli, W., Hoelzle, M., Suter, S. and Frauenfelder, R. (eds), Zurich.

Isaksen, K., Vonder Muehll, D., Gubler, H.U., Kohl, Th. and Sollid, J.L. (2000) Ground surface temperature reconstruction based on data from a deep borehole in permafrost at Janssonhaugen, Svalbard. *Annals of Glaciology* 31, 287-294.

Jackson, L.E., Hungr, O., Gardner, J.S. and Mackay, C. (1989) Cathedral mountain debris flows, Canada. *Bulletin of the International Association for Engineering Geology* 40, 35-54.

Jin, H., Qui, G., Zhao, L., Wang, S. and Zeng, Z. (1993) Thermal regime of alpine permafrost in the upper reach of Urumqi river, Xinjiang, China. In: *Proceedings of 6th International Conference on Permafrost, 5-9 July 1993, Beijing, China.* Vol. 1, pp. 307-311.

Johannesson, T., Raymond, C.F. and Waddington, E.D. (1989) Time-scale for adjustment of glaciers to changes in mass balance. *Journal of Glaciology* 35, 355-369.

Kääb, A., Gudmundsson, G.H. and Hoelzle, M. (1998) Surface deformation of creeping mountain permafrost. Photogrammetric investigations on Murtèl Rock Glacier, Swiss Alps. In: Proceedings of7th International Conference on Permafrost, Yellowknife, N.W.T., 23-26 June 1998. Quebec City, Quebec. *Collection Nordicana 57*, 531-537.

Körner, H.J. (1983) Zur Mechanik der Bergsturzströme vom Huascaran, Peru. *Hochgebirgsforschung 6*, Universitätsverlag Wagner, Innsbruck, 71-110.

Kotlyakov, V.M. and Krenke, A.N. (1982) Investigations of hydrological conditions of alpine regions by glaciological methods. *IAHS Publication 138*, 31-42.

Lachenbruch, A.H. and Marshall, B.V. (1986) Changing climate: geothermal evidence from permafrost in the Alaskan arctic. *Science* 234, 689-696.

Letréguilly, A. and Reynaud, L. (1990) Space and time distribution of glacier mass balance in the northern hemisphere. *Arctic and Alpine Research* 22, 43-50.

Leverington, D. (1995) A field survey of late-summer depths to frozen ground at two study areas near Mayo, Yukon Territory, Canada. *Permafrost and Periglacial Processes* 6, 373-379.

Lipman, P.W. and Mullinaux, D.R. (1981) The 1980 eruptions of Mount St. Helens, Washington. *US Geological Survey Professional Paper 1250*.

Lüthi, M. (1994) Stabilität steiler Gletscher: Eine Studie über den Einfluss möglicher Klimaänderungen; Untersuchungen am Beispiel eines Hängegletschers in der Westflanke des Eigers. Diploma thesis, ETH Zurich.

Mackay, J.R. (1970) Disturbances to the tundra and forest tundra environment of the western Arctic. *Canadian Geotechnical Journal* 7, 420-432.

Mackay, J.R. (1972) The world of underground ice. *Annals, Association of American Geographers* 62, 1-22.

Mackay, J.R. (1977) Changes in the active layer from 1968 to 1976 as a result of the Inuvik fire. *Report of Activities, part B. Geological Survey of Canada, Paper 77-1B*, 273-275.

Mackay, J.R. (1983) Downward water movement into frozen ground, western Arctic coast, Canada. *Canadian Journal of Earth Sciences* 20, 120-134.

Mackay, J.R. (1992) Lake stability in an ice-rich permafrost environment: examples from the western arctic coast. In: Robarts, R.D. and Bothwell, M.L. (eds), *Aquatic Ecosystems in Semi-arid Regions: Implications for Resource Management*.

Environment Canada, Saskatoon, Sask. National Hydrology Research Institute Symposium Series, 7, 1-26.

Mackay, J.R. (1995) Active layer changes (1968 to 1993) following the forest-tundra fire near Inuvik, N.W.T., Canada. *Arctic and Alpine Research* 27, 323-336.

Mackay, J.R. and Mathews, W.H. (1973) Geomorphology and Quaternary history of the Mackenzie River valley near Fort Good Hope, Northwest Territories, Canada. *Canadian Journal of Earth Sciences* 10, 26-41.

Major, J.J. and Newhall, C.G. (1989) Snow and ice perturbation during historical volcanic eruptions and the formation of lahars and floods. *Bulletin of Volcanology* 52, 1-27.

Margreth, S. and Funk, M. (1999) Hazard mapping for ice and combined snow/ice avalanches - two case studies from the Swiss and Italian Alps. *Cold Regions Science and Technology* 30, 159-173.

Matsuoka, N., Hirakawa, K., Watanabe, T., Haeberli, W. and Keller, F. (1998) The role of diurnal, annual and millennial freeze-thaw cycles in controlling alpine slope instability. In: Proceedings of 7th International Conference on Permafrost, Yellowknife, N.W.T., 23-26 June 1998. Quebec City, Quebec. *Collection Nordicana 57*, 711-717.

Mayo, L. (1989) Advance of Hubbard Glacier and 1986 outburst of Russel Fjord, Alaska, U.S.A. *Annals of Glaciology* 13, 189-194.

McSaveney, M.J. (1993) Rock avalanches of 2 May and 16 September 1992, Mount Fletcher, New Zealand. *Landslide News* 7.

McSaveney, M.J., Chinn, T.J. and Hancox, G.T. (1992) Mount Cook rock avalanche of 14 December 1991. *Landslide News* 6.

Meier, M.F. and Bahr, D.B. (1996) Counting Glaciers: Use of Scaling Methods to Estimate the Number and Size Distribution of the Glaciers on the World. In: Colbeck, S.C. (ed.), *Glaciers, Ice Sheets and Volcanoes: a Tribute to Mark F. Meier*. CRREL Special Report 96-27, US Army, Hanover, New Hampshire, pp. 1-120.

Muller, S.W. (1947) *Permafrost or permanently frozen ground and related engineering problems*. Edwards, Ann Arbor, Michigan.

O'Connor, J.E. and Costa, J.E. (1993) Geologic and hydrologic hazards in glacierized basins in North America resulting from 19th and 20th century global warming. *Natural Hazards* 8, 121-140.

Ødegard, R.S. and Sollid, L.S. (1993) Coastal cliff temperatures related to the potential for cryogenic weathering processes, western Spitsbergen, Svalbard. *Polar Research* 12, 95-106.

Oerlemans, J. (1994) Quantifying global warming from the retreat of glaciers. *Science* 264, 243-245.

Osterkamp, T.E. and Gosink, J.P. (1991) Variations in permafrost thickness in response to changes in paleoclimate. *Journal of Geophysical Research* 96, 4423-4434.

Osterkamp, T.E. and Romanovsky, V.E. (1996) Characteristics of changing permafrost temperatures in the Alaskan Arctic. *Arctic and Alpine Research* 28, 267-273.

Osterkamp, T.E. and Romanovsky, V.E. (1999) Evidence for warming and thawing of discontinuous permafrost in Alaska. *Permafrost and Periglacial Processes* 10, 17-37.

Overpeck, J., Hughen, K., Hardy, D., Bradley, R., Case, R., Douglas, M., Finney, B., Gajewski, K., Jacoby, G., Jennings, A., Lamoureux, S., Lasca, A., MacDonald, G., Moore, J., Retelle, M., Smith, S., Wolfe, A. and Zielinski, G. (1997) Arctic environmental change of the last four centuries. *Science* 278, 1251-1256.

Pierson, T.C., Janda, R.J., Thouret, J.-C. and Borrero, C.A. (1990) Perturbation and melting of snow and ice by the 13 November 1985 eruption of Nevado del Ruiz, Colombia, and consequent mobilization, flow and deposition of lahars. *Journal of Volcanology and Geothermal Research* 41, 17-66.

Pollard, W.H. and French, H.M. (1980) A first approximation of the volume of ground ice, Richards Island, Pleistocene Mackenzie Delta, Northwest Territories, Canada. *Canadian Geotechnical Journal* 17, 509-516.

Rampton, V.N. (1988) Quaternary geology of the Tuktoyaktuk Coastlands, Northwest Territories. *Geological Survey of Canada, Memoir 423*.

Rebetez, M., Lugon, R. and Baeriswyl, P.-A. (1997) Climatic warming and debris flows in high mountain regions: the case study of the Ritigraben, Switzerland. *Climate Change* 36, 371-389.

Richardson, S.D. and Reynolds, J.M. (2000) An overview of glacial hazards in the Himalayas. *Quaternary International* 65/66, 31-47.

Rickenmann, D. and Zimmermann, M. (1993) The 1987 debris flows in Switzerland: documentation and analysis. *Geomorphology* 8, 175-189.

Riseborough, D.W. and Smith, M.W. (1993) Modelling permafrost response to climate change and climate variability. In: *Proceedings of 4th International Symposium on Thermal Engineering and Science for Cold Regions, Hanover, New Hampshire*. US Army Cold Regions Research and Engineering Laboratory Special Report, 93-22: 179-187.

Romanovsky, V.E. and Osterkamp, T.E. (1995) Interannual variations of the thermal regime of the active layer and near-surface permafrost in northern Alaska. *Permafrost and Periglacial Processes* 6, 313-335.

Roots, E.F. (1989) Climate change: high latitude regions. *Climatic Change* 15, 223-253.

Rouse, W.R. (1976) Microclimatic changes accompanying burning in subarctic lichen woodland. *Arctic and Alpine Research* 8, 357-376.

Shumskii, P.A. (1964) *Principles of Structural Glaciology*. Transl. D. Kraus. Dorer, New York.

Shur, Yu.L. (1988) The upper horizon of permafrost soils. In: *Proceedings of 6th International Conference on Permafrost, 2-5 August 1988. Trondheim, Norway*. Vol. 1, pp. 867-871.

Sidle, R.C., Pearce, A.J. and O'Loughlin, C.L. (1985) *Hillslope stability and land use*. Water Resources Monograph, no. 11, American Geophysical Union, Washington, DC.

Smith, C.A.S., Burn, C.R., Tarnocai, C. and Sproule, B. (1998) Air and soil temperature relations along an ecological transect through the permafrost zones of northwestern Canada. In: Proceedings of 7th International Conference on Permafrost, Yellowknife, N.W.T., 23-26 June 1998. Quebec City, Quebec. *Collection Nordicana* 57, 1009-1015.

Smith, M.W. and Riseborough, D.W. (1983) Permafrost sensitivity to climatic change. In: *Proceedings of 4th International Conference on Permafrost, 17-22 July 1983, Fairbanks, Alaska*. National Academy Press, Washington, DC., 1, pp. 1178-1183.

Spring, U. and Hutter, K. (1981) Numerical studies of jökulhlaups. *Cold Regions Science and Technology* 4, 227-244.

Swanson, D.K. (1996) Susceptibility of permafrost soils to deep thaw after forest fires in interior Alaska, U.S.A. and some ecologic implications. *Arctic and Alpine Research* 28, 217-227.

Taylor, A.E. (1991) Holocene paleoenvironmental reconstruction from deep ground temperatures: a comparison with paleoclimate derived from the $\delta^{18}O$ record in an ice core from the Aggassiz Ice Cap, Canadian Arctic Archipelago. *Journal of Glaciology* 37, 209-219.

Thie, J. (1974) Distribution and thawing of permafrost in the southern part of the discontinuous permafrost zone in Manitoba. *Arctic* 27, 189-200.

Thouret, J.-C. (1990) Effects of the November 13, 1985 eruption on the snow pack and ice cap of Nevado del Ruiz volcano, Colombia. *Journal of Volcanology and Geothermal Research* 41**,** 177-201.

Tufnell, L. (1984) *Glazier Hazards*. Longman, London and New York.

Viereck, L.A. (1982) Effects of fire and firelines on active layer thickness and soil temperatures in interior Alaska. In: *Proceedings, 4th Canadian Permafrost Conference, Calgary, Alta, 2-6 March 1980.* National Research Council of Canada, Ottawa, Ontario, pp. 123-135.

Vitt, D.H., Halsey, L.A. and Zoltai, S.C. (1994) The bog landforms of continental western Canada in relation to climate and permafrost patterns. *Arctic and Alpine Research* 26, 1-13.

Voight, B. (1981) The 1980 eruptions of Mount St. Helens, Washington. Time scale for the first moments of the May 18 eruption. *US Geological Survey Professional Paper 1250*, 69-86.

Vonder Mühll, D., Stucki, T. and Haeberli, W. (1998) Borehole temperatures in alpine permafrost: a ten year series. In: Proceedings of 7th International Conference on Permafrost, 23-26 June 1998,Yellowknife, N.W.T. Quebec City, Quebec. *Collection Nordicana* 57, 1089-1095.

Vuichard, D. and Zimmermann, M. (1987) The 1985 catastrophic drainage of a moraine-dammed lake Khumbu Himal, Nepal: cause and consequences. *Mountain Research and Develeopment* **7,** 91-110.

Walder, J.S. and Costa, J.E. (1996) Outburst floods from glacier-dammed lakes: the effect of mode of laid drainage on flood magnitude. *Earth Surface Processes and landforms* 21, 701-723.

Warrick, R.A., Le Provost, C., Meier, M.F., Oerlemans, J., Woodworth, P.L., Alley, R.B., Bindschadler, R.A., Bentley, C.R., Braithwaite, R.J., de Wolde, J.R., Douglas, B.C., Dyurgerov, M., Flemming, N.C., Genthon, C., Gornitz, V., Gregory, J., Haeberli, W., Huybrechts, P., Jóhannesson, T., Mikolajewicz, U., Raper, S.C.B., Sahagian, D.L., van de Wal, R.S.W. and Wigley, T.M.L. (1996) Changes in sea level. Climate Change 1995: The Science of Climate Change. *Contribution of Working Group I to the Second Assessment Report of the Intergovernmental Panel on Climate Change*. Cambridge University Press, UK, pp. 359-405.

Wegmann, M., Gudmundsson, G.H. and Haeberli, W. (1998) Permafrost Changes in Rock Walls and the Retreat of Alpine Glaciers: a Thermal Modelling Approach. *Permafrost and Periglacial Processes* 9, 23-33.

Weller, G., Chapin, F.S., Everett, K.R., Hobbie, J.E., Kane, D., Oechel, W.C., Ping, C.L., Reeburgh, W.S., Walker, D. and Walsh, J. (1995) The Arctic flux study: a regional view of trace gas release. *Journal of Biogeography* 22, 365-374.

Yamada, T. (1998) *Glacier lake and its outburst flood in the Nepal Himalaya*. Data Center for Glacier Research, Japanese Society of Snow and Ice, Monograph 1, Japan.

Zimmermann, M. and Haeberli, W. (1992) Climatic change and debris flow activity in high mountain areas; a case study in the Swiss Alps. *Catena Supplement* 22, 59-72.

Zoltai, S.C. (1975a) Structure of subarctic forests on hummocky permafrost terrain in northwest Canada. *Canadian Journal of Forest Research* 5, 1-9.

Zoltai, S.C. (1975b) Tree ring record of soil movements on permafrost. *Arctic and Alpine Research* 7, 331-340.

Zoltai, S.C. (1993) Cyclic development of permafrost in the peatlands of northwestern Alberta, Canada. *Arctic and Alpine Research* 25, 240-246.