GEOLOGICAL SURVEY OF CANADA
BULLETIN 524

GEOMORPHOLOGICAL PROCESSES IN THE ALPINE AREAS OF CANADA

J.M. Ryder

1998

Natural Resources Canada
Ressources naturelles Canada

Canada

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GEOMORPHOLOGICAL PROCESSES IN THE ALPINE AREAS OF CANADA:

the effects of climate change and their impacts on human activities

J.M. Ryder

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Catalogue No. M42-524E
ISBN 0-660-17489-8

Available in Canada from
Geological Survey of Canada offices:

601 Booth Street
Ottawa, Ontario K1A 0E8

3303-33rd Street N.W.
Calgary, Alberta T2L 2A7

101-605 Robson Street
Vancouver, B.C. V6B 5J3

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in selected public libraries across Canada

Price subject to change without notice

Cover illustration

Flood Glacier in the Boundary Ranges of the Coast Mountains, British Columbia. The low forested ridge in the foreground that is cut by a single river channel is the Little Ice Age moraine of Flood Glacier. The former large terminal lobe of this glacier has melted due to climate amelioration since the late nineteenth century. Floor Lake, which is impounded by the glacier in an ice-free tributary valley, lies behind the steep forested ridge at left centre. It emits periodic outburst floods that drain to Stikine River (directly below the photographer) via subglacial conduits and the outwash channels that are visible in the lower part of the photo. Photograph by J. Ryder. GSC 1998-032

Critical reviewer
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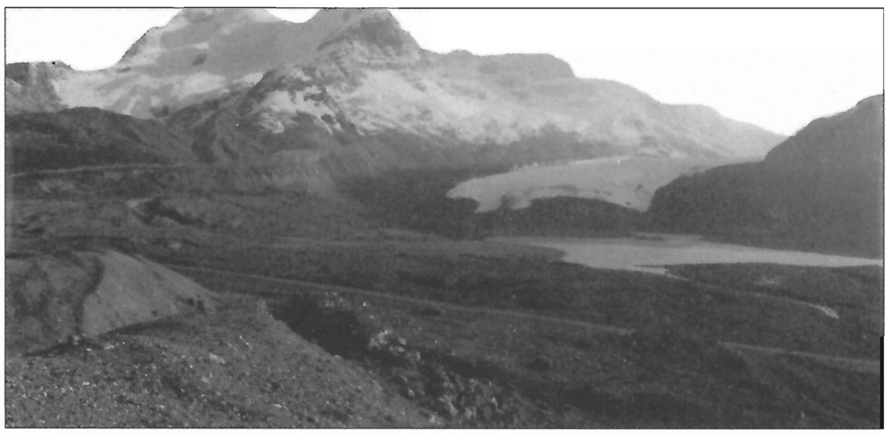
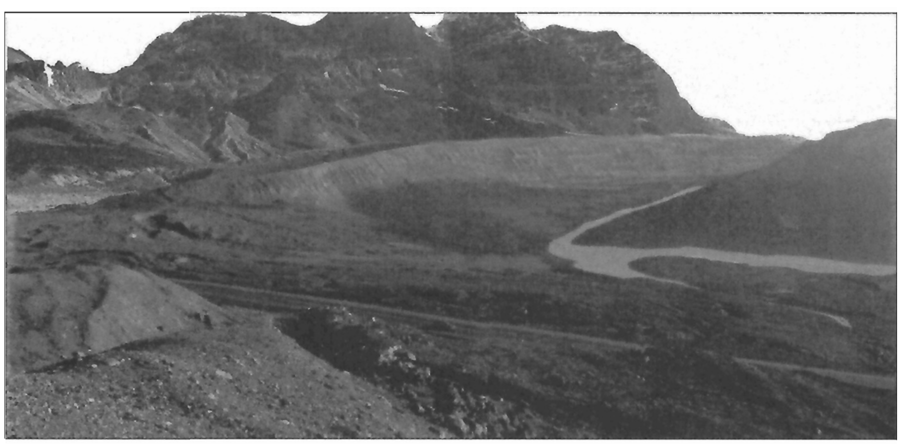
Original manuscript submitted: 1994-05-25
Final version approved for publication: 1995-09-22





Frontispiece: *Athabasca Glacier: (1917, 1986, and the future?). Climate change could result in significant retreat of large glaciers such as this and related reduction in attractiveness to tourists. (Photos: 1917: Alberta/B.C. Provincial Boundary Commission; 1986: B.H. Luckman; image of the Athabasca Glacier in the future if the glacier continues to recede.)*

FOREWORD

Change is a welcomed or feared challenge. It is welcomed when the outcome is understood and expected to be good; it is feared when the outcome is unknown or expected to be bad. Global climate change is the most significant projected change currently facing humanity. Global climate change is a feared change because so many unknowns are involved. What will be the rate and level of climate change? How will global climate change be "shared" or impact on various regions? How will the complex earth ecosystems be affected? Most importantly, how will humanity cope?

One element of the global change picture that falls within the mandate of the Geological Survey of Canada is surface geological processes. Surface geological processes include the various forces which act to change the Earth's surface. The ones with which we are most familiar are water and wind. These act on the surface by eroding (removing) materials from one place and depositing them in another. Climate plays a major role in driving these processes and changes in climate will result in changes in their nature and intensity.

Through time we have gained knowledge which lets us predict what to expect from surface geological processes and hence to mitigate or avoid their harmful effects. For example, application of process knowledge has resulted in development of local practices such as planting shelter belts and strip farming to control soil erosion on individual fields and in government initiatives such as removing land from cultivation to offset damage on a regional scale. If future global climate changes result in changes in natural processes, then adaptive measures will have to be developed. Developing and implementing new coping strategies requires research, planning, and time, hence the sooner we can gather information on what to expect the better prepared we will be to take action as the changes occur.

The Geological Survey of Canada is preparing a series of overview reports that look at several of the more widespread kinds of geological changes occurring in Canada. Each of these reports will show areal distribution of processes today and areas where these processes are most sensitive to climate change, and each looks at how different factors control these processes, discusses the sensitivity of the processes to climate changes, and considers the impact of the effect of processes on human activity. These are not intended as research documents predicting what might be expected in each part of the country, but as warnings to draw attention to potential "hotspots" or

AVANT-PROPOS

Le changement est, selon le cas, bienvenu ou craint. Bienvenu, si les conséquences sont connues et qu'on les prévoit favorables; craint, si les conséquences sont inconnues ou présumées nuisibles. Le changement climatique à l'échelle du globe est le plus important parmi ceux auxquels devra faire face l'humanité. Il est craint parce qu'il soulève de nombreuses questions sans réponse. À quelle vitesse se produira-t-il et quelle sera son importance? Comment ses conséquences se répartiront-elles dans les différentes régions du monde? Comment les écosystèmes complexes seront-ils touchés? Et, surtout, comment l'humanité réagira-t-elle à ses effets?

L'étude de l'un des éléments du cadre général du changement climatique à l'échelle du globe relève du mandat de la Commission géologique du Canada. Il s'agit de l'étude des processus géologiques superficiels ou, en d'autres mots, des multiples forces qui modifient la surface de la Terre, parmi lesquelles les plus connues sont l'eau et le vent. Ces agents érodent les matériaux, les transportent et les déposent ailleurs. Le climat est une des forces motrices de ces processus. Voilà pourquoi le changement climatique se traduira par des modifications dans leur nature et de leur intensité.

Avec le temps, nous avons acquis suffisamment de connaissances pour en arriver à bien connaître les processus géologiques superficiels et, par conséquent, à en atténuer ou en éviter les effets néfastes. Ainsi, l'utilisation de ces connaissances a permis d'élaborer des méthodes d'intervention locale, comme celles d'ériger des plantations brise-vent et de faire de la culture en bandes pour contrer l'érosion du sol de certains champs, mais aussi de mettre en oeuvre des initiatives gouvernementales comme celle de soustraire à la culture certaines terres pour compenser les dommages causés à l'échelle régionale. Si le changement climatique planétaire devait avoir des répercussions sur les processus naturels, il faudrait alors élaborer des mesures correctrices. La conception et la mise en oeuvre de nouvelles stratégies de correction demande de la recherche, de la planification et du temps. Moins nous tarderons à recueillir des données sur ce qui risque de se produire, mieux nous serons préparés à réagir aux changements lorsqu'ils surviendront.

La Commission géologique du Canada prépare une série de rapports de synthèse sur plusieurs des processus géologiques les plus répandus au Canada. Dans chacun de ces rapports, on présente comment se répartit actuellement l'action de ces processus sur le territoire canadien, les régions où le changement climatique influe le plus sur eux, les facteurs qui les régissent, leur vulnérabilité ainsi que leurs répercussions sur l'activité humaine. Ces rapports ne se veulent pas des documents de recherche dans lesquels seraient présentées les prévisions des conséquences du changement climatique à l'échelle du globe dans toutes les parties du pays, mais des avertissements pour attirer l'attention sur les zones névralgiques où les effets de ce grand changement sur les processus seront les plus areas where the processes in question are likely to be most affected by global climate change. The hope is that these studies will foster and focus follow-up research that will determine potential impact more precisely and provide information for planning mitigative measures.

Other reports published in this series are: 'Sensitivity of eolian processes to climate change in Canada' (GSC Bulletin 421) and 'Sensitivity of the Canadian coast to sea-level rise' (GSC Bulletin 505).

R.J. Fulton,
Co-ordinator,
Impact of Global Climate Change on Geological Processes Project

intenses. Il est à espérer que ces études seront le catalyseur de nouveaux projets de recherche et permettront de mieux cerner leur sujet, pour en arriver à déterminer avec plus de précision les modifications possibles et à constituer une source d'information dans la planification de mesures correctrices.

D'autres documents ont déjà été publiés dans cette série. Ce sont *Sensitivity of eolian processes to climate change in Canada* (bulletin 421 de la CGC) et *Sensitivity of the Canadian coast to sea-level rise* (bulletin 505 de la CGC).

R.J. Fulton
Coordonnateur
Projet d'étude des répercussions du changement climatique à l'échelle du globe sur les processus géologiques

Preface

This is the third overview report in the Geological Survey of Canada project: "Impact of Global Climate Change on Geological Processes". These overviews provide the information on the possible impacts of global climate change on geological processes that permit us to see the potential magnitude of the problem.

This bulletin looks at the relationship between climate and geological processes that occurs in alpine areas. It not only describes how these processes affect human activities, it also discusses how predicted global climate change could result in modification of alpine processes, and provides a map classification of alpine areas of Canada that is based on climate, relief, and glacier cover. This document is the first step in the identification of the problems which could result in Canada's alpine areas because of global climate change. It highlights the critical regions where studies must now be conducted to outline the probable local impact of global change and to develop the best coping mechanisms.

This report was prepared under the auspices of the Global Change Program of the Geological Survey of Canada with the partial support of the Canada Green Plan Fund.

M.D. Everell
Assistant Deputy Minister
Earth Sciences Sector

Préface

Voici le troisième document de synthèse produit par la Commission géologique du Canada dans le cadre de son Projet d'étude des répercussions du changement climatique à l'échelle du globe sur les processus géologiques. Ces documents contiennent des informations sur les répercussions possibles du changement climatique à l'échelle du globe sur les processus géologiques et permettent d'avoir un aperçu des proportions que pourraient prendre le problème.

Le présent bulletin se penche sur les liens qui existent entre le climat et les processus géologiques dans les régions alpines. Il contient non seulement une description de la façon dont ces processus influent sur l'activité humaine, mais fait également état de la façon dont le changement climatique planétaire pourrait modifier les processus actifs dans les régions alpines. De plus, il est accompagné d'une carte de classification des régions alpines du Canada en fonction du climat, du relief et de la présence ou de l'absence de glaciers. Ce document constitue la première étape d'une démarche conduisant à l'identification des problèmes qui risquent de survenir dans ces régions du pays à cause du changement climatique à l'échelle planétaire. Il met l'accent sur les zones névralgiques où des études doivent dès maintenant être entreprises afin de prévoir les répercussions locales du changement climatique et d'élaborer les meilleurs mécanismes pour y faire face.

Ce bulletin a été préparé dans le cadre du Programme des changements à l'échelle du globe de la Commission géologique du Canada, avec l'appui partiel du Fonds du Plan vert du Canada.

M.D. Everell
Sous-ministre adjoint
Secteur des sciences de la Terre

CONTENTS

GEOMORPHOLOGICAL PROCESSES IN THE ALPINE AREAS OF CANADA: THE EFFECTS OF CLIMATE CHANGE AND THEIR IMPACTS ON HUMAN ACTIVITIES

Abstract

Geomorphological processes are related to climate in a complex manner. Estimates of how processes will react to anticipated climate changes, such as warming and increased precipitation, are necessarily tentative and qualitative. They indicate the kinds of changes that may occur in the alpine landscapes, but not the magnitude or precise rates of such changes.

If climate change proceeds as forecast, then one of the most apparent results will be a rise in the timberline. This will result in a decrease in the extent of the alpine zone, thus altering the appearance of many mountain areas and their potential for recreation and tourism.

Anticipated changes in snowfall include reduced snow and increased winter rain at elevations close to the present winter snowline, and increased snowfall on higher and colder mountains. This will have a detrimental effect on ski resorts at low elevations, but improve winter access for other activities. Warmer summer temperatures will probably lead to a rise in the summer snowline and significant loss of glaciers, changing the scenery in southern mountain areas, such as Banff and Jasper national parks.

Climate change will probably have some hazardous effects. Melting of permafrost and more intense rainfall may initiate or increase the frequency of slides and debris flows in some areas. Heavier snowfall may increase avalanche danger, and glacier recession will probably result in more floods and debris flows due to the failure of glacier and moraine dams. Increased sediment input to glacier-fed rivers may lead to increased channel instability, erosion, and flooding. The hazard zones related to most of these fluvial processes will extend a long way beyond the limits of the alpine zone.

Thinner winter snowpacks, rapid spring runoff, and a reduced extent of permanent snow and ice will reduce both the seasonal and longer term water storage capacity of alpine areas. This effect could lead to serious water shortages in regions such as the Okanagan Valley, British Columbia, where summer meltwater maintains extensive agriculture and industry.

The impact of climate change on permafrost in alpine areas is not likely to be severe, except possibly in small areas. Warming of ground temperatures will cause a slow reduction in the extent and thickness of permafrost. This will have noticeable effects only where soils contain a high proportion of ice, conditions that are probably not extensive in most alpine areas.

Résumé

Les processus géomorphologiques ont des liens complexes avec le climat. L'établissement de la façon dont les processus réagiront aux changements climatiques attendus, comme le réchauffement du climat et l'augmentation des précipitations, ne peut être qu'une démarche provisoire et qualitative. On peut déterminer quels types de changements se produiront dans les régions alpines, mais pas leur importance ni leur vitesse précise.

Si le changement climatique se passe comme prévu, l'un des résultats les plus apparents sera le déplacement vers le haut de la limite forestière. Il s'ensuivra une diminution du territoire de la zone alpine et, par conséquent, un changement dans l'aspect de nombreuses régions montagneuses et une modification de leur potentiel comme zone récréative et touristique.

Parmi les changements présumés dans les précipitations, mentionnons une réduction des chutes de neige et un accroissement des chutes de pluie aux altitudes proches de la limite actuelle des neiges d'hiver, mais aussi une augmentation des chutes de neige sur les montagnes plus élevées où il fait plus froid. Il en résultera des effets néfastes sur les centres de ski de basse altitude mais des avantages pour d'autres activités hivernales. L'élévation des températures estivales se traduira probablement par un déplacement vers le haut de la limite des neiges d'été et une fonte significative des glaciers, d'où une transformation du paysage des zones montagneuses méridionales comme celles des parcs nationaux de Banff et de Jasper.

Le changement climatique aura fort probablement des effets qui seront la source de risques. La fonte du pergélisol et les pluies plus abondantes pourraient déclencher des glissements de terrain et des coulées de débris dans certaines régions, ou augmenter leur fréquence. L'augmentation des chutes de neige pourrait accroître les risques d'avalanche. Quant au recul des glaciers, il causera probablement des crues et des coulées de débris plus nombreuses découlant de la rupture de barrages glaciaires et morainiques. Le plus grand apport de sédiments vers les rivières fluvioglaciaires pourrait intensifier l'instabilité des chenaux, leur érosion et les crues. Les zones dangereuses associées à la plupart de ces processus s'étendront bien au-delà des limites de la zone alpine.

L'amincissement du manteau nival en hiver, l'accélération du ruissellement printanier et la réduction de l'étendue de la neige et de la glace permanentes se traduiront par une diminution de la capacité de stockage saisonnier et, à plus long terme, de l'eau dans les régions alpines. Les régions comme la vallée de l'Okanagan, en Colombie-Britannique, où les eaux de fonte estivales alimentent les vastes terres agricoles et les exploitations industrielles, pourraient connaître de graves pénuries d'eau.

Le changement climatique ne devrait pas avoir de répercussions importantes sur le pergélisol des régions alpines, sauf peut-être par endroits. Le réchauffement des températures du sol réduira lentement l'étendue et l'épaisseur du pergélisol. Ce changement aura des effets visibles là où le sol contient une grande proportion de glace, ce qui est probablement plutôt rare dans la plupart des régions alpines.

SUMMARY

Geomorphological processes in alpine areas range from dramatic events such as debris flows and snow avalanches to the imperceptible downslope movement of soil and the slow flow of glaciers. In one way or another, all of these processes affect human endeavours in the alpine zone, and they are all, in some manner, linked to climate. Increases in atmospheric greenhouse gases are predicted to lead to significant climate changes, resulting in related changes in the nature, frequency, or rates of operation of geomorphological processes. For example, avalanches and debris flows may occur in areas not affected before, and there may be more catastrophic floods in valleys carrying glacier meltwater.

The objective of this report is to provide information about geomorphological processes in Canadian alpine areas, especially processes that impinge on human activities, and to estimate how they may be affected by predicted climate change. This study is not intended as a focused prediction of what can be expected, but as a starting point for defining problems of dealing with global change in a rather sensitive environment and deciding where future work should be directed.

For the purpose of this study, alpine areas are defined as regions lying above the upper limit of continuous forest: *timberline* is the boundary of the alpine areas shown in Figure 1. The alpine zone includes subalpine parkland, alpine tundra, glaciers and snowfields, and barren rocky ground. Mountainous areas in the Arctic are excluded because, unlike high-elevation areas farther south, they are not distinguished from adjacent lowlands by distinctive geomorphological processes or their sensitivity to climate change.

SOMMAIRE

Les processus géomorphologiques dans les régions alpines vont des plus catastrophiques telles les coulées de débris et les avalanches, aux plus imperceptibles mouvements vers le bas d'une pente et à l'avancée très lente des glaciers. D'une façon ou de l'autre, tous ces processus se répercutent sur les entreprises humaines ayant leur siège dans la zone alpine et ils sont, d'une certaine manière, liés au climat. Il est prévu que l'augmentation de la quantité de gaz à effet de serre dans l'atmosphère modifiera significativement le climat, ce qui influera sur la nature, la fréquence ou la vitesse des processus géomorphologiques. Par exemple, des avalanches et des coulées de débris pourraient se produire là où il n'y en a jamais eu, tout comme des crues plus catastrophiques pourraient être déclenchées dans les vallées empruntées par les eaux de fonte des glaciers.

Le présent document contient des informations sur les processus géomorphologiques qui agissent dans les régions alpines du Canada, en particulier sur ceux qui ont des répercussions sur les activités humaines. Il vise à évaluer comment le changement climatique prévu modifiera ces processus. Il ne se limite pas à des prévisions de ce qui arrivera; il se veut aussi un point de départ pour cerner les problèmes découlant du changement planétaire dans un milieu plutôt fragile et de décider des axes de recherche pour les travaux futurs.

Pour les besoins de la présente étude, les régions alpines sont définies comme celles qui se trouvent au-delà de la limite supérieure de la forêt continue : la *limite forestière* est la limite des régions alpines montrées à la figure 1. La zone alpine inclut le parc subalpin, la toundra alpine, les glaciers et les champs de neige, ainsi que les étendues de roche nue. Les régions montagneuses de l'Arctique sont exclues du fait que, contrairement aux régions de haute altitude plus au sud, elles ne se distinguent pas des basses terres adjacentes par des différences dans les processus géomorphologiques ou dans leur vulnérabilité au changement climatique.

Estimates of the effects of future climate change on alpine geomorphological processes are tentative for several reasons: the anticipated climate modification has been defined only in very general terms; the exact nature of climate controls on most geomorphological processes is not well understood; and the relations between climate and geomorphological processes are complex. The broad climatic predictions upon which this paper is based indicate: that warming in high northern latitudes will exceed the predicted global mean increase of 2.5°C in winter but be closer to the average in summer, that continental areas will warm more than maritime regions, that precipitation and evaporation will increase throughout the year in high latitudes and in winter in mid-latitudes, and that net effects will result in a decrease in the area of seasonal snow cover.

Les estimations des effets du changement climatique futur sur les processus géomorphologiques en zone alpine sont provisoires et ce, pour plusieurs raisons : pour le moment, la modification prévue du climat n'a été définie qu'en termes très généraux; la façon dont le climat agit sur la plupart des processus géomorphologiques n'est pas bien connue; et les liens entre le climat et les processus géomorphologiques sont complexes. Les grandes prédictions climatiques sur lesquelles se base le présent document pointent vers les faits suivants : que le réchauffement du climat aux latitudes nordiques dépassera la hausse moyenne prévue de 2,5 °C en hiver, mais qu'il sera plus proche des valeurs moyennes en été; que les zones continentales se réchaufferont davantage que les régions maritimes; que les précipitations et l'évaporation augmenteront pendant toute l'année aux hautes latitudes, mais seulement pendant l'hiver aux moyennes latitudes, et que les effets nets se traduiront par une diminution de la superficie de la couverture nivale saisonnière.

Geomorphological processes are dynamic actions or events that occur at the Earth's surface due to the application of forces generated by atmospheric processes, such as rainfall and temperature variations, and terrestrial effects, such as gravity (expressed as slope steepness) and seismic shaking. Thus the processes under review include erosion and deposition by flowing water and ice, rockfalls and soil creep influenced by freezing and thawing, landslides triggered by torrential rain or melting permafrost, and avalanches following heavy snowfall. A major difficulty in determining the response of a process to climate change is that many of the significant climatic variables, such as precipitation intensity during major storms and the annual number of freeze-thaw cycles, are neither routinely measured at most climate stations nor referred to in climate predictions. Consequently, only tentative and qualitative suggestions about the likely effects of climate change can be provided.

Les processus géomorphologiques sont dynamiques du fait qu'ils se produisent à la surface de la Terre et résultent de l'application, d'une part, de forces créées par des processus atmosphériques (comme les précipitations et les variations de la température) et, d'autre part, de forces terrestres (comme la gravité, exprimée par l'inclinaison de la pente, et les secousses sismiques). Aussi les processus à l'étude incluent-ils l'érosion et le dépôt de matériaux par l'écoulement de l'eau et de la glace, les éboulements et les reptations de sol dues à l'action du gel et du dégel, les glissements de terrain déclenchés par les pluies torrentielles ou la fusion du pergélisol et les avalanches découlant d'abondantes chutes de neige. Une des principales difficultés dans la détermination des effets du changement climatique sur un processus réside dans le fait que de nombreuses variables significatives, comme par exemple l'intensité des précipitations durant les grosses tempêtes et le nombre de cycles de gel-dégel au cours d'une année, ne sont pas habituellement mesurées à la plupart des stations climatiques ni ne sont mentionnées dans les prévisions climatiques. Par conséquent, les suppositions sur les effets possibles du changement climatique ne peuvent être que provisoires et qualitatives.

Timberline, the lower limit of the alpine zone, is a dynamic boundary. Its elevation is controlled primarily by summer temperatures and the duration of winter snowpack. Climate warming will probably result in increasingly favourable conditions for trees at high elevations, and thus timberline will rise, although other changes, especially increasing snowfall, could have opposing effects. A rise in timberline will result in a reduction in the extent of alpine areas, especially alpine meadows, and the alpine zone will disappear altogether on mountains and plateaus which rise only a short distance above the present limit of trees.

La limite forestière, marquant le passage à la zone alpine, est dynamique. L'altitude à laquelle elle se trouve dépend principalement des températures estivales et du temps que reste le manteau nival d'hiver. Le réchauffement climatique devrait probablement faire en sorte qu'à de plus fortes altitudes, il existe des conditions qui se prêtent de plus en plus à l'implantation d'arbres, déplaçant ainsi la limite forestière vers le haut; cependant, d'autres changements, comme la plus grande abondance des chutes de neige, pourraient avoir des effets contraires. Un tel déplacement de la limite forestière réduirait l'étendue des régions alpines, en particulier celle des alpages; ainsi, la zone alpine pourrait disparaître tant sur les montagnes que sur les plateaux dont l'altitude excède à peine celle de la limite forestière actuelle.

The extent of permafrost in alpine areas has not been mapped in detail, although it is well known that frozen ground is most widespread in more northerly regions and in continental (inland) mountains, where winters are relatively cold and the winter snowpack is thin. Modification of permafrost due to climate warming is likely to take place slowly over many centuries because the ground is well insulated from the atmosphere by soil and vegetation. Effects are likely to be

L'étendue du pergélisol dans les régions montagneuses n'a pas été cartographiée en détail, mais il est bien établi que le sol gelé est plus abondant dans les régions les plus septentrionales et dans les secteurs de montagnes continentales (intérieures), là où les hivers sont relativement froids et où le manteau nival est mince. La modification du pergélisol à la suite d'un réchauffement climatique devrait progresser à un rythme lent et s'échelonner sur de nombreux siècles du fait que les parties gelées sont bien isolées de l'atmosphère par le sol et la végétation. Les effets seront complex, because soil characteristics, especially moisture content, and vegetation will also be modified. Climate warming will be most effective where permafrost is relatively warm and thin (i.e. discontinuous permafrost regions in Fig. 1) resulting in thickening of the seasonally thawed soil (*active layer*), and the development of a zone of unfrozen ground between the active layer and the top of the permafrost. Thin permafrost may disappear. The results of ground warming will be noticeable in areas underlain by ice-rich permafrost, which is probably not extensive in many alpine areas. Where ice lenses are present, melting will result in subsidence of the land surface, collapse of peat plateaus in bogs, slope instability, and mudflows. Such effects, once initiated, may persist at specific sites for several decades or longer. Other typical problems that might be expected are an increase in road ditch maintenance and increasing pressure on building foundations as the rates of soil creep increase due to warming permafrost and a thickening active layer. In general, however, direct human actions, such as removal of insulating soil and vegetation, are potentially of greater significance for localized permafrost degradation than the effects of climate change.

vraisemblablement complexes puisque les caractéristiques pédologiques (en particulier la teneur en humidité) et la végétation seront également modifiées. Les répercussions du réchauffement climatique seront maximales dans les zones de pergélisol relativement chaud et mince (soit celles de pergélisol discontinu de la figure 1), ce qui se traduira par un épaississement du sol dégelé selon les saisons (mollisol) et la formation d'une zone de sol non gelé entre le mollisol et le sommet du pergélisol. Le pergélisol en minces couches pourrait disparaître. Les effets du réchauffement du sol s'observeront là où le pergélisol contient beaucoup de glace, ce qui probablement plutôt rare dans bien des régions alpines. Aux endroits où des lentilles de glace sont présentes, la fonte pourrait faire en sorte que la surface du sol s'affaisse, que des plateaux palsiques s'effondrent, que des pentes deviennent instables et que des coulées de boue soient déclenchées. Ces phénomènes, une fois amorcés, pourraient persister à des sites spécifiques pendant plusieurs décennies, voire même sur une plus longue période. Parmi les autres problèmes caractéristiques qui pourraient survenir, mentionnons qu'il se peut que, à mesure que la reptation du sol s'intensifie en réponse à un réchauffement du pergélisol et à un épaississement du mollisol, l'entretien des fossés demande plus de travail et les fondations des bâtiments soient soumises à de plus fortes pressions. En général, cependant, l'intervention humaine (notamment l'élimination de la couche d'isolation que représente le sol et la végétation) est peut-être un facteur encore plus important de la dégradation localisée du pergélisol que les effets du changement climatique.

Frost action (daily freezing and thawing) is most effective in continental and high latitude mountains where seasonal snow cover is thin. It contributes to shattering of bedrock, heaving and churning of soil, and slow downslope creep of the unfrozen soil mantle in summer. The effects of climate change on these processes are hard to estimate, but they are likely to be minor in comparison with the normal place to place variations in frost action intensity that result from differences in slope steepness, aspect, soil moisture, vegetation cover, and snow depth.

L'action du gel et du dégel quotidiens est particulièrement efficace dans les régions de montages continentales et de hautes montagnes où le manteau nival saisonnier est mince. Ce processus provoque l'éclatement du substratum, le soulèvement et le brassage du sol et la reptation lente de la couche de sol dégelé en été. Les effets du changement climatique sur ces processus sont difficiles à déterminer. Ils devraient être négligeables en comparaison avec les variations normales de l'intensité de l'action du gel d'un endroit à l'autre, compte tenu des différences dans l'inclinaison et la morphologie des versants, l'humidité du sol, la couverture végétale et l'épaisseur de la neige.

In mountainous regions with relatively mild winter temperatures (see table in Fig. 1), the combined effects of winter warming and increased precipitation will result in a rise in elevation of the winter snowline, a decrease in the extent of seasonal snowcover at low elevations, and increased snowfall at higher elevations. An adverse effect on low-elevation ski areas (shorter season, wetter snow conditions, increased rain) is expected, whereas increased snowfall may benefit resorts at higher elevations. In relatively cold mountain regions, a more general increase in snowfall is expected. Activities such as transportation and mining in alpine areas will also be affected by changing snow conditions, with the adverse effects of increased snowfall perhaps being more widespread than the beneficial effects of less snow and a shorter snow season.

Dans les régions montagneuses où les températures hivernales sont relativement douces (voir tableau de la figure 1), les effets combinés du réchauffement et de l'augmentation des précipitations pendant la saison froide se traduiront par une élévation de la limite des neiges d'hiver, une réduction de l'étendue du manteau nival saisonnier à faible altitude et une abondance accrue des chutes de neige à haute altitude. On prévoit des effets néfastes pour les stations de ski de basse altitude où la saison sera plus courte, la neige plus mouillée et les pluies plus abondantes. Quant aux centres de villégiature de plus haute altitude, ils tireront profit de la plus grande abondance de neige. Dans les régions montagneuses relativement froides, on s'attend à une augmentation plus générale des chutes de neige. Certaines activités comme le transport et l'exploitation minière dans les régions alpines devront également s'adapter à l'évolution des conditions, les effets négatifs de l'accroissement des chutes de neige pesant peut-être plus lourd que les effets bénéfiques d'une saison hivernale plus courte marquée par des neiges moins abondantes.

In regions where snowfall will increase, snow avalanche activity might also be expected to increase. However, this tendency may be offset by the effects of warmer atmospheric

Dans les régions où les chutes de neige augmenteront, on peut s'attendre à une multiplication des avalanches. Cette tendance pourrait néanmoins être contrebalancée par les effets d'une temperatures, which could reduce temperature gradient in the snow pack thereby increasing its stability, and tree colonization of initiation zones formerly above timberline. Future changes in storm patterns, including wind strengths and directions, will also have effects that cannot be predicted yet. In areas close to snowline, avalanche hazards are likely to decrease.

The extent of permanent snow and ice cover, in the form of glaciers and icefields, is controlled both by snowfall amounts, which determine ice accumulation, and by summer temperatures, which determine melting. In general, and especially in warmer areas, the effects of increased temperature are expected to dominate, resulting in glacier recession. Quantitative estimates of the change in elevation of the summer snowline suggest that many small, relatively low-elevation glaciers will disappear, and larger glaciers and icefields will experience drastic shrinkage. Thus the appearance of alpine landscapes will change dramatically, possibly with long-term impacts on tourism in national parks.

During glacier recession, hazards from floods and debris flows are likely to increase. (See Fig. 1 for generalized distribution of glaciers.) Catastrophic floods that far exceed flows generated by rainstorms and snowmelt result from the bursting of glacier dams and moraine dams. On steep slopes, outburst floods may be transformed into debris flows, and debris flows may also be generated by intense precipitation on steep moraines recently uncovered by glacier recession. Flows and floods generated in alpine areas may travel downstream below timberline for many tens of kilometres, posing a hazard to humans and structures such as bridges and roads.

The effects of climate change on gravitationally induced processes, such as landslides, rockfalls, and nonglacial debris flows, are hard to predict. Increased precipitation, particularly high intensity rainfall, may increase the frequency of some types of mass movement, such as debris slides and flows. Warmer temperatures and increased evaporation, on the other hand, may increase the stability of some slopes by reducing soil moisture and reducing the intensity of freeze-thaw and related processes, such as frost shattering and rock fall. (The intensity of mass movement processes is proportional, in a very general sense, to the elevations of the mountains shown in Fig. 1.)

If, as is likely, climate warming is accompanied by increasing storminess and intense rainfall, then peak stream discharges are likely to increase, resulting in increased erosion, sediment transport, and deposition downstream. Increased sediment input to streams in recently deglaciated areas will augment these effects, and, like outburst floods, effects may become apparent at considerable distances downstream from alpine areas. In both cases, hazards will include increased flooding (possibly both magnitude and frequency),

hausse des températures atmosphériques, laquelle se traduirait par une réduction du gradient de température du manteau nival qui, à son tour, amènerait une plus grande stabilité du manteau et la pousse d'arbres dans des zones auparavant au-dessus de la limite forestière, où les avalanches se déclenchaient. Les modifications futures dans le régime des tempêtes, notamment de la force et de la direction des vents, auront également des incidences encore imprévisibles. Dans les régions voisines de la limite des neiges, les risques d'avalanche devraient s'atténuer.

L'étendue de la neige et de la glace permanentes, sous la forme de glaciers et de champs de glace, dépend de l'abondance des chutes de neige (qui détermine quelle épaisseur de glace s'accumulera) et des températures estivales (qui déterminent quelle épaisseur de glace fondra). Dans les régions plus chaudes, les effets d'une augmentation des températures devraient en général dominer et causer le recul des glaciers. Les estimations quantitatives du déplacement vers le haut de la limite des neiges en été révèlent que de nombreux petits glaciers d'altitude relativement faible disparaîtront et que les gros glaciers et champs de glace connaîtront un retrait spectaculaire. L'apparence des paysages alpins sera donc profondément modifiée, ce qui risque d'avoir des répercussions à long terme sur le tourisme dans les parcs nationaux.

Durant le recul d'un glacier, il est fort problable que les risques de crue et de coulée de débris grimpent (figure 1, répartition générale des glaciers). La rupture de barrages glaciaires et morainiques peut provoquer des crues beaucoup plus catastrophiques que celles causées par de fortes pluies et la fonte de la neige. Sur les versants abrupts, les débâcles glaciaires peuvent se transformer en coulées de débris. Il est cependant à noter que ces dernières peuvent également être provoquées par des précipitations intenses sur les moraines à pente abrupte, récemment mises au jour par le recul d'un glacier. Les coulées et les crues qui se produisent dans les régions alpines peuvent descendre un cours d'eau sur plusieurs dizaines de kilomètres, bien au-delà de la limite forestière, mettant en danger des vies humaines et risquant de détruire des ouvrages comme des ponts et des routes.

Les répercussions du changement climatique sur les processus déclenchés par gravité (gravitaires), comme les glissements de terrain, les éboulements et les coulées de débris non glaciaires, sont difficiles à prévoir. Les précipitations accrues, surtout les pluies de forte intensité, pourraient augmenter la fréquence de certains types de mouvement de masse, comme les glissements de terrain et les coulées de débris. Par ailleurs, l'élévation des températures et l'accentuation de l'évaporation pourraient stabiliser certains versants en réduisant l'humidité des sols et en atténuant l'action du gel-dégel et des phénonèmes associés comme la gélifraction et les éboulements. L'ampleur des mouvements de masse est proportionnelle, dans un sens très général, à l'altitude des montagnes montrées sur la figure 1.

Si, ce qui est fort probable, le réchauffement climatique s'accompagne de tempêtes et d'orages en plus grand nombre, le débit de pointe des cours d'eau devrait s'accroître, de manière à amplifier l'érosion, le transport des sédiments et la sédimentation en aval. Le plus grand apport de sédiments dans les cours d'eau des zones récemment déglacées accentuera les effets du réchauffement et, tout comme les débâcles glaciaires, les conséquences pourraient se manifester sur des distances considérables en aval des régions alpines. Dans les deux cas, les dangers seront liés aux rise of the stream bed causing increased elevation of flood waters, and increased bank erosion and lateral instability of river channels. Figure 1 shows streams draining from alpine areas.

Where snow packs are reduced, the water storage capacity and the spring and early summer water yield of drainage basins will be reduced. Figure 1 shows major rivers with spring and summer flow levels that are significantly augmented by snowpack and glacier melt. Where glacier recession reduces the extent of permanent snow and ice, late summer water yield will be reduced also, and the buffering capacity of a glacier to reduce year-to-year variations in runoff from precipitation will also decline. These potential changes are particularly significant in regions that are dry in summer, such as the Okanagan Valley in south central British Columbia, where agriculture and industry receives its water supply from adjacent mountains. Where snow accumulation is increased, water storage and runoff will be augmented.

crues (probablement plus intenses et plus fréquentes), à la plus grande épaisseur du lit des cours d'eau (augmentant le niveau des eaux de crue), à l'érosion plus intense des berges et à l'instabilisation latérale des chenaux. La figure 1 montre les cours d'eau ayant leur source dans les régions alpines.

Là où le manteau nival aura diminué, on observera une réduction de la capacité de stockage de l'eau et de l'apport d'eau par les bassins versants au printemps et au début de l'été. La figure 1 montre les grandes rivières de la région où le débit printanier et estival est significativement accru par la fonte du manteau nival et des glaciers. Là où le recul glaciaire réduira l'étendue de la neige et de la glace permanentes, l'apport d'eau de la fin de l'été s'en trouvera diminué. Il en sera de même de la capacité tampon d'un glacier à réduire les variations annuelles dans l'écoulement des eaux de ruissellement. Ces changements potentiels sont particulièrement importants dans les régions où le climat est sec en été, comme dans la vallée de l'Okanagan, dans le centre sud de la Colombie-Britannique, où l'agriculture et l'industrie utilisent l'eau provenant des montagnes adjacentes. Là où la neige sera plus abondante, le stockage de l'eau et le ruissellement augmenteront.

INTRODUCTION, OBJECTIVES, AND SCOPE OF REPORT

In alpine areas, many effects of severe climate are quite obvious in the landscape. Trees are absent or reduced to wind-sculpted shrubs; other vegetation seeks the shelter of the ground, and snowbanks persist well into summer. Human occupancy in these regions has also adapted to the climate, with some activities, such as skiing, even taking advantage of it.

Less apparent to the casual observer are the geomorphological processes that result from the interaction of climate with the rocks and soils of alpine areas. These range from dramatic events such as snow avalanches and debris flows, to the imperceptibly slow but effective downslope movement of soil and glaciers. Although catastrophic events can be hazardous to people engaged in even the most transient of pursuits, such as hiking and cross-country skiing, the effects of slow but persistent processes must be accommodated in the design of roads and other structures. In one way or another, these processes impinge on all human endeavours in this environment. With increasing use of alpine areas for both recreation and industry in the future, the impact of the geomorphological processes will similarly increase.

Geomorphological processes are related to climate in a complex manner, and neither they nor the anticipated climate changes are known sufficiently well to provide other than tentative and qualitative conclusions about the future. Also, in many cases, climate changes will give rise to opposing effects, for example both warmer temperatures and increased snowfall in relatively cold alpine areas, and it is hard to predict which of these effects will exert the stronger influence on phenomena such as glaciers and avalanches. In general, the changes to alpine areas that are described here will take place slowly over the course of decades, although some specific effects, such as a glacier outburst flood, may occur suddenly.

The frequencies, magnitudes, and rates of geomorphological processes will likely be altered due to human-induced global warming, just as they have changed in the past due to natural modifications of climate. Some evidence of past climate fluctuations is preserved in alpine areas, for example, in the moraines which indicate the former greater extent of glaciers and in the relict soil patterns that indicate the former extent of permafrost. From this kind of information and from process studies, some of the likely effects of future climatic warming can be estimated, although in a rather general fashion.

This report is designed to be used with Figure 1 (in pocket). Report objectives are to inventory, describe, and classify the geomorphological processes of the alpine areas in Canada, and to indicate how these processes may be affected by prospective changes in climate, paying special attention to those processes that impinge on human activities.

The nature of the expected climate change is not yet well defined. Current global climate models do not include realistic treatment of some factors that have a profound influence on climate, such as the effects of ocean-atmosphere interaction, clouds, sea-ice, land-surface hydrology, and the biosphere. Also, the sources and sinks of greenhouse gases, and natural climate variability are incompletely understood. Another shortcoming is that global climate models do not provide information about the rates of climate modification over time.

The best quantitative estimate of climate change that is currently available comes from a subjective comparison and summary of the outputs of various global climate models by the Intergovernmental Panel on Climate Change (IPCC) (Mitchell et al., 1990). This paper concluded that for a doubled atmospheric concentration of greenhouse gases, the equilibrium global mean temperature will increase by about 2.5°C (1.5°C to 4.5°C) and global average precipitation and evaporation will increase by 3 to 15%. The IPCC projections show that doubling of greenhouse gases could occur within the next 30 to 40 years if nothing is done to limit emissions.

Changes in temperature and precipitation across Canada, based on IPCC reports, which are summarized in "Canada's National Report on Climate Change, 1994" (Government of Canada, 1994), are as follows. Global climate models predict that temperature increases will be above the global mean in winter, with greater increases (6°C) in the continental interior and less (4°C) near the Pacific and Atlantic coasts. Some global climate models predict a winter temperature increase of as much as 10°C in the High Arctic. Summer temperature increases will be less, reaching 4°C except along the Pacific and Arctic coasts. Increased winter precipitation is predicted

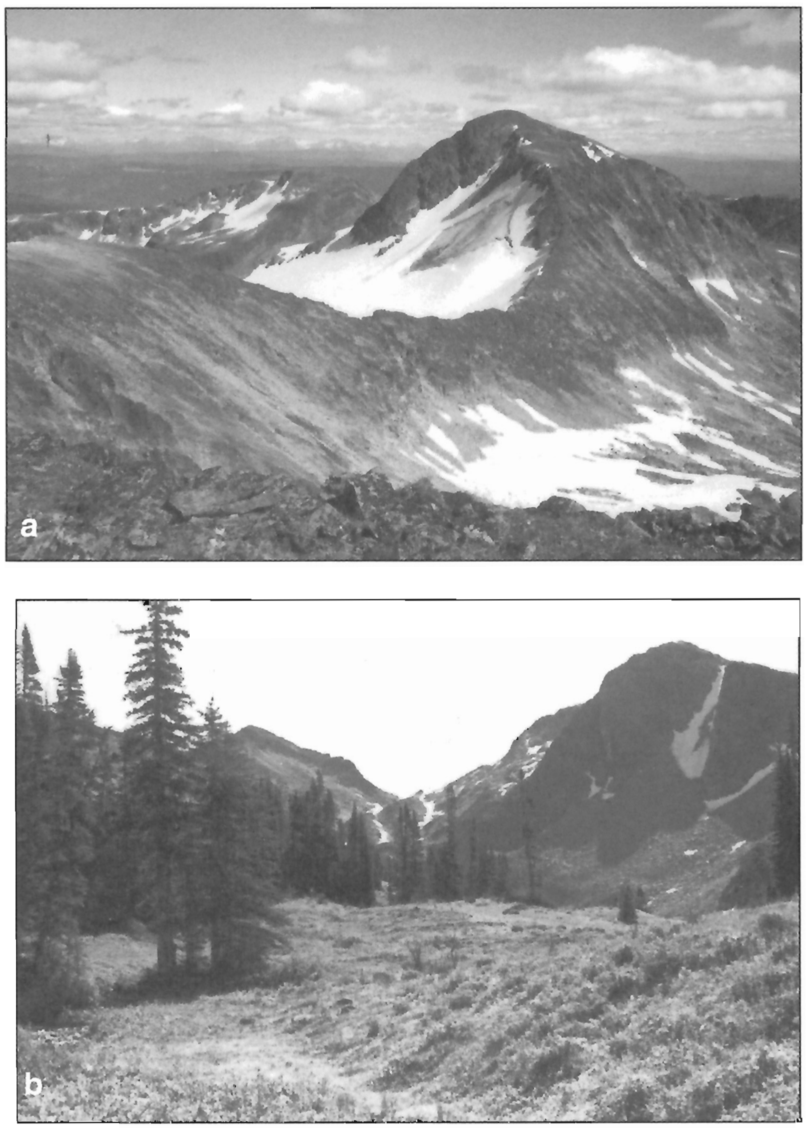


Figure 2. a) *Alpine area at about 2500 m on Trophy Mountain, Cariboo Mountains, central British Columbia. Small glaciers occupy north-facing cirques on subdued glacial horns; diamond shaped scar on nearest glacier marks a recent slab avalanche; vegetation is restricted to small herbaceous plants and lichen. GSC 1997-012A;* ***b)*** *Subalpine parkland in the Stein River basin at about 1850 m, Coast Mountains, British Columbia. Scattered clumps of subalpine fir and Englemann spruce separate openings with lush herbaceous vegetation. GSC 1997-012B*

for the entire country. Summer precipitation is projected to increase north of 60°N, but there is no consensus as to the change expected south of this latitude. Thus southern Canada could experience reduced snowfall and increased rainfall, and winter snow cover may become intermittent in regions where it now persists throughout the winter. Significant reductions in the duration of seasonal snow and ice cover could occur in higher latitudes, although snowpack might increase in areas where winter temperatures remain below 0°C.

The effects of climate change in the alpine zone, which typically consists of small discrete areas, are not specifically addressed by global climate models. Computing limitations require that models be based on a coarse network of points, typically equivalent to about 5° latitude, and climate outputs are presented for sea level (Barry, 1992). The latter restriction can be overcome to some extent for small areas by using quantitative techniques to determine the effects of elevation. For the purposes of this report, changes in alpine climates are assumed to reflect the broad regional trends, although the magnitude of the anticipated changes is not known and will probably differ with elevation and from one alpine area to another. Consequently, this report will address the effects of climate change as described in the above paragraph on Canada, but in a generalized, nonquantitative manner. Another difficulty stems from uncertainty about changes of significance to alpine areas that result from the interaction of opposing effects, such as snowpack thickness, which depends upon the balance between increased snowfall and warmer winter temperatures, and soil moisture content, which is largely dependent on the balance between evaporation (i.e. summer temperatures) and precipitation.

ALPINE AREAS OF CANADA

Definition

The alpine zone includes areas of ground-hugging, shrubby and herbaceous vegetation (alpine tundra), barren rocky areas at high elevations, glaciers, and perennial snowbanks. For the purposes of this report, *alpine areas* are defined as land above the altitudinal timberline (the upper limit of forest) (Fig. 2).

Under this definition, mountains in treeless arctic terrain, such as those of Baffin and Ellesmere islands, are excluded, and areas of low relief that lie above timberline, such as high plateaus, are included. The justification for this is as follows. Geomorphological processes are strongly influenced by climate, and climate varies with elevation. In the temperate zone, geomorphological processes and landforms at high elevations are dissimilar to those at elevations near sea level. Conditions above timberline are characterized by widespread glaciers and permafrost, relatively heavy precipitation, severe frost action, and the influence of forest vegetation is lacking. In arctic areas, differences between mountains and lowlands are much less pronounced. Permafrost and severe frost action are ubiquitous, glaciers commonly descend to sea level, and deep-rooted vegetation is absent. Thus arctic mountains are more appropriately discussed together with their neighbouring lowlands. Geomorphological processes and landscapes in the alpine tundra zone are dissimilar to those of their surroundings. These are the subject of this report.

The subalpine zone, with its discontinuous tree cover and stunted, wind-shaped trees, is included in the above definition of alpine areas. An excellent nontechnical account of the physical and biological conditions of this transition zone is provided by Arno and Hammerly (1984). They also described the variability of the timberline environment, both physically and floristically, within the various mountain ranges of North America.

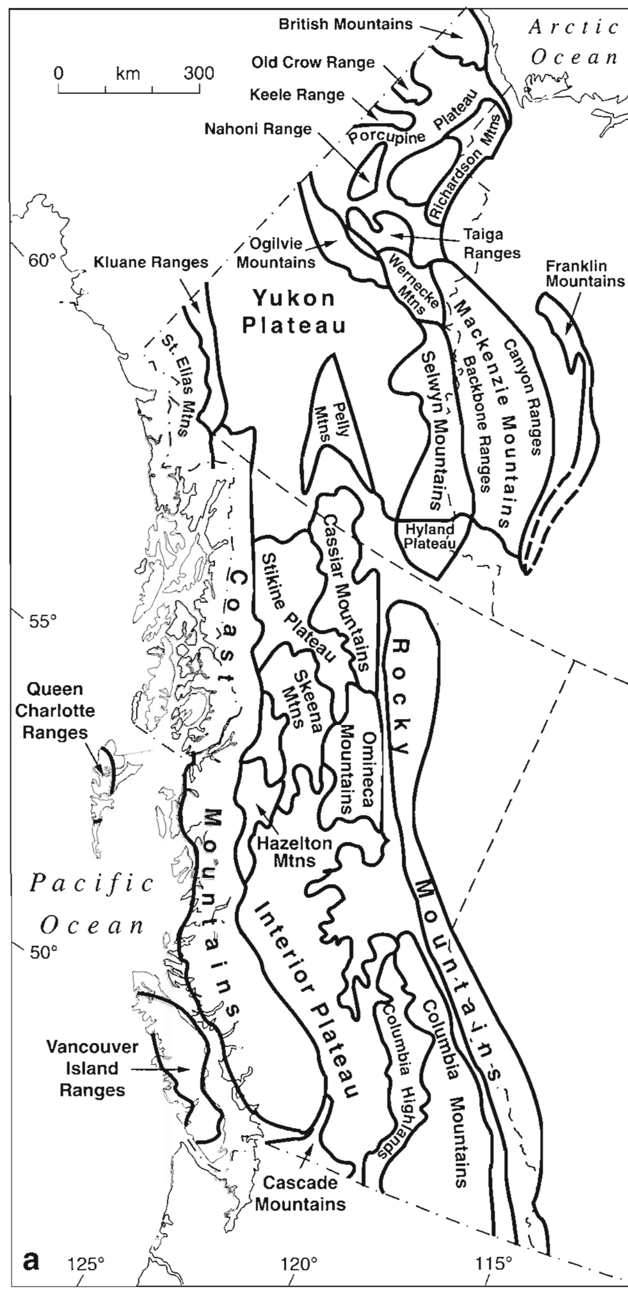


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Figure 3. *Location maps for physiographic regions that include alpine areas:* ***a)*** *western Cordillera;* ***b)*** *eastern Canada (after Bostock, 1970).*

Regional distribution and characteristics

Alpine areas are widespread in the mountains and plateaus of the Western Cordillera, and large tracts of the mountains and uplands of Labrador and Newfoundland also lie above timberline (Fig. 1, 3, 4). The highest timberline in Canada exists at an elevation of about 2100 m in the Rocky Mountains of southwestern Alberta and southeastern British Columbia (see 'Timberline' section). Westward and northward, timberline decreases in elevation, descending to about 1500 m in the vicinity of Vancouver and slightly lower than this in the Rocky Mountains at the boundary between British Columbia and Yukon Territory. Many peaks in the Rocky Mountains, the Columbia-Omineca Mountains, and the Coast Mountains are more than 1000 m above timberline, and so a substantial amount of rugged topography lies within the alpine zone (Fig. 5). The continuity of this alpine terrain is commonly interrupted by deep valleys, however, especially in the southern parts of the Cordillera.

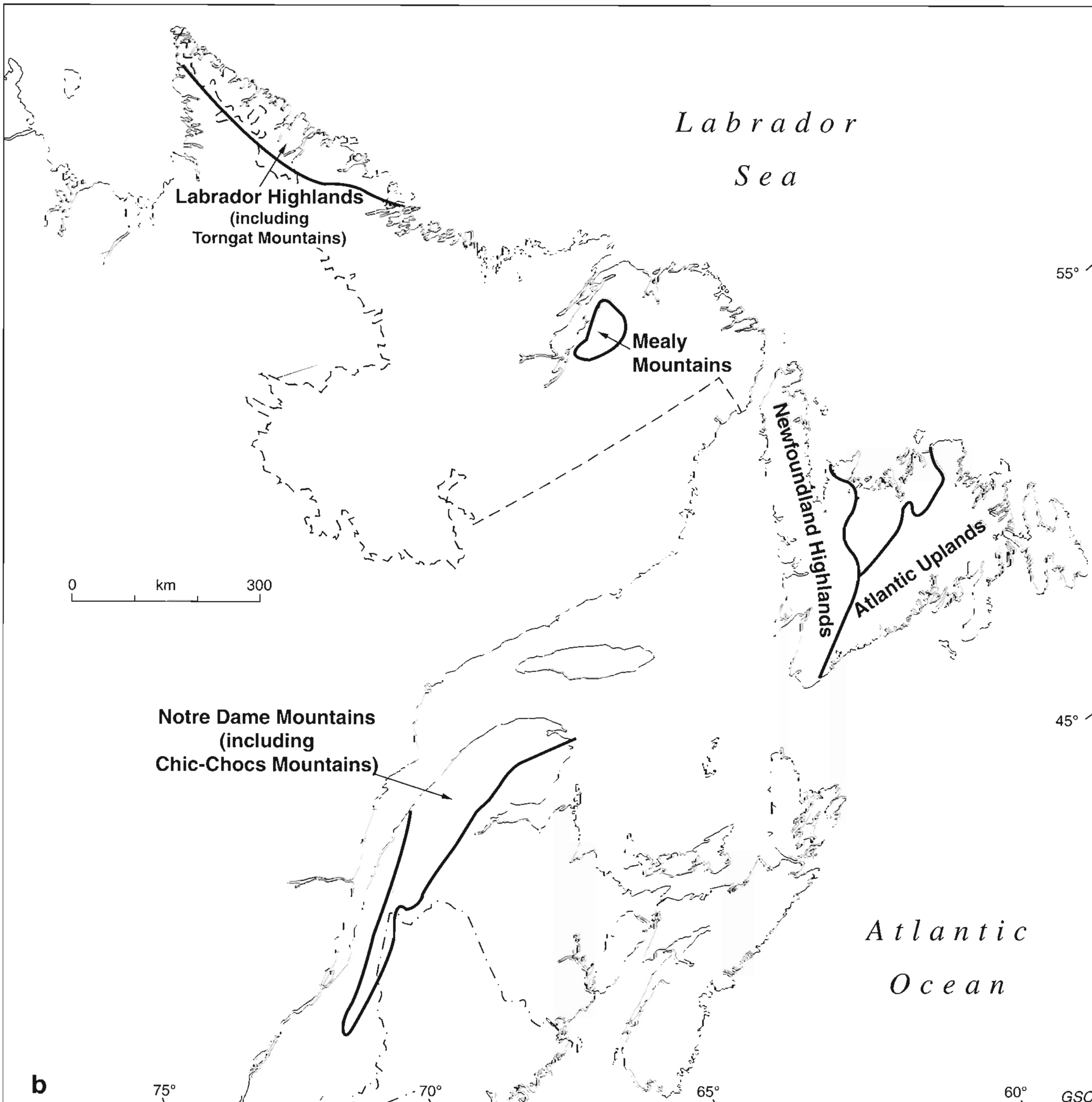


Figure 3. (cont.)

Farther north, the altitudinal timberline descends gradually until it merges with the latitudinal timberline (latitude at which tree growth stops) in the mountains of northernmost Yukon Territory and adjacent District of Mackenzie. The British Mountains and northern Richardson Mountains are included as alpine areas in Figure 1, although the scattered stands of trees that occur here hardly constitute a continuous timberline (Fig. 6a). In these more northerly regions, alpine areas are larger and more continuous. In northern British Columbia and Yukon Territory, many plateaus rise above timberline and support extensive areas of alpine terrain (Fig. 6b).

In eastern Canada, timberline declines northward from 1000-1100 m on Mount Jacques-Cartier in the Chic-Chocs Mountains of Gaspésie (Boudreau and Payette, 1974), to about 550 m in the Mealy Mountains of southern Labrador, to sea level at about 58°N (Fig. 7). Along the coasts of Labrador and Newfoundland the elevation of timberline varies greatly in response to contrasts in exposure to severe weather conditions; a strip of tundra extends southward along the coast of Labrador, and broad, treeless areas extend down to sea level in Newfoundland (Fig. 8). These areas are included with the alpine areas discussed in this report because geomorphological processes in the treeless landscapes are essentially the same

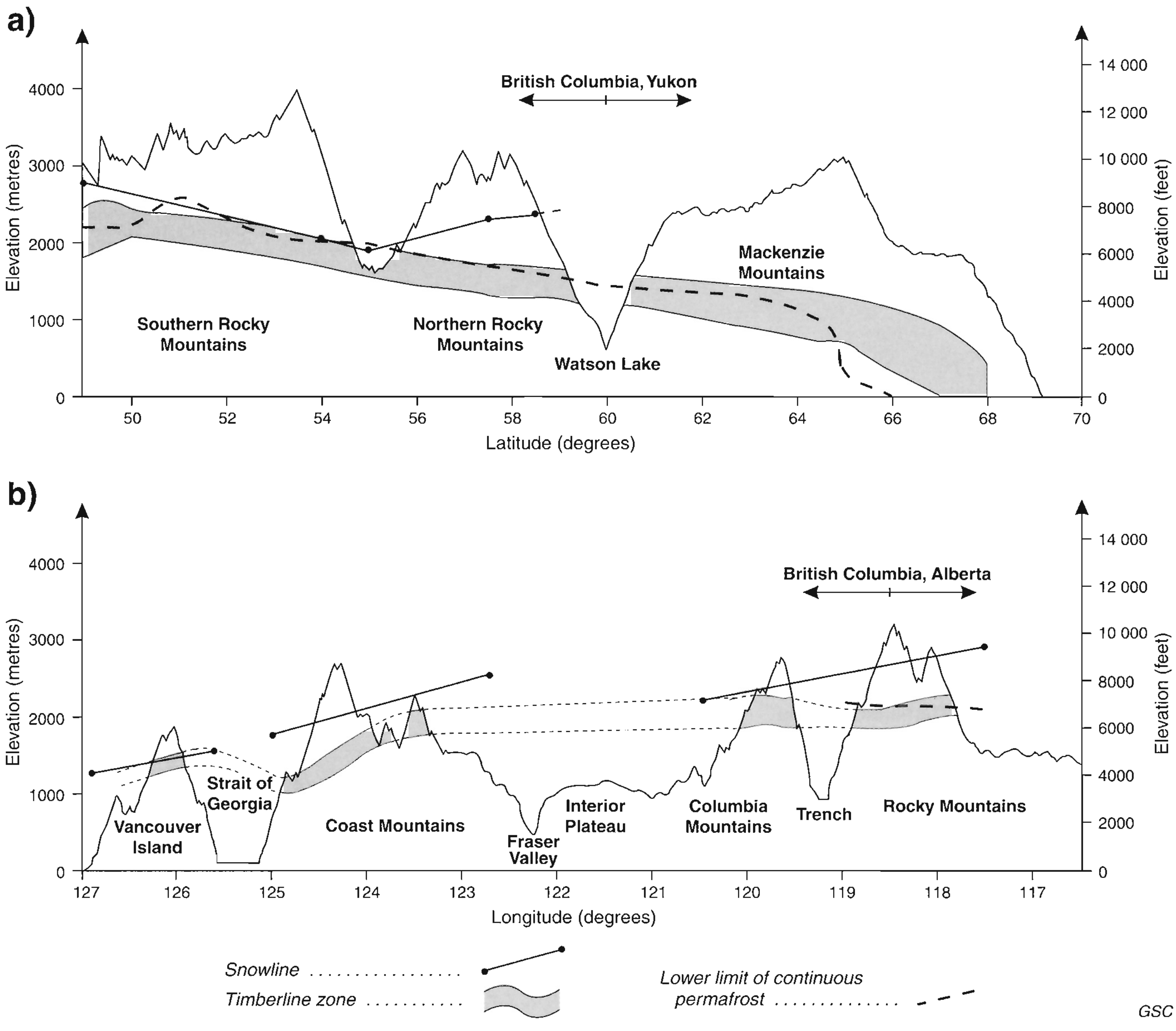


*Figure 4. **a)** Profile from 49°N to the Arctic Ocean along the crests of the Rocky Mountains and the Mackenzie Mountains, showing the timberline zone (Arno and Hammerly, 1984), lower limit of continuous permafrost (Harris, 1986), and annual snowline (glaciation limit) after Ostrem (1966) and Evans (1990). **b)** East-west profile across the Cordillera between 49° and 55° N showing the same features, except the permafrost limit is shown for the Rocky Mountains only (same sources as 4a).*

Figure 5. ***a)*** *Typical alpine area in the Coast Mountains (Tantalus Range) British Columbia; note serrate ridge crests above glacier basins (cirques), small glaciers, perennial snow patches, and avalanche tracks. GSC 1997-012C* ***b)*** *Typical alpine area in the Cascade Mountains near Mount Hewitt Bostock, British Columbia; cirques are common but horns and serrate ridges are absent. Rounded ridge crests are remnants of an old erosion surface, the upland surface of the interior plateaus. All summits were overridden during the last Pleistocene glaciation. GSC 1997-012D* ***c)*** *Alpine landscape in the Rocky Mountains, including Mount Assiniboine, British Columbia and Alberta; note serrate skyline, cirques and troughs, and stratification of sedimentary rocks. GSC 1997-012E*

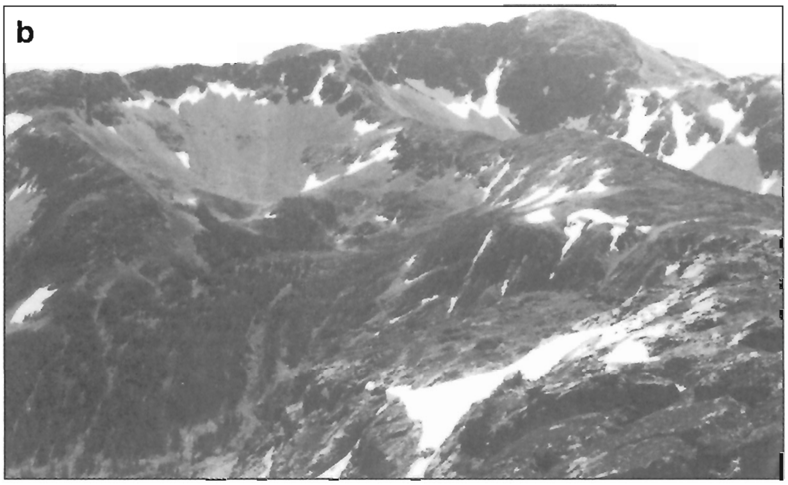
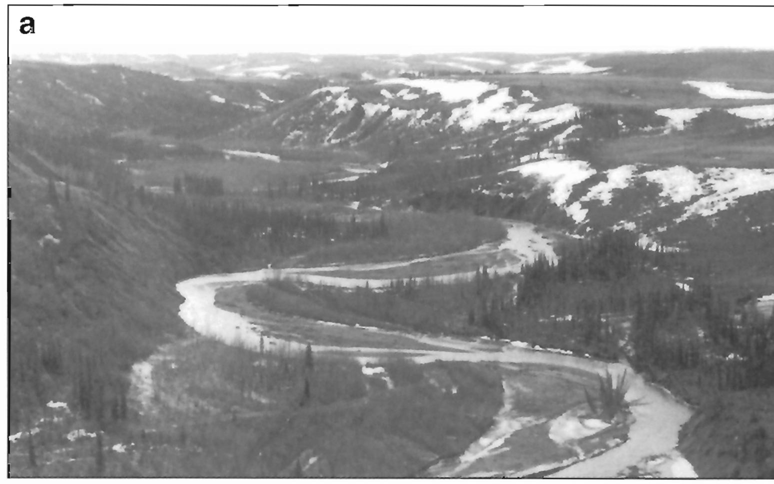
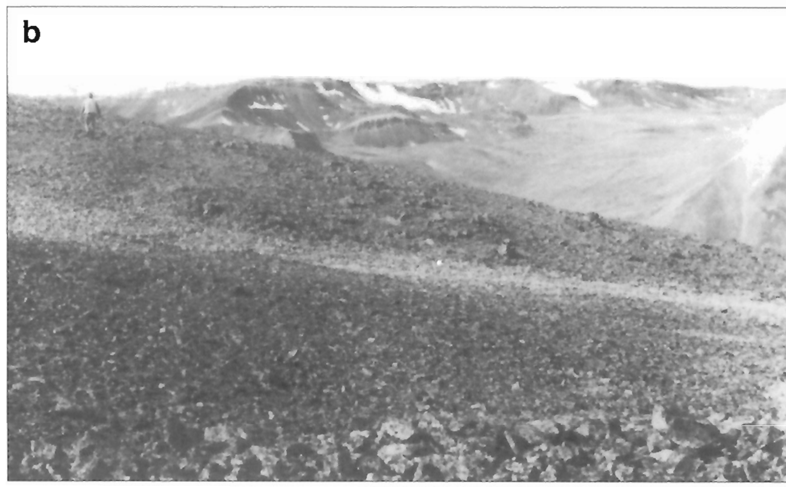


Figure 6.

a) *Landscape along Stoney Creek in the northern Richardson Mountains, Northwest Territories. No continuous treeline exists here: trees cluster in sheltered locations; note gently undulating alpine area in distance. Photograph courtesy of M. Church.* ***b)*** *Alpine plateau on Level Mountain, Stikine Plateau, north-central British Columbia. Note cirques (with snowbanks) cut into plateau rim; frost-shattered rock in foreground. GSC 1997-012F*





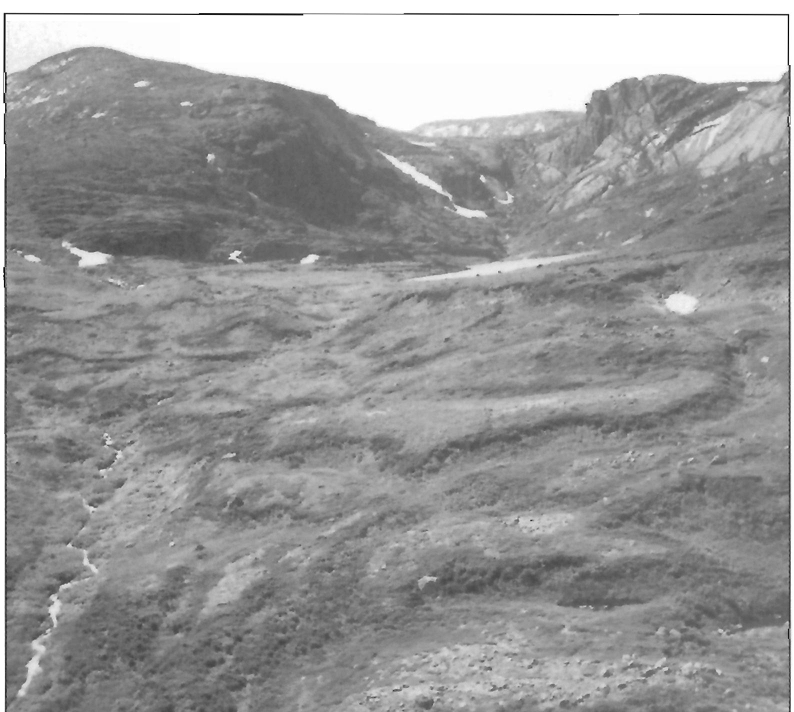


Figure 7.

Alpine landscape in the Mealy Mountains of Labrador. Note nivation hollows and benches in the foreground. Photograph by R.J. Fulton. GSC 161595

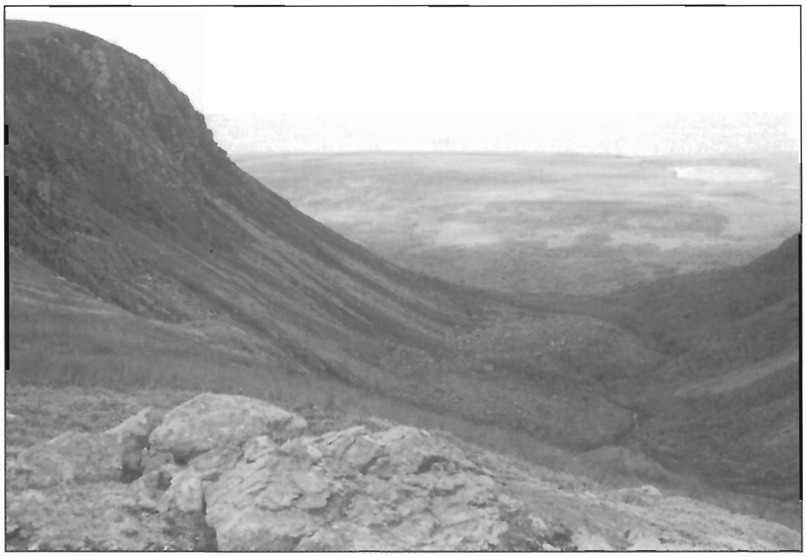


Figure 8.

Long Range Mountains, Newfoundland; view westward from 450 m elevation, down an old glacial trough that was not occupied by ice during the last glaciation. Note talus and rock glaciers on north-facing slope with patchy cover of dwarf forest, and coastal plain beyond with a mosaic of sedge bogs (pale patches) and dwarf spruce. Photograph courtesy of I.A. Brookes.

Figure 9. *Rugged unglaciated terrain in the northern Mackenzie Mountains, Northwest Territories; vegetation is sparse on slopes underlain by limestone. Photograph courtesy of M. Church.*

as those in alpine areas at higher elevations. Along the eastern side of Ungava Peninsula, the latitudinal and altitudinal timberlines merge into an extremely broad transition zone on a long, gentle incline (Arno and Hammerly, 1984). In Figure 1, timberline in this area was drawn arbitrarily at about 300 m a.s.l.

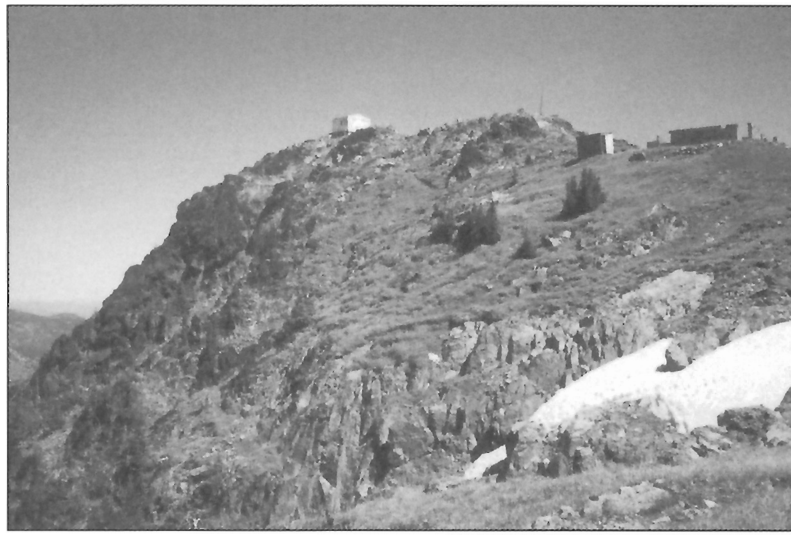
The types of geomorphological processes that operate within alpine areas and their severity are strongly influenced by local topography, particularly slope steepness, by weather and climate, by local bedrock and surficial materials and their weathering products, and by the presence or absence of permafrost. Some of these conditions are shown in Figure 1 and described briefly here (see Appendix A). The lack of trees and the absence of the stabilizing effects of their extensive root systems also affect slope processes in alpine areas.

The topographic characteristics of alpine areas are shown in a generalized form in Figure 1 because slope steepness and topographic detail cannot be represented adequately on a small scale map. Most physiographic regions within which the alpine areas occur are shown as either *mountains* or *plateaus* (sometimes called highlands). Mountains are characterized by steep slopes at high elevations, rugged topography, and serrate skylines (e.g. Fig. 2, 5a, c). By way of contrast, plateaus consist of level or gently sloping terrain at high elevations, overlooking deep, steep-sided valleys (e.g. Fig. 6b). A few relatively low-elevation, gently sloping alpine areas occur on *hills* or in areas of *low relief* (Fig. 6a).

Alpine landscapes display strong evidence of their history. Most alpine areas were sculpted by ice during the glaciations of the Pleistocene Epoch. In the mountains, glacial horns, knife-edged ridges, cirques, and U-shaped valleys dominate the scenery (Fig. 5a, c). On plateaus, many of which were overridden by an ice sheet during glacial maxima, horns are absent, and knife-edged ridges are uncommon, although cirques and U-shaped valleys are abundant (Fig. 5b, 6b). Glacial deposits are widespread in both landscapes. The

Figure 10.

Old Glory Mountain in the Monashee Mountains, southern British Columbia. The ruined buildings at upper right are the remains of the highest weather station in Canada (2350 m), which operated from 1945 to 1967 when it was destroyed by fire. The white cabin on the summit is a forest fire lookout. GSC 1997-012G



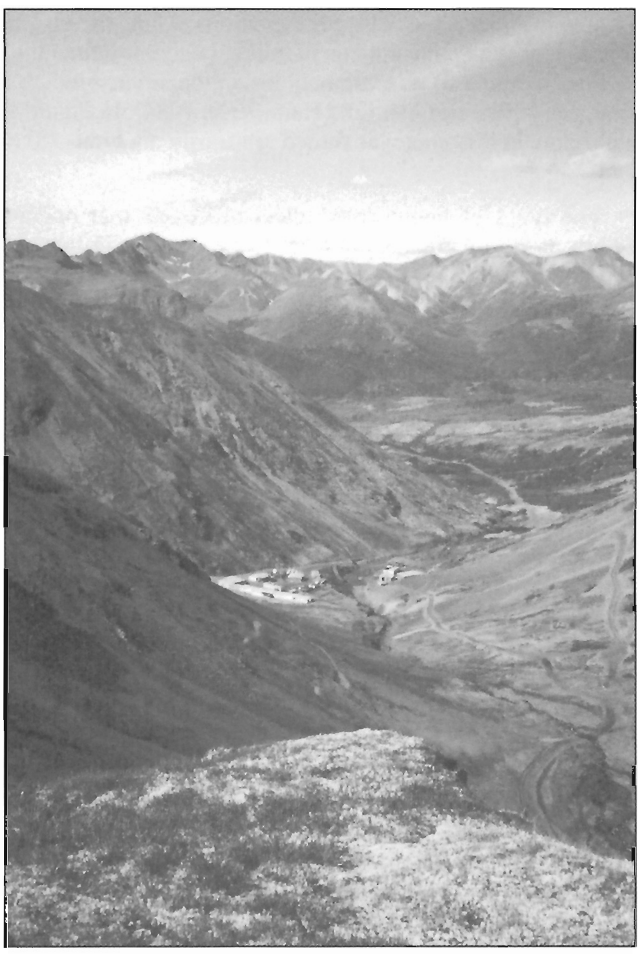


Figure 11. *A mining camp and exploration roads at 1225 m elevation in the rugged Selwyn Mountains, Yukon Territory. GSC 1997-012H*

nonglacial geomorphological processes described below have modified these glacial landscapes during the relatively short span (10 000 to 12 000 years) of postglacial time.

Large areas of central and northern Yukon Territory and the adjacent Mackenzie Mountains in the Northwest Territories remained ice-free during Pleistocene glaciations (cf. Prest et al., 1967). Landforms here reflect the effects of weathering, running water, mass movement, and other geomorphological processes operating over a time span of millions of years (Fig. 9).

Alpine weather and climate are characterized by their temporal and spatial variability, rather than by any particular set of attributes. Topographic irregularities and elevation differences give rise to marked changes over short distances in parameters such as temperature, precipitation, snow depth, and soil moisture. Most of this variability is not documented by standard climate records because there are almost no climate stations in alpine areas (Fig. 10). The broad regional climatic trends of precipitation and temperature that are recorded by lowland weather stations, however, do also occur in alpine areas, and, in fact, are reflected by alpine geomorphological features. These broad patterns have been indicated in Figure 1 by the delimitation of climatic regions. Regional boundaries represent gradual transitions of climate. Thus regions of wet, windward slopes are distinguished from the considerably drier rainshadow areas on the leeward sides of the mountains; for example, see the east and west subdivisions of the southern Rocky Mountains. Relatively temperate alpine regions are distinguished from colder subarctic regions; for example, see the southern and northern subdivisions of the Rocky Mountains. In this report, the terms *maritime* and *continental* are used to distinguish coastal mountains, which experience equable, oceanic climates, from inland mountains that experience drier, more extreme (greater annual and diurnal temperature ranges) climates.

Human use of the alpine zone includes such diverse activities as recreation, hunting, mining, cattle grazing, and use by native peoples, with recreation predominating in the less extensive but relatively accessible alpine areas adjacent to southern population centres (Fig. 11). Roads, power lines, pipe lines, communication beacons, and other facilities are also located in the alpine zone. Settlements in alpine areas range from the overnight camps of hikers and cross-country skiers to seasonally occupied mining and hunting camps and year-round resorts and lodges. Many users of the alpine zone are exposed to potentially hazardous geomorphological processes, such as snow avalanches and rockfalls. *Hazards* are processes with considerable potential to damage structures and injure humans, and not readily controlled by engineering works. (Hazardous weather conditions are not considered in this report.) Some activities in the alpine zone are affected by terrain conditions that are a nuisance or that make construction more expensive, but are not life-threatening, such as boggy ground (i.e. organic soils) and permafrost. These are referred to as *constraints* rather than hazards.

TIMBERLINE

Definition

Timberline is the boundary between montane forests and alpine tundra. Although this boundary appears as a line on most maps, when viewed more closely, it is a belt or zone of transitional vegetation (the timberline ecotone) that lies between continuous forest and the treeless alpine tundra (Arno and Hammerly, 1984) (Fig. 12). This zone, which is included in the definition of alpine areas used in this report, includes forest with openings, sometimes referred to as forest tundra, park-like areas with tree clusters (tree islands and ribbon forests), stunted and wind-distorted trees, and the krummholtz (low, horizontally-spreading bushy conifers) zone. The transition takes place over a range in elevation of about 300 m, although it may be narrower or broader, depending upon local conditions.

One of the most apparent effects of climate change in alpine areas is the up- or downslope migration of timberline. This can result in a great change in the areal extent of alpine areas (and the significance of alpine geomorphic processes), and so warrants discussion here.

Climatic controls

In general, the elevation of timberline is controlled by temperature and snowfall. Local variations in elevation derive from the effects of wind, severe winter conditions such as desiccation and intense sunlight, aspect (slope orientation),

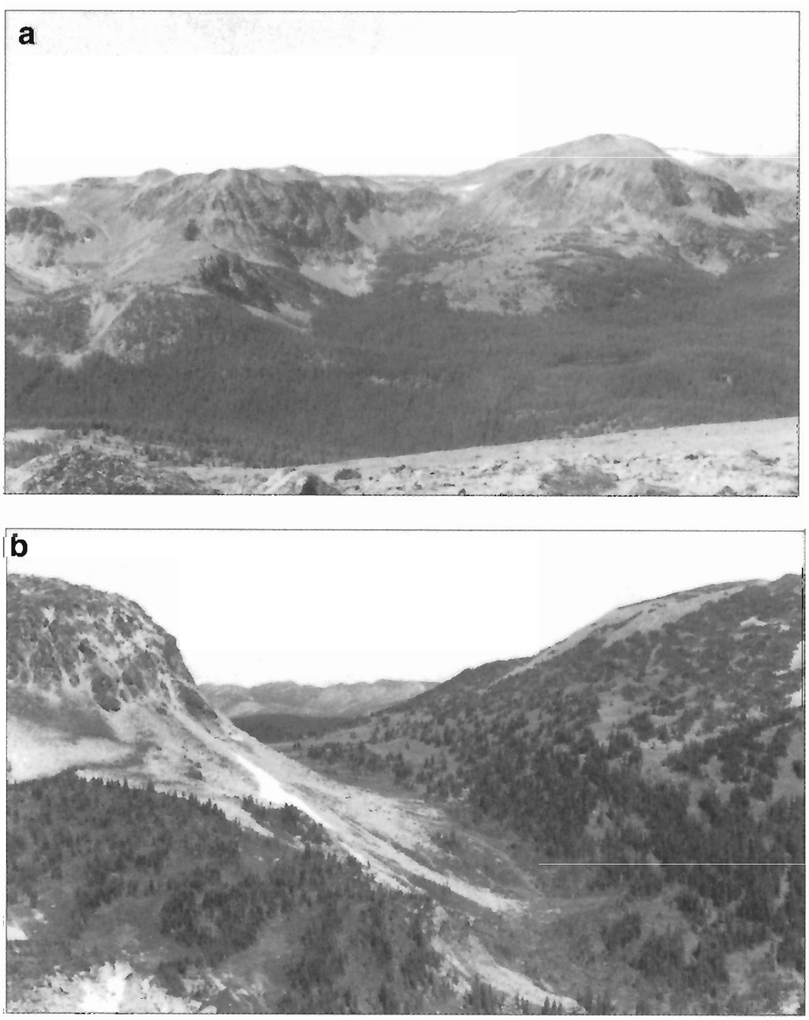


Figure 12.

a) *Timberline in Cathedral Provincial Park, southeastern Cascade Mountains, British Columbia. Note how the elevation and width of the timberline ecotone varies with terrain characteristics such as slope steepness and rockiness. GSC 1997-012I* ***b)*** *Timberline ecotone near Mount Hewitt Bostock, Cascade Mountains, British Columbia. Tree rows and islands occupy low mounds and ridges that have a longer snow-free season than adjacent depressions. The resulting parkland extends over a vertical range of about 300 m. GSC 1997-012J* slope steepness, soil types and moisture conditions, permafrost, and geomorphological processes such as snow avalanches (Arno and Hammerly, 1984). Timberline elevation is also influenced by nonclimatic effects such as fire and overgrazing (Griggs, 1938; Arno and Hammerly, 1984, p. 50-51).

The most significant limiting factor for tree growth is summer temperature. The relations between temperature and tree growth are complex; for example, different temperature controls apply to each biological process, such as germination, growth, and photosynthesis. A growing season of about two months is an additional requirement for tree growth. On a regional scale, however, both arctic and alpine timberlines appear to coincide with the 10°C isotherm for the warmest month of the year, usually July (Tranquillini, 1979). This straightforward measure can thus be used to estimate timberline positions. (That coniferous trees can survive extremely low temperatures during their period of winter dormancy is demonstrated by their growth in the coldest parts of Canada.)

The second significant factor, snowfall, influences timberline elevation through the adverse effects of late-lying snow on successful germination. Hence timberline is depressed in regions where snowfall is heavy and where cool summers retard melt. At the regional scale, other components of precipitation and soil moisture conditions probably do not restrict tree growth at timberline in Canadian mountains (Arno and Hammerly, 1984).

As a result, regional patterns of summer temperature and snowfall can thus be used to explain most of the observed variation in timberline elevation. The general west to east rise in timberline that occurs across each of the northwest-trending mountain belts in British Columbia reflects decreasing snowfall and increasing summer warmth (Fig. 4b). The general northward decline of timberline reflects increasingly cool summers at higher latitudes.

Effects of climate change

There have been numerous studies of the effects of past climate change on timberline elevations (e.g. see Beaudoin, 1989, and references therein for examples from the Canadian Rocky Mountains; and Lamb, 1985, for an example from Labrador). Evidence provided by pollen and macrofossils (wood, leaves, and seeds) preserved in bogs and lakes indicates that several times during postglacial time (the 10 000 to 12 000 years since the last major glaciation), timberline elevation changed by several tens of metres. The total postglacial range of timberline elevations in the southern Rocky Mountains may have been as much as 200 m, accompanying a 1.0°C range in average July temperatures (Luckman and Kearney, 1986). Relatively abrupt changes in climate that have occurred during the Little Ice Age of the past few centuries are evidenced by relicts of former forests at elevations above present timberline (Fig. 13). For example, Luckman (1994) has shown that dead trees on the slopes of Mount Wilcox near the Columbia Icefields colonized the slope as timberline rose between the late 1300s and mid-1500s; subsequently, a major dieback of trees occurred during cooling in the latter half of the 17th century. Young living trees on this same slope indicate that it is being recolonized, and that timberline is currently advancing (Kearney, 1982).

Many studies have reported that a current advance of alpine timberlines is a widespread phenomenon, although others have noted that the response of timberline vegetation to climate change is not straightforward. Invasions of tree seedlings into alpine meadows and increases in tree density in forests close to timberline have been reported from the mountains of both eastern and western Canada (e.g. Brink, 1959; Kearney, 1982; Payette and Filion, 1984, and references therein). Although in general, this phenomenon can be attributed to the climatic warming that has occurred since the end of the Little Ice Age (about the mid-nineteenth century), the process is complex. Successful germination depends upon the coincidence of high seed production with snow-free conditions, hence light snowfall or early melt, and warmth. Subsequently, the survival of seedlings is dependent upon

Figure 13. *Timberline ecotone on Old Glory Mountain, Monashee Mountains, British Columbia: standing snags are considerably larger than living trees and may be relicts from a time of warmer climate and higher timberline prior to the Little Ice Age of the seventeenth to nineteenth centuries. GSC 1997-012K*

summer warmth, a relatively long growing season, adequate summer moisture, and protective snow cover in winter. Some workers have noted the failure of timberline to respond, or a delayed response, to warming trends. Brink (1964), for example, noted that soil instability resulting from frost action and gravitational effects retarded colonization of ground exposed by glacier recession. Also, Payette et al. (1989) concluded that vegetation resilience to environmental change is a significant factor: recent warming has not compensated for the detrimental effects of Little Ice Age cooling at their study site at an exposed location in northern Quebec. Thus, climatic warming will probably result in a general upward migration of the timberline ecotone, although the vegetation response at any particular site may be strongly influenced by local physical conditions or retarded by biological constraints.

Accordingly, both the magnitude of future timberline rise and the associated decrease in the extent of alpine areas are hard to estimate. The average environmental lapse rate (temperature decrease with increasing elevation) is 0.6°C/100 m. Thus if temperature were the only variable involved, timberline would eventually rise by about 167 m for each 1°C of warming of mean July temperature. This assumption is obviously overly simplistic because it does not consider changes in other climate parameters, such as snowfall and soil moisture, nor changes in climate variability. It does suggest, however, that the projected timberline advance due to climate warming in the next century may be greater than that of the previous 10 000-12 000 years. The effect on the extent of alpine areas of a 200 m rise in timberline is illustrated in Figure 14. The reduction in the areal extent of the alpine zone will be greatest where extensive alpine areas lie at elevations close to present-day timberline, such as in some plateau areas of northern British Columbia, Yukon Territory, and Labrador.

The impact of a reduction in the extent of treeless alpine areas is hard to assess because it will occur slowly and irregularly, over many decades, as young trees expand upslope. Alpine meadows at relatively low elevations will be transformed into parkland or forest, and present subalpine parkland will develop a continuous tree cover. Clearly, the visual and aesthetic qualities of the present alpine areas will be significantly changed, and hence their use for recreation and tourism. The extent of alpine meadows available for wildlife and for summer grazing of cattle will decline, unless the establishment of tree seedlings is prevented by the browsing of the animals or fire. In the long run, a rise in timberline will result in an increase in the land base for the timber industry, and a decrease in the zone where most snow avalanches are initiated (<u>see</u> section on 'Glacial processes').

GEOMORPHOLOGICAL PROCESSES

Geomorphological processes and climate

Geomorphological processes are dynamic actions or events that occur at the Earth's surface due to the application of natural forces. Such forces result from gravity, temperature

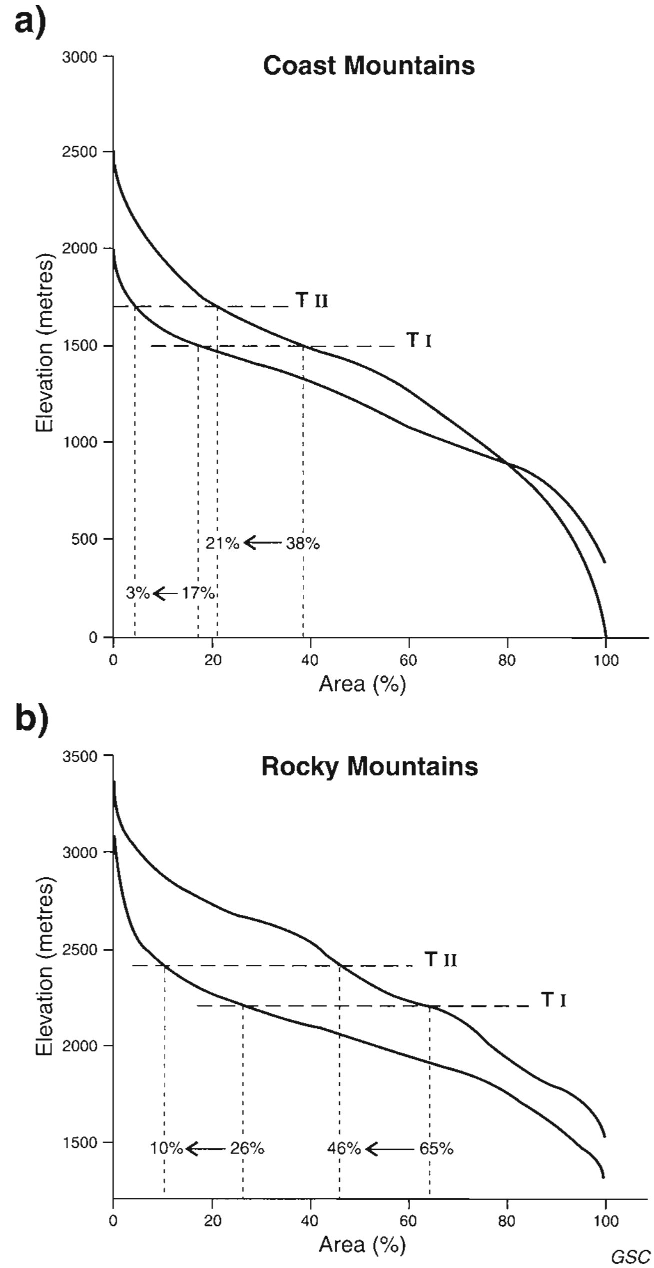


Figure 14. *Hypsographic curves for two drainage basins in the Coast Mountains **a)** and two drainage basins in the Rocky Mountains **b)** of southern British Columbia, showing the effects of potential timberline rise on the extent of alpine areas. Hypsographic curves are graphs that express the relation between proportional land area (horizontal axis) and altitude (vertical axis). Curves were constructed for two typical, small drainage basins in each area. In the Coast Mountain basins, alpine areas presently occupy 38 and 17% of the area. Timberline rise of 200 m (TI to TII) would reduce these areas to 21 and 3% respectively. In the Rocky Mountain basins, alpine areas presently occupy 65 and 26% of the area. Timberline rise of 200 m would reduce these areas to 46 and 10% respectively.*

changes, freezing and thawing, chemical reactions, seismic shaking, and the agencies of wind and moving water, ice, and snow. Where and when a force exceeds the strength of the earth material (rock, soil, ice), the material is changed by deformation, translocation, or chemical reactions.

The magnitude of the force that results from gravity depends upon slope steepness, and rock mineralogy determines chemical reactions, but the active controls on geomorphological processes are weather and climate. The climatic parameters that are commonly the most important in this regard are not the standard ones, such as monthly means of temperature and precipitation. Rather, the magnitude of exceptional events and the variability of temperature are significant. Thus for example, an unusually high-intensity rainstorm may generate more debris flows during a few hours than have occurred during the previous century. Also, the intensity of processes such as frost heave is much more closely related to the number of freeze-thaw cycles during a season than to mean monthly temperatures.

The resistance of earth materials to the forces that act upon them is a function of their inherent strength, internal cohesion, and interparticle friction. The strength of intact rock varies widely with rock type, but is generally greater than the strength of surficial materials (soils) such as glacial deposits. The strength of surficial materials depends on characteristics such as particle size, water content, bulk density, the presence or absence of ice as a cementing agent, and the extent and type of vegetation.

Vegetation is significant for several reasons. It increases slope stability and reduces erosion potential because interlocking roots increase bulk soil cohesion and thus strength. Leaves and stems absorb the impact of raindrops, and the vegetation mat slows surface water flow and, over a longer term, evaporation and transpiration from vegetation reduce runoff. Of particular significance in alpine (and arctic) areas is the way in which vegetation insulates the soil and influences its thermal regime.

The relations between climate and geomorphological processes are complex. Many geomorphological systems absorb and accumulate energy and matter (moisture and sediment) until some critical threshold value is exceeded, and then undergo a period of rapid change. Systems may be in stable or unstable states of equilibrium. In the former case, an external force effects only a temporary modification of the system, which eventually returns to its original state. A system in unstable equilibrium, however, would be permanently changed. Feedback loops are common within geomorphological systems, and feedback may enhance or delay the operative process. In addition, the history of a landscape is usually significant. Many geomorphic systems are still responding to the effects of the last Pleistocene glaciation which ended between 12 000 and 10 000 years ago.

Thus each geomorphological process that is operating within a system is a result of the complex response of earth materials to climate, local conditions (material type, vegetation, and slope steepness), past events, and the effects of other, contemporaneous processes. Each process has its own thresholds and response times.

Our knowledge of geomorphological processes, as described in the following sections of this report, is based on site monitoring, modelling and simulation, and studies of relict landforms related to paleoclimates. Numerous investigations have been carried out in Canada, and the results of work from alpine environments elsewhere are also applicable here. In general, processes have been well described and controlling factors identified in qualitative or semiquantitative terms, but there are few precise quantitative descriptions of the mechanisms involved.

For purposes of predicting changes in alpine geomorphology due to global warming, lack of quantitative data about processes is not a serious drawback because little quantitative information exists about predicted climate changes. Changes in climate parameters significant to geomorphological processes, such as the magnitude of extreme events and diurnal temperature fluctuations, have not been specified; regional patterns and time frame are only vaguely known.

As a consequence of considerations outlined in the above two paragraphs, this paper contains only qualitative suggestions as to the nature and direction of the response of geomorphological processes to climate warming and related effects.

Geomorphological processes in alpine areas

Some geomorphological processes, such as those associated with glaciers, are restricted to cold environments. Others, such as the effects of streams and wind, operate in a variety of environments, including the alpine zone. Processes that are restricted to cold and alpine environments are emphasized in this report; the discussion of other processes is relatively brief.

In this report, geomorphological processes are classified as follows:

Permafrost processes are related to the presence of continuously frozen ground, and to long-term (usually decades or longer) freezing or thawing of the ground;

Frost action processes are dominated by the effects of short-term (annual or less) freezing and thawing;

Nival processes result from the accumulation and melting of the annual snowpack, including snow avalanches;

Glacial processes are associated with permanent snow and ice;

Mass movement is the downslope movement of earth materials induced by gravity;

Stream processes include erosion and deposition by running water; and

Wind processes involve erosion and deposition by wind.

PERMAFROST PROCESSES

Introduction

The term *permafrost* refers to the thermal condition of ground that remains at zero or subzero (°C) temperatures for more than one year. Thus permafrost exists where the depth reached by winter freezing exceeds that of summer thaw. The shallow surface zone, known as the active layer, that thaws in summer and refreezes in autumn, ranges in thickness from a few centimetres to a few metres. In the arctic and the highest alpine areas permafrost is typically tens or hundreds of metres thick, and is continuous, except beneath deep lakes and rivers that do not freeze to their base in winter. Toward lower latitudes and elevations, permafrost thins and becomes patchy and discontinuous (Fig. 1). Near the margins of permafrost areas, the presence or absence of permafrost is difficult to ascertain because it commonly has no surface expression.

In Canada, the widespread permafrost in arctic and subarctic regions has received considerable attention from scientists and engineers. Permafrost in alpine areas is of relatively limited extent and has received little attention.

Ice is commonly present in permafrost. In coarse textured soils (sand, gravel, rubbly weathered rock) it usually fills only the pore spaces between adjacent particles, but finer materials (silt, clay) commonly contain segregated ice (distinct masses of ice) which may make up as much as 90% of the volume of the frozen ground. Melting of this ice-rich permafrost can result in erosion, landslides, ground subsidence, and other adverse effects. Coarse materials with interstitial ice and bedrock at subzero temperatures are the common types of permafrost in most alpine areas, and melting can contribute to slope failure.

The distribution of permafrost is controlled primarily by atmospheric temperatures, with the mean annual soil temperature usually being several degrees warmer than the mean annual air temperature. On the "Permafrost in Canada" map, Brown (1967) used the 17°F (-8°C) and 30°F (-1°C) mean annual isotherms to approximate the southern limits of continuous and discontinuous permafrost respectively. On a local scale, ground temperatures are strongly influenced by terrain and vegetation conditions. Permafrost is favoured by a relatively thin winter snow cover, thick peaty soils that are poor conductors of heat in summer when they are dry, presence of moss and vegetation cover (summer insulation), northerly slope aspect, and cold air drainage (downslope movement of cold air causing unusually cold temperatures). The thermal properties of soil and rock and local hydrological conditions are also important.

The distribution of frozen ground may be further complicated by the presence of relict permafrost. This is frozen ground that is not in thermal equilibrium with present conditions because it dates from a period of cooler climate. It warms and melts very slowly due to the insulation that is provided by overlying earth materials and vegetation. Depending upon the original temperature, thickness, and ice content of the permafrost, melting may take a year or two, or it may require millennia.

Distribution and character of permafrost in alpine areas

The distribution of permafrost in alpine areas is not well known because the extent and thickness of the frozen ground are exceedingly irregular. This patchy distribution results from extreme variability in the character of land surface and vegetation cover. In rugged terrain, air and ground temperatures, snowpack thickness, slope aspect and steepness, soil and rock types, vegetation characteristics, and moisture conditions can change markedly over distances of only a few metres (cf. Ives, 1974; Péwé, 1983). Permafrost limits are hard to map because the frozen ground is not usually a visible feature of the landscape. There are some distinctive landforms (Fig. 15), but these are not present on every patch of permafrost, and so its limits cannot be identified precisely by air photo interpretation (Brown, 1974). Subsurface investigations by probing or drilling are only feasible for small areas such as construction sites. Further, where permafrost is thin, and in response to short-term climate fluctuations, patches of frozen ground may melt and refreeze over time spans of just a few years. Deeper patches of relict permafrost may be present as well. The extent of permafrost is already gradually decreasing, and this is expected to continue, exacerbating the mapping problem.

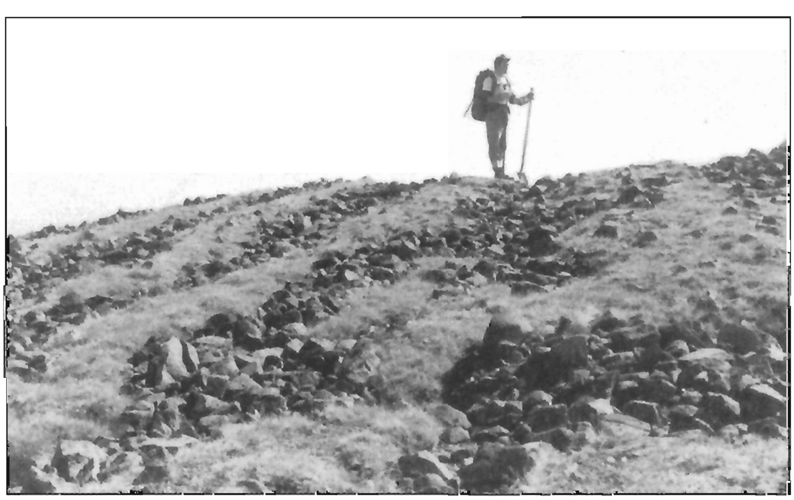


Figure 15.

Stone stripes – rows of boulders separated by turf bands – on Cathedral Ridge, southeastern Cascade Mountains, British Columbia. These features suggest that permafrost is (or has recently been) present here. GSC 1997-012L

Generalized elevation limits of permafrost in the eastern Cordillera (Rocky Mountains-Mackenzie Mountains) have been compiled by Harris and Brown (1982; also in Harris, 1986) on the basis of extensive field observations (Fig. 4a). The lower limit of continuous permafrost appears to decline northward from about 2200 m at 49°N in southern Alberta to 1500 m at 60°N, to about sea level at 67°N. Below this limit in the Alberta Rocky Mountains is a broad zone of discontinuous permafrost where the distribution of frozen ground is closely controlled by the terrain and vegetation factors mentioned above. Most discontinuous permafrost lies within the subalpine and montane forest zones however, because in these relatively dry regions, the lower limit of continuous permafrost roughly coincides with timberline.

Westward across British Columbia, the permafrost limit generally rises in response to warmer mean annual temperatures and increasing snowfall (Fig. 4b). Because timberline decreases westward, the alpine zone includes an increasingly broad band of permafrost-free terrain. Across each mountain range, permafrost limits probably ascend steeply from east to west in response to the increasing thickness of winter snowpacks. This effect has been mapped only in the Rocky Mountains (Harris and Brown, 1982), but it is evident elsewhere in the predominance of rock glaciers (glacier-like masses of rubble with cores of permafrost), rather than true glaciers on the drier, eastern side of each mountain range (Fig. 16). In the western Coast Mountains and the wetter parts of the Columbia Mountains, permafrost is relatively rare (cf. Mathews, 1955), although patches of frozen ground may exist at exposed, windswept locations where winter snow cover is thin or absent.

Permafrost is virtually continuous within the alpine areas of Yukon Territory (cf. Harris, 1983), and is hundreds of metres thick in the northern parts of the territory.

Permafrost is present throughout the alpine areas of Labrador and Quebec, where the limits of frozen ground correspond approximately with timberline (Ives, 1962; Brown, 1979). In central Labrador-Ungava, permafrost is 60 to 100 m thick below treeless ridge crests that are swept free of snow by strong winds (Granberg, 1973; Ives, 1974). The presence of permafrost at the summits of the Mealy Mountains, Labrador and the Otish Mountains, Quebec, is indicated by surface features (Brown, 1979). Thick permafrost in the Torngat Mountains probably developed when these mountains were ice-free during the last glaciation (Ives, 1979). No permafrost investigations have been published for Newfoundland, although permafrost is present on the higher ridges of the Long Range Mountains (Natural Resources Canada, 1995). Permafrost underlies the 1000-1100 m summit plateaus of the Chic-Chocs Mountains of Gaspésie, south of the Gulf of St. Lawrence. The frozen ground, which is in equilibrium with present-day temperatures, is about 60 m thick on Mount Jacques-Cartier, and its extent generally corresponds to that of alpine tundra (Gray and Brown, 1979).

Permafrost sensitivity to climate change

The sensitivity of frozen ground to climate change has been demonstrated by the response of arctic and low-elevation permafrost to past fluctuations in climate. For example, relict patterned ground, which is widespread in both alpine and arctic environments, is a clear indication of colder climates in the distant past. More recently, climate warming between the late 1800s and 1940s resulted in an increase in mean annual ground temperature of 3°C in the Mackenzie Valley; subsequently, and prior to 1975, cooling of 1°C occurred (Mackay, 1975). Mackay estimated that the 3°C warming may have brought about a 320 km (200 mile) northward shift of the southern boundary of continuous permafrost. This is roughly equivalent to an increase in elevation of about 500 m for the permafrost limit in mountainous terrain. More recently, several authors have reported recession of permafrost as a result of climate amelioration (see Williams and Smith, 1989, p. 80-81).

Recent studies of the potential effects on permafrost of climate warming, summarized in Williams and Smith (1989) and Smith (1990), have shown that, in general, the thermal

Figure 16.

A rock glacier on the eastern side of the southern Coast Mountains near Lytton, British Columbia. GSC 1997-012M

regime of permafrost will respond to changes in mean annual (atmospheric) temperature, but the response will be slow because undisturbed ground is buffered from the atmosphere by soil, peat, and vegetation. The response will also be complex because, for example, the nature of the vegetation itself may change slowly with changing climate. The precise thermal response at any given location will depend upon the characteristics of the natural surface and site conditions, especially wetness, including the ice content of permafrost at the base of the active layer, and snow cover (cf. Smith and Riseborough, 1983; Smith, 1986).

The predicted increase in the amount of winter snowcover further complicates the estimation of ground temperature response (Williams and Smith, 1989, p. 81). It has been shown that ground temperatures will increase in response to thickening of snow pack (Goodrich, 1982; Smith, 1990). Much more rapid changes in ground temperatures over a small area can result from direct human modification of soil, vegetation, and hydrological conditions, and from fire.

Smith (1990) indicated that in the case of continuous permafrost, where ground temperatures are many degrees below zero, predicted warming will not result in significant changes in the distribution of permafrost. The chief effect would be a thickening of the active layer in response to warmer summer temperatures. In areas of discontinuous permafrost, however, where much permafrost is warmer than -3°C, widespread disappearance of permafrost would occur under the predicted climate warming, although this would take many centuries (Williams and Smith, 1989, p. 80).

Effects of climate change and impact on human activities

Permafrost that is in thermal equilibrium with its environment is an essentially static feature, but permafrost degradation (melting) due to slow warming of the ground as a result of climate change or human disturbance gives rise to distinctive geomorphological processes. An increase in atmospheric temperatures causes warming of the ground from the top down. This may result in a thickening of the active layer only, or, where permafrost is relatively warm, it may cause thawing of the upper part of the permafrost below the depth reached by seasonal temperature fluctuations, or, where permafrost is warm and thin, it may eventually bring about its partial or total disappearance. Partial disappearance leaves lenses of relict permafrost below the active layer. The potential effects of these changes on the natural environment and related impacts on human activities depend largely on the ice content of the permafrost, and, to a lesser extent, on the morphology of the terrain, especially slope steepness.

Thawing of ice-rich permafrost releases water and reduces soil strength, resulting in local collapse and subsidence of the ground. This is the basic cause of most geotechnical problems in permafrost areas. In areas of low relief, this thaw-settlement produces irregular small hills and water-filled depressions, referred to as thermokarst terrain after similar features that result from the chemical dissolution of limestone. Thaw lakes and depressions may rapidly enlarge due to thermal erosion (i.e. melting of ice) by moving water in streams or by waves on lakes.

Widespread relict and active thermokarst features associated with ice-rich permafrost in the western Arctic have received much attention. Rampton (1973, 1982) concluded that thawing of ice-rich permafrost during a past period of warmer climate gave rise to the numerous lakes (thaw lakes) of the western Arctic Coastal Plain, and Mackay (1975) suggested that recently active thaw lakes in the upper Mackenzie Valley date from the late 1800s to 1940s warm period. Burn and Smith (1990) indicated that the development of thaw lakes near Mayo, Yukon Territory, may have been influenced by forest fires as well as by climate warming. They suggested that future climate warming could accelerate present rates of lake expansion.

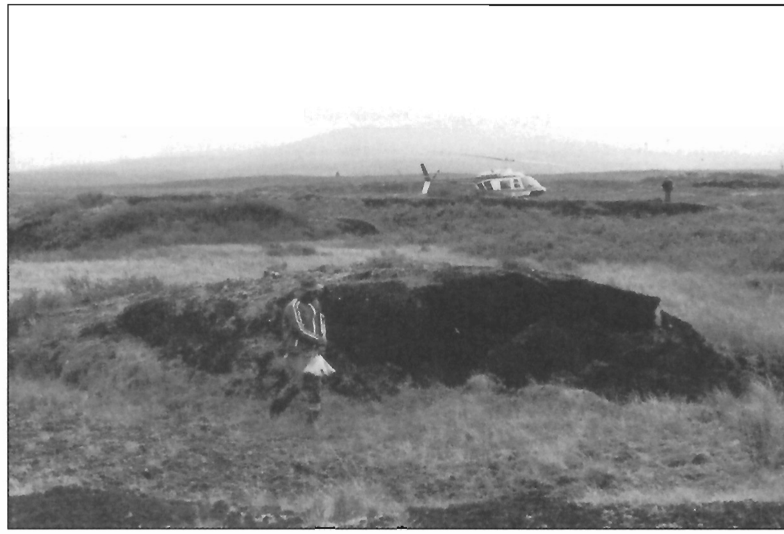
Thermokarst terrain is not widespread in alpine areas, except on debris-covered glaciers, due to the limited extent in the mountains of the thick silty soils that typically have a high content of ground ice. There are no published studies known to the writer of alpine thermokarst in Canada. Although highly generalized, the permafrost map of Canada (Natural Resources Canada, 1995) indicates that the ground ice content of mountain permafrost is very low, ranging from 0% to less than 10%. Small enclaves of ice-rich permafrost likely exist in valleys and depressions within the alpine zone where fine-textured soils, such as glacial lake sediments, are present, and climate warming could trigger thermokarst development at such sites. In undisturbed areas, thaw-settlement will probably occur slowly with no major environmental impact. Where roads or other developments have been built on ice-rich permafrost, thawing could be more rapid, causing local subsidence and erosion. The extent of the problems that arise will probably depend on the measures taken to protect the permafrost during the initial construction, such as insulation with gravel.

Another group of permafrost features that may be modified by thaw comprises plateaus and mounds (palsas) of peat. Palsas have been raised one or two metres above their surroundings due to the growth of ice lenses in the underlying peat (Fig. 17). They are widespread in bogs in the discontinuous permafrost zone, and are common near the latitudinal and elevation limits of permafrost. They are not permanent features, but pass through a cycle of formation and decay even in the absence of climate change (Seppälä, 1986), such that some mounds may be growing while others nearby are melting and collapsing. In some areas, the extent of these small landforms is declining (Kershaw and Gill, 1979). Climate warming, particularly warmer winters, and increasing snowfall will result in melting of the permafrost lenses and lead to a further decline in the extent of palsas in presently marginal permafrost areas. The effect of this on human activities is not detrimental, as roads and settlements generally avoid boggy ground, and reduction of permafrost will simplify construction should it be necessary.

Slope instability due to reduction in the strength of surface materials and addition of water is likely to be a much more widespread effect of permafrost degradation in alpine

Figure 17.

Palsas (peat plateaus) on the Stikine Plateau, north-central British Columbia. GSC 1997-012N



areas than thermokarst alone (cf. Smith, 1986). The significance of the resulting slope processes will depend upon the nature of the materials that are melting and their ice content. For example, thaw of ice-poor rock may only increase the frequency of rockfalls from cliffs where rockfalls already occur, but thaw of ice-rich materials may lead to major slope instability.

Landslides and debris flows caused by melting of ice-rich permafrost can be broadly divided into two types: shallow skin slides and flows, and deeper-seated slides, slumps, and associated debris flows.

Skin slides and flows occur where a thawing, relatively thin layer of water-saturated material becomes unstable, slides over the underlying frozen ground, and then flows rapidly downslope as a viscous fluid, rafting away surface soil and shallow-rooted vegetation. Initial failure can occur on slopes as gentle as 5° (Carson and Bovis, 1989). Downslope, where a flow comes to rest, soil and vegetation are buried by muddy debris. If the bare ground of the slide scar is underlain by icy permafrost, further melting and repeated failure may occur, enlarging the affected area and creating a scar that may take decades or longer to heal. This process occurs sporadically under present-day conditions, with the timing of slides influenced by factors such as the rate of thawing of the lower active layer (Lewkowicz, 1992). Thickening of the active layer and/or melting of the upper part of the permafrost due to climate warming would likely increase the frequency of such events and the extent of disturbed ground.

Skin slides and flows are related to the presence of ice-rich permafrost, and, as noted above, this is not extensive in most alpine areas, particularly in the more southerly mountains. Shallow failures in icy permafrost will be most likely where hillsides are underlain by predominantly silty materials, such as fine-textured, weakly consolidated or weathered sedimentary rocks, and silty glacial deposits.

Skin slides and flows will have a significant impact on any human activities in their path. Conversely, human disturbance of vegetation and terrain can accelerate this process, especially during a period of climate warming. Tundra fires are an additional contributing factor because these also result in loss of vegetation and soil cover. Landslide scars on mountainsides will be clearly visible from great distances, a negative visual impact in the opinion of many people.

Deep-seated landslides (e.g. thaw slumps), typically occur on slopes underlain by ice-rich permafrost and are commonly triggered by wave or stream erosion at the toe of stream-side bluffs (cf. French, 1976, p. 119-125; Burn and Lewkowicz, 1990). Mechanical and thermal erosion by the moving water leads to the exposure of ice lenses and icy sediments. The resulting melting triggers slumping and rapid recession of the bluffs, and commonly, the generation of semifluid debris and mudflows at the bluff toe. These processes are well known from arctic coastal lowlands, but as in the case of thermokarst, they likely affect only small areas in the alpine zone. Acceleration of these processes will likely accompany future warming of permafrost and surface waters.

Slope movements can also result from melting of interstitial ice in coarse-textured materials that underlie steep slopes, and from melting of ice veins in bedrock. For example, in continental mountains where unvegetated moraines recently exposed by glacier recession lie in the permafrost zone, the coarse, rubbly morainal material is held together by interstitial ice. Melting of this ice results in ravelling and slumping on steep moraines, and provides material for debris flows that can be generated by heavy rain (cf. Zimmermann and Haeberli, 1992). Similarly, melting of vein ice that fills cracks in bedrock can result in increased frequency and magnitude of rockfalls from cliffs.

A major engineering problem associated with permafrost is determination of the precise extent of frozen ground in the discontinuous permafrost zone. Construction on permafrost requires special measures to insulate and maintain the ground in its frozen state, and is normally more expensive than similar construction on unfrozen ground. Thus it is desirable to minimize construction costs by avoiding permafrost if possible and by applying special measures only where permafrost exists. Difficulties arise because, as mentioned earlier, permafrost limits are hard to map. There is, as well, the perverse prospect that where a building has been erected on permafrost, the permafrost may subsequently disappear, requiring the maintenance of an artificial climate in the soil that supports the building!

Williams and Smith (1989) pointed out that, even without thawing, warming of frozen ground results in a loss of strength because the deformation of ice occurs more readily at warmer temperatures. Thus, for example, warming could lead to a decrease in the bearing capacity of piles driven into frozen ground.

Summary

Most problems of construction on permafrost or environmental change due to permafrost melting are associated with frozen ground that has a high ice content. As far as is known, icy permafrost is not extensive in the mostly coarse-grained soils of the alpine zone, although it is likely present where Quaternary sediments or weakly consolidated sedimentary rocks give rise to fine-textured (predominantly silt or fine sand) soils. The effects of climate warming on permafrost are likely to occur slowly because the ground is insulated from the atmosphere by soil and vegetation. In general, the thermal regime of permafrost is more sensitive to human actions, such as removal of vegetation cover and tundra fires, than to climate change.

Long-term results of climate warming will likely include an increase in thickness of the active layer, melting of the upper part of permafrost below the base of the active layer, disappearance of permafrost in marginal areas, and a reduction in the strength of frozen ground due to warming. Surface expressions of these changes, which will be evident in some alpine areas, will include slow subsidence and development of small lakes (thermokarst), and an increase in the frequency of landslides and debris flows. The effects of climate warming and associated problems will be greatest in areas of discontinuous and thin permafrost.

FROST ACTION PROCESSES

A variety of geomorphological processes is associated with the short term freeze-thaw cycles that occur daily or seasonally. The processes result from stresses that are generated by the freezing and thawing of moisture in the pores and voids of soils and sediments, and in the pores and cracks of rocks. Results include shattering and disintegration of bedrock, and heaving, churning, and downslope movement of the active layer. Because most of these processes do not pose serious constraints on human use of alpine areas, the following discussion is relatively brief.

Frost shattering

Frost shattering of bedrock is ubiquitous in alpine areas. This weathering process widens cracks in bedrock and liberates angular rock fragments. Frost shattering on level to moderately sloping ground results in the development of blockfields

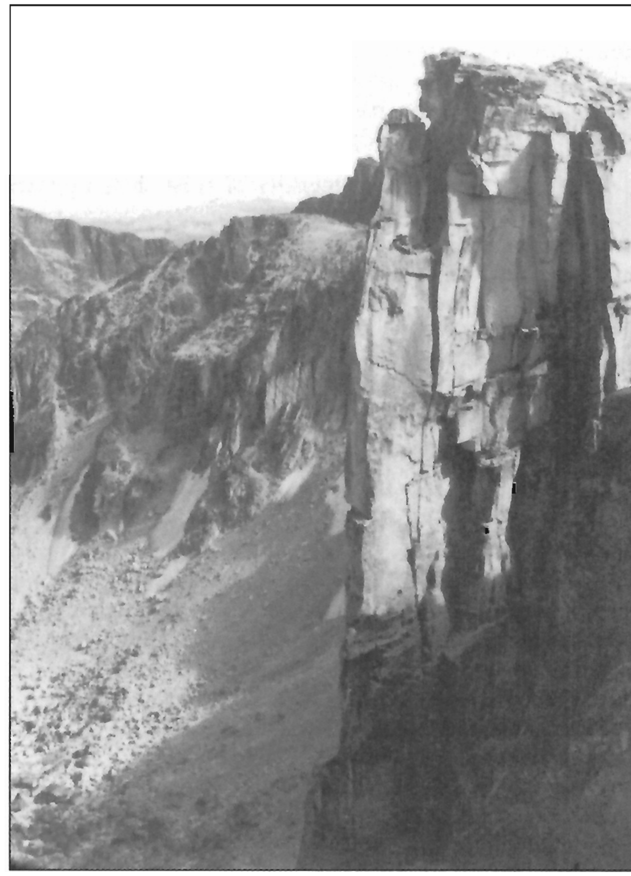


Figure 18. *Frost effects lead to the widening of joints (vertical cracks in this example) and rockfalls, resulting in the accumulation of talus below rock faces. (Smoky the Bear, southeastern Cascade Mountains, British Columbia). GSC 1997-012*

(felsenmeer) and veneers of rubbly detritus on mountain tops, and slopes covered with rocky debris. The release of rock fragments from bedrock cliffs results in rockfalls, the only hazardous effect of frost action (Fig. 18).

Frost shattering has been loosely attributed to stress generated by the freezing and expansion of water in bedrock crevasses and pore spaces. The specific mechanism whereby this occurs, however, is not fully understood (cf. White, 1976) and is currently under investigation. The important climatic parameter that influences the intensity (rate and magnitude) of the process is, in the traditional view, the frequency of freeze-thaw cycles within rock crevasses and pores. Recent work (e.g. Walder and Hallett, 1985; Hall, 1986), however, suggests that frost intensity, which is the temperature depression below 0°C, and the duration of subzero temperatures (Gardner, 1970; Church et al., 1979), may be more significant. Obviously, the presence of moisture is essential.

Regional variations in the intensity of frost shattering in alpine areas have not been investigated. If frost intensity is the most significant parameter, then frost shattering will be more effective in mountains with relatively cold winters, such as the northern Rocky Mountains and the Torngat Mountains, and least effective in maritime mountains with mild winters, such as the west side of the southern Coast Mountains. Thick snow packs also reduce frost intensity on all but steep, snow-free slopes in maritime areas.

Local variations in the intensity of frost shattering are controlled by the size and spacing of openings in bedrock and microclimate, which is determined by slope aspect, steepness, exposure, soil, vegetation, and upslope sources of water such as snow patches. These factors give rise to considerable variation in the intensity of frost shattering over short distances in rugged terrain. Although such variation has not been measured directly, a related phenomenon, rockfall frequency, has been observed to vary markedly between closely spaced sites in response to local factors (e.g. Luckman, 1976).

Given our present lack of knowledge about the relation between frost shattering and climate and about the details of future climate warming, it is pointless and potentially misleading to try to predict future changes in frost shattering intensity. Climate warming (warmer winters and drier conditions) might result in reduced frost intensity in some areas but, in cold regions, warmer summers might promote frost shattering by increasing thaw. In any case, the magnitudes of these changes may be insignificant given the local variability of the frost shattering process.

Frost heaving and cryoturbation

Frost heave results from the stress exerted by the growth of ice crystals. For example, upheaval of surface soil and pebbles due to the overnight growth of ice needles associated with ground frost is a common phenomenon, even in urban areas. Cryoturbation probably results from differential rates of freezing and heaving and differential rates of thawing and settling that are induced by variations in soil characteristics, such as particle size and moisture content. The precise mechanics of these processes are not known however (see discussion in Williams and Smith, 1989), and variations in the intensity of these processes within alpine areas have not been described. As in the case of frost shattering, freeze-thaw processes are probably most effective in mountains with continental and subpolar climates where snowfall is relatively light.

Heaving and churning (cryoturbation) of soils due to annual freezing and thawing result in the development of a variety of soil structures (see excellent review in Washburn, 1979); they also influence some human actions. Frost-related soil structures are common in both alpine and polar regions, and include contorted and convoluted soil horizons, earth hummocks, and some kinds of patterned ground. Stones are brought up to the ground surface by frost action, and large chunks of bedrock can be heaved upward. Piles, telephone poles, and other emplaced objects are also subject to frost heave, and roads can be broken up by the seasonal growth of subsurface ice lenses. Damage can be minimized or prevented by the use of appropriate materials and designs.

In general, frost heaving and churning are a constraint to some forms of construction in alpine areas, but one that can be overcome by geotechnical measures. Climate warming could reduce frost effects in relatively warm areas, but potential effects cannot be predicted in any meaningful way at present.

Solifluction

This is the slow downslope movement of soil, usually affecting a layer that is less than about 2 m thick (see Washburn, 1979, for a comprehensive description). Soil movement occurs during freezing and thawing, and possibly also during wetting of the soil from melting ice. Total movement is probably the cumulative effect of numerous small displacements (see discussion in Williams and Smith, 1989). Rates of downslope movement are dependent upon local factors, such as slope steepness, aspect, and vegetation (cf., Price, 1970), rather than regional conditions. Rates of a few millimetres to a few centimetres per year have been indicated by measurements from several locations (cf. Price, 1970; Smith, 1987), but a relation between rate of movement and soil or air temperature has not been recognized. No quantitative regional

Figure 19.

The leading edge of a series of solifluction lobes, southern Okanagan Range, Cascade Mountains, British Columbia. GSC 1997-012P comparisons have been made, although Thompson (1990) suggested that gelifluction (solifluction over permafrost) may be more active on mountains in continental and subarctic climates. Movement occurs on virtually all soil-covered surfaces with gradients of more than a few degrees. The soil may move as a featureless blanket, but most commonly, distinctive lobes or terraces are present (Fig. 19).

As in the case of other frost effects, this process is a constraint to some forms of development in alpine areas, but it can be accommodated by appropriate geotechnical measures. Climate warming and decreases in soil moisture may reduce solifluction activity in relatively warm areas. However, more specific predictions are unwarranted at present and local variation may be more significant than any changes resulting from climatic warming.

Summary

The potential effects of climate change on frost action cannot be reliably predicted at present due to incomplete understanding of the mechanics of these complex processes, the role of regional versus local conditions, and imprecise knowledge of climate change. With regard to all frost action processes, the effects of climate change likely will be minor compared to the local variability of these processes that results from the differences in influential factors such as slope aspect and steepness, vegetation cover, soil moisture, and snow depth and duration.

THE WINTER SNOWPACK, SNOW AVALANCHES, AND NIVATION

Snowpack

Regional patterns of snow accumulation in alpine areas are directly related to snowfall amounts shown for the climatic regions in Figure 1. Where snowfall is heavy, such as in the Coast Mountains and Columbia Mountains, the snowpack commonly attains a thickness of several metres by the end of winter. In drier continental regions such as the Rocky Mountains, the average snowpack is thinner, but deep drifts develop at protected sites where wind-blown snow accumulates.

Anticipated climate modifications include two changes that can have opposite effects on the snowpack, and so the following suggestions about their combined impact are very uncertain. Increasing winter temperatures suggest decreasing snowfall amounts and snowpack thicknesses, but conversely, increasing winter precipitation suggests increases in snowfall and snowpack. Although the effect of winter warming is expected to predominate, resulting in a decrease in the extent of seasonal snowcover (Mitchell et al., 1990; Government of Canada, 1994), there could be significant departures from this norm in some regions. In particular, contrasts between warmer and colder regions could be enhanced.

In southerly mountain regions, where the winter snowline is normally above sea level, the depth of the winter snowpack increases upslope because more snow falls and less melting occurs at progressively higher elevations. The rate at which snow depth increases upslope depends on factors such as the altitudinal variability of the freezing level throughout the winter and the total amount of winter precipitation. With increased winter temperatures and increased snowfall, there will be less snow at low elevations and snowline will rise. Yet at higher elevations that remain above the freezing level (i.e. at subzero temperatures) through most of the winter, snowfall will increase. Consequently, the rate at which snow thickens upslope will increase. In colder areas where winter temperatures will remain below zero despite several degrees of warming, there may be a more uniform increase of snowfall and snowpack thickness.

Other significant modifications of snow characteristics may accompany climate change. Snow temperatures will be affected, along with related properties such as snow particle size and shape, cohesion, and the internal structure of the snowpack (see 'Snow avalanches' for effect on avalanches).

The magnitude of the anticipated changes in the thickness, extent and thermal characteristics of the snowpack cannot be estimated at present, but it is possible that these effects will have adverse impacts on some snow-based activities. Ski resorts and ski runs at relatively low elevations could suffer from a decrease in the average duration of snow-cover, a tendency toward warmer and wetter snow conditions, and increased rain. Presently reliable ski areas may become only marginally viable, and presently marginal resorts may be eliminated, resulting in large financial losses. Other forms of snow-based recreation and commercial activities that depend upon the snow pack (and low temperatures), such as winter haul roads, will be similarly affected. On the other hand, reductions in snowfall and snowpack will benefit some activities in the alpine zone due to increased accessibility and reduced snow removal costs.

The seasonal snowpack is also the natural reservoir that provides the spring and summer water supply to many areas that experience summer drought, such as British Columbia's Okanagan Valley. It also provides water for hydropower generation. Changes in snowpack thickness, temperature, and water-content, together with increased winter runoff, will influence river discharge regimes and hence water availability for agriculture and industry. Figure 1 shows rivers that are fed by alpine meltwater.

Snow avalanches

Introduction

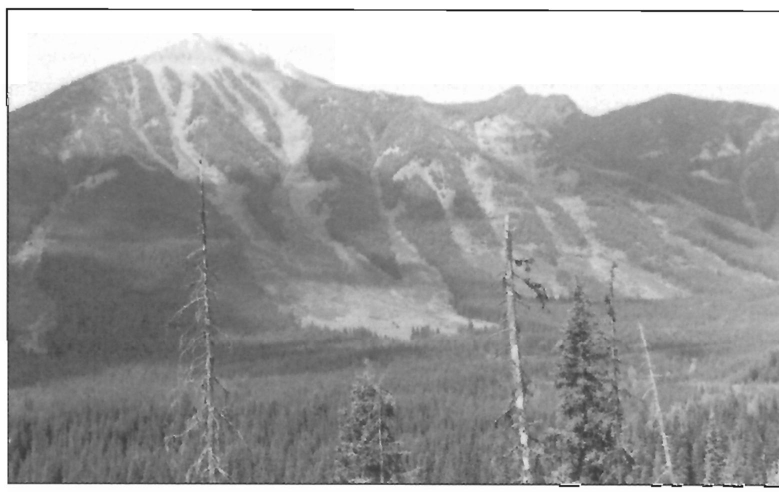
The rapid downslope movement of snow is one of the most dangerous and destructive natural processes. Avalanche hazards exist along many highways and railroads in western Canada. Along the most dangerous sections, for example at Rogers Pass, where the Trans-Canada Highway intersects more than 140 avalanche paths (Fig. 20), danger zones are well known and the slopes above are closely monitored. Although a large number of sites that are threatened by avalanches lie in the montane forest zone, many of the avalanches start in the treeless alpine zone (Fig. 21). In alpine areas, avalanches are common on moderate to steep slopes. They pose a serious hazard to recreational skiers and

Figure 20.

Avalanche at Rogers Pass, Selkirk Mountains, British Columbia. Photograph courtesy of D. McClung.

Figure 21.

Avalanche tracks in summer, Elk Valley, southern Rocky Mountains, British Columbia. Note that most tracks begin on steep, treeless slopes. Although the tracks are clearly defined by contrasts in vegetation types, they do not provide a totally reliable indicator of the extent of hazard because large avalanches will occasionally crash into mature forest. GSC 1997-012Q



snowmobilers that is increasing, regardless of climate change, as larger numbers of inexperienced people gain access to snow-covered terrain.

Characteristics and causes

On forested slopes, most of the paths followed by avalanches are clearly marked by avalanche tracks (Fig. 21) – strips of bright green shrubby vegetation where tree growth is suppressed. Occasionally, an avalanche will crash into mature forest, but the distribution of avalanche tracks shows where avalanches regularly occur and thus where the highest hazard exists. In alpine areas, avalanches are more widespread; some avalanche paths are indicated by debris deposits, but others are not well marked, and thus hazardous areas are harder to define.

Many avalanche types have been defined according to modes of initiation and movement. For example, loose snow avalanches result from surface failure at a point, whereas slab avalanches (Fig. 2a) result from detachment of cohesive blocks of snow. Snow moves downslope, either by sliding or by flowing, with the latter process most common in avalanches. Flow mechanisms, velocity, and the destructive power of an avalanche are strongly influenced by the characteristics of the snow, the amount of snow in motion, and terrain characteristics such as surface roughness and the length and shape of the avalanche path. Perla and Martinelli (1976) and McClung and Schaerer (1993) provided comprehensive discussions of avalanche processes. Some kinds of avalanches commonly transport vegetative debris and rock fragments, increasing their destructive power and resulting in debris cones at the lower end of avalanche paths (cf. Luckman, 1978).

In general, avalanches are favoured by heavy snowfall (see table in Fig. 1 for regions of heavy snowfall), but a variety of weather conditions and local landscape characteristics contribute to snowpack instability. Failure of a slab, the most destructive kind of avalanche, occurs when the shear stress exerted by the snowpack (the downslope component of snow weight) exceeds the shear strength of a weak layer at the base of the slab. Thus the amount of snow that falls during each storm is a major control on avalanche occurrence. The shear strength of the snowpack increases with snow density, decreases with a rise in temperature, and varies in a complex way with snow texture (grain size and shape) and freeze-thaw history. Snow density, temperature, and texture are not constant but change with time under the influence of atmospheric temperature, wind, additions of new snow or water from surface melt, and other factors. Thus the probability of an avalanche event on a particular slope varies with time and weather conditions. In continental areas, for example, persistent cold weather sets up a steep temperature gradient in the snowpack, resulting in the upward migration of water vapour away from the base of the snowpack where large, poorly interconnected and noncohesive ice crystals (depth hoar) grow by sublimation. The shear strength of this zone is relatively low and if it fails, large avalanches can result.

On any particular mountain or valley side, some slopes are more prone to avalanches than others. Factors such as wind, slope aspect, temperature and snow characteristics (i.e. powder snow vs. wet snow), slope steepness, terrain roughness, characteristic drifting patterns, and the vegetation of the potential avalanche initiation zones are all important in this regard (Schaerer, 1977). For example, the amount of snow deposited during a storm varies greatly from place to place depending upon wind strength and direction and snow type. Snow drifts from which avalanches may originate accumulate on lee slopes and in hollows, whereas deposition on windward slopes may be minimal. Particularly exposed sites may be swept bare and all snow transported to drifts on leeward sites if the snow is dry. Once the snow is deposited, rough and irregular slopes provide support for a snowpack which would slide off smoother terrain, and, of course, trees usually function as anchors.

As already noted, many major avalanche tracks that run down steep, forested mountainsides start at initiation zones that lie in the alpine zone just above timberline. Grassy meadows and glacially smoothed rock slabs promote avalanches that are initiated by failure at the base of the snowpack.

Effect of climate change and impact on human activities

In general, the magnitude and frequency of avalanches might be expected to increase in areas where snowfall increases, particularly if the amount of snow deposited during individual storms also increases. The converse effect may occur when and where snowfall declines. These simple relations are made more complex, however, by the additional effects of changes in snowpack temperature regimes and patterns of snow accumulation. In warmer winters, for example, newly fallen snow may stabilize and reduced temperature gradients may decrease the likelihood of depth hoar, resulting in fewer avalanches. Shifts in prevailing wind directions and storm paths may lead to changes in the patterns of snow loading on slopes. In addition, tree colonization of some avalanche initiation zones may stabilize the snowpack.

If avalanche magnitude were to increase due to climate change, or if avalanches were to begin to occur on slopes not previously affected due to changes in snow accumulation patterns, larger and new runout zones would develop in forested zones, and old tracks would become increasingly unreliable indicators of the extent of avalanche hazards.

In summary, climate change could cause changes in the magnitude, frequency, and locations of snow avalanches, and hence the potential avalanche hazard to human activities. Avalanche hazards will probably decline in some areas, but may increase in alpine areas that lie well above the winter snowline. Thus avalanche danger may increase at ski resorts located in the alpine zone.

Nivation

Summer snow patches in alpine areas occupy shallow depressions that are formed by a variety of processes related to the prolonged presence of snow, such as freezing and thawing of snow meltwater, corrosion by meltwater, and slow creep of saturated overburden. These processes are referred to collectively as *nivation* (Washburn, 1979). After snow patches have melted, nivation hollows, which are typically a few metres deep and a few tens of metres across, can be identified by their lack of vegetation or by the presence of late-developing herbaceous vegetation, as well as by their morphology (Fig. 22).

In the event of a decline in snowfall and snowpack thicknesses, snow patches will be shorter lived, and the effects of nivation less conspicuous. Sparsely vegetated nivation hollows will probably be colonized by plants from adjacent areas. If snowfall were to increase, the extent of snow patches and areas affected by nivation would also increase, possibly resulting in a dying-off of vegetation unable to survive under more prolonged snow cover.

Such changes would occur slowly and would have relatively little impact on human activities in the alpine zone.

GLACIAL PROCESSES

Introduction

Glaciers, together with the landforms and sediments that result from glacial erosion and deposition, are, to many people, the most interesting and significant features of the alpine environment. Glacier expansion and contraction is one of the best known manifestations of climate change. The effects of the great Pleistocene glaciations, when ice expanded to cover most of Canada and other large areas in high latitudes, are particularly evident in alpine areas, as are the moraines and abraded rocks that date from the recent glacier advances of the Little Ice Age (Fig. 23).

A glacier is a perennial accumulation of snow and ice that flows slowly downslope. It begins to develop where and when mean annual snow accumulation exceeds snow removal by ablation (melting and evaporation). As snow depth increases, the lower layers of the snowpack are slowly transformed to ice, and eventually, the accumulated mass imparts sufficient stress to deform the lower ice layers and to cause the basal ice to slide over its bed. Then the glacier expands downslope until its advance is halted by increased ablation at lower elevations.

Figure 22.

Nivation hollows in late summer, Three Brothers Mountain, Cascade Mountains, British Columbia. The shallow depressions where turf is absent have been recently vacated by melting snow. GSC 1997-012R

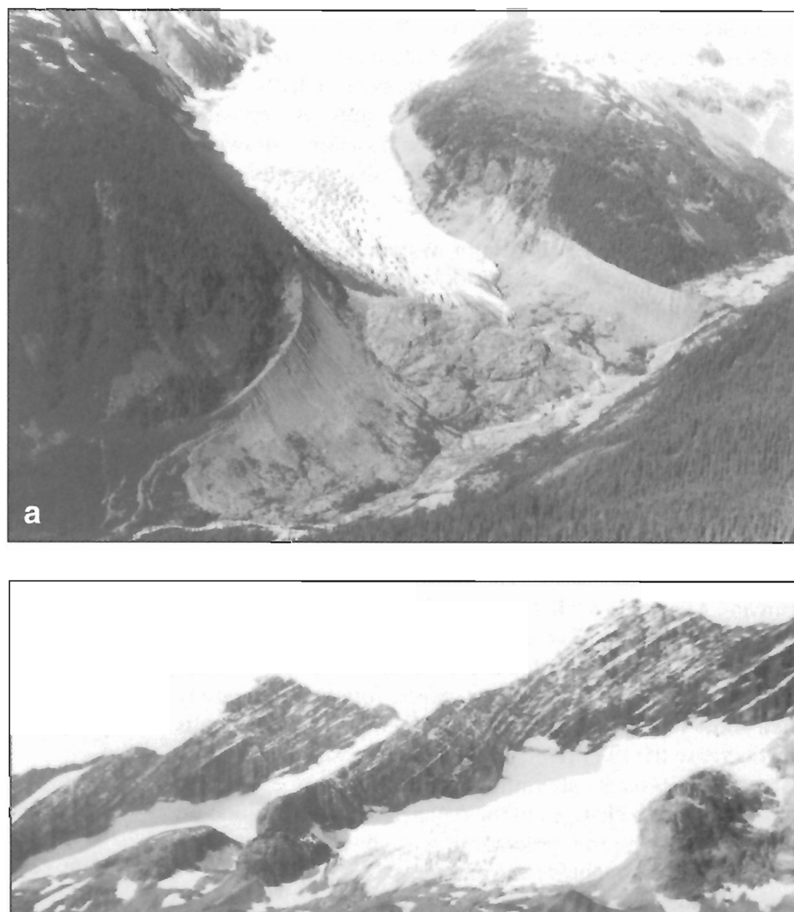


Figure 23.

***a)** Glacier with prominent Little Ice Age moraines near the head of Bute Inlet, southern Coast Mountains, British Columbia. GSC 1997-012S **b)** Small glaciers with Little Ice Age moraines near the Elk Valley, southern Rocky Mountains, British Columbia. Small glaciers like these may disappear. GSC 1997-012T*

Glacier mass balance and relation to climate

The response of a glacier to climate is determined by the amounts of snow and ice that accumulate and ablate each year. In relatively dry continental mountains, where annual snowfall is light, the volume of snow and ice that is lost each year by ablation is similarly small. In maritime mountains, where winter snow accumulation of several metres is common, equivalently large volumes are lost to ablation, mostly to meltwater runoff. Large maritime glaciers commonly extend downvalley to altitudinal zones that are forested. Although the lower parts of these ice tongues are below timberline and not in the alpine zone as defined earlier in this report, they are included in the discussion of glacier hazards that follows.

More significant, however, than total accumulation and ablation with respect to climate change is the mass balance of a glacier; that is, the marginal amount by which accumulation exceeds ablation or vice versa. The magnitudes of accumulation and ablation are clearly controlled by climate, but the relations are not simple. For example, climate in the immediate vicinity of a glacier may differ significantly from regional climate and the mass and energy exchange processes at a glacier surface are commonly complex. Mass can be added by a variety of processes, including sublimation (hoar frost), freezing of rainwater, snow drifting and avalanches, as well as direct snowfall. Ablation is determined by incoming solar energy, the reflectivity of the glacier surface, wind speeds, and atmospheric humidity, as well as air temperature. (For a full discussion of the relations between glaciers and climate, see Meier, 1965; for glacier physics, see Paterson, 1994.) In general, winter precipitation (snowfall) and summer energy inputs (indexed by temperature) are the major controlling factors on mass balance in most glacial environments.

Changes in mass balance over time determine the response of a glacier to climate change. If, on average, a positive mass balance is maintained, then the glacier will advance: its volume will increase, the ice will thicken and expand downvalley. A negative mass balance will result in glacier recession: the glacier terminus will melt and retreat up valley. The effects of precipitation and temperature are hard to separate: a positive mass balance can result from either snowier winters or cooler summers, or both; a negative mass balance expresses either drier winters or warmer summers, or both.

The annual budget of a glacier is rarely balanced (i.e. mass balance = 0), and glacier termini are rarely stable. Advance and recession of alpine glaciers in response to climate cycles of a few years to centuries or millennia in duration is the normal state of affairs. The response of a glacier to climate-induced changes in mass balance commonly lags behind the climate change. Advance after an increase in mass balance usually lags by a few years for alpine glaciers (Paterson, 1994). Retreat after a decrease in mass balance induced by warm summers is usually immediate, as most melting occurs near the glacier snout, which retreats accordingly. Differences in glacier size, microclimate, and morphology commonly result in asynchronous behaviour, even between adjacent glaciers (e.g. McCarthy and Smith, 1994).

At the present time, alpine glaciers throughout the world abut sparsely vegetated zones that have been uncovered by glacier recession only since the culmination of Little Ice Age advances in the eighteenth and nineteenth centuries (Fig. 23). This recession has occurred at variable rates, interrupted at times by pauses and minor readvances in response to climate fluctuations that have been recorded by meteorological instruments. For example, Mathews (1951) correlated rapid glacier shrinkage in Garibaldi Park, southern Coast Mountains, since the second and third decades of this century (and before 1950) with warmer temperatures recorded at New Westminster, British Columbia. Luckman et al. (1987) attributed glacier advances in the Premier Range of the Cariboo Mountains during the mid 1950s to mid 1980s to increased winter precipitation and low summer temperatures between 1951 and 1976, recorded at Valemount, British Columbia. Field (1975) summarized the recent behaviour of many Canadian glaciers, and Brugman (1992) described trends in mass balance data and glacier snout positions observed over the past few decades.

Few quantitative data are available for glacier shrinkage in Canada following the Little Ice Age. Broad regional comparisons are not possible, but some examples are available. Mathews (1951) estimated that between 1860 and 1947, the area of Helm Glacier, Garibaldi Park, decreased by 39%, and between 1911 and 1947, glaciers in the park thinned at rates ranging from 1.4 to 3.8 m/a. Harding (1985) and Luckman (1990) reported that between the mid-nineteenth century and 1970, while average global temperature rose by 0.5°C, ice cover in part of the Premier Range (northern Columbia Mountains) declined by 23% and the average elevation of glacier snouts rose by 262 m. Brugman (1992) reported that ice volumes in monitored glaciers have decreased by 10 to 50% during the last 30 years. McCarthy and Smith (1994) showed that the area covered by most glaciers in part of the Rocky Mountains of southwestern Alberta decreased by 30 to 50% (total range 15-80%) between 1916 and 1988. While amounts of recession and volume loss can be expected to vary considerably with differences in glacier regimes, the data presented here do indicate the magnitude of the changes that resulted from a climate amelioration that is small compared to that predicted for next century.

How much of the post-Little Ice Age recession is the result of natural climate amelioration and how much is the result of the atmospheric modification that has already occurred since the industrial revolution has not been determined.

Present day distribution and extent of alpine glaciers

The extent of alpine glaciers and icefields is indicated in a generalized manner in Figure 1. Small-scale maps showing individual glaciers in western Canada are available (Falconer et al., 1965; Henoch and Stanley, 1967a, b; Field, 1975). The glaciers are described by Field (1975); glaciers in western Yukon Territory and northwestern British Columbia are also described in a more recent account by Field (1990).

The largest glaciers and alpine areas with the highest proportion of snow and ice are in the higher and more northerly parts of the maritime mountains of western Canada. In the northern Coast Mountains of British Columbia and the St. Elias Mountains of Yukon Territory, glaciers and icefields are more extensive than ice-free alpine terrain. Smaller icefields and valley glaciers also exist in the moister ranges of central and southern British Columbia, and in Alberta near the Columbia Icefields. Glaciers are absent from cold, but dry subpolar mountains: the Canyon Ranges of the Mackenzie Mountains and the British and Richardson mountains. These areas also remained largely ice free during Pleistocene glaciations (Duk-Rodkin and Hughes, 1992). In the Torngat Mountains of eastern Canada, about 50 small cirque glaciers comprise about 5 km^2 of ice (Fahn, 1975).

Glacier elevations vary from region to region in accord with present-day climate. The most significant, visible, climatically controlled feature on a glacier is its equilibrium line, the late-summer snowline that separates the permanent snowfields of the upper part of the glacier from the exposed ice of the lower part of the glacier. It is the elevation on the glacier where accumulation equals ablation. Glaciers with similar climate usually have a similar ELA (equilibrium line altitude), even though they may differ markedly in size. ELAs rise eastward across British Columbia and southern Alberta, roughly paralleling timberline, as a consequence of decreasing snowfall and increasing summer temperatures (cf. Evans, 1990). Northward, ELAs decline in elevation as a result of generally cooler temperatures. Ostrem (1966, 1972) determined that the glaciation limit (the critical height that has to be exceeded if a mountain is to bear a glacier; about 100 m above ELA) rises from 1600 m to 3000 m across southern British Columbia and Alberta, and from 1200 m to 2700 m across northern British Columbia (Fig. 3).

Effects of future climate change

Climate warming, particularly increasing summer temperatures, is expected to continue the general 20th century pattern of rising ELAs, predominantly negative glacier mass balance, and glacier recession. Increasing winter precipitation will tend to offset this trend in some areas, however, but may be insufficient to reverse it. Hence future glacier behaviour is uncertain, although recession is considered to be most likely. Glacier response may vary between maritime and continental mountains and between middle and high latitudes (Haeberli, 1990). For example, because gross accumulation and ablation amounts are greatest in maritime mountains, changes in mass balance may be greatest there, and so maritime glaciers may be more sensitive to climate change. If increased snowfall in cold regions is not offset by warmer summer temperatures, ELAs could decline, causing glaciers to advance. This latter effect, which has not been seriously considered in the literature, is probably unlikely and is not discussed in detail here.

Insufficient data about climate change exist for reliable determination of the magnitude of glacier recession or ELA rise. If temperature alone were to change, a very crude estimate of ELA rise could be made by applying the standard lapse rate of -0.6°C/100 m. Thus for every degree of warming, snowline might be expected to rise by about 170 m (for 2.5°C warming, 425 m). During attempts to estimate changes in sea level resulting from glacier melt, Kuhn (1985) calculated that a temperature increase of 4°C over 100 years, together with a minor increase in precipitation, would lead to a rise of ELAs of alpine glaciers by 265 m, or a rate of rise of 2 to 3 m per year; rates of glacier melting would increase to a maximum within a few decades. Although it is unlikely that any of these predictions are accurate, they indicate the direction, and possibly the order of magnitude, of the changes that will occur.

A rise of snowline of a few hundred metres would result in a drastic reduction in the extent of glaciers. In more southerly mountains, where present snowline is commonly less than a few hundred metres below the mountain summits, many glaciers would disappear (Fig. 23b). Farther north, smaller glaciers at low elevations would disappear and larger, higher icefields and glaciers would experience massive recession. Extensive areas of bare ground would be exposed by glacier retreat and eventually colonized by tundra vegetation or forest.

Impact of glacier recession on human activities

Impact on tourism and outdoor recreation

Glaciers and snowfields are a significant component of mountain scenery for thousands of visitors, many of whom may never leave the vicinity of their vehicles. Apparently barren glacier forefields, recently emerged from beneath the ice, are considerably less attractive. Thus major glacial recession, accompanied by significant changes in scenery, could have a marked impact on tourism in popular mountain areas, such as the southern Rocky Mountains (see frontispiece).

Visitors can enjoy close contact with the ice at the Athabasca Glacier of the Columbia Icefields (Sanguin and Gill, 1990). In the past, they enjoyed easy access to the Illecillewaet (Great) Glacier near Rogers Pass. Just as the snout of the Illecillewaet Glacier has receded during this century to a location that is now accessible only to mountaineers, so the Athabasca Glacier may recede over the next several decades (see frontispiece). Mountaineers also will notice the effects of glacier recession. Replacement of glacier ice by smooth, glacially abraded rock slabs could significantly alter the character of many climbing routes, and either increase or decrease their difficulty.

Hazards resulting from glacier recession

Glacier-related hazards will continue to threaten human activities in, and downstream from, many alpine areas, and they will probably be amplified by the effects of rapid glacier recession. At the same time, increasing use of mountain valleys and alpine areas for recreational and industrial purposes will bring many more people into potential hazard zones.

Hazardous events that are associated with present-day glaciers and that may become more common as glaciers recede are catastrophic floods and debris flows from ice-dammed and moraine-dammed lakes, landslides and debris flows initiated on recently deglaciated, unstable slopes, and ice avalanches (see Eisbacher and Clague, 1984, and Evans and Clague, 1992 for reviews of these processes). Glacier surges could result from warming of basal ice. These hazards are discussed below. Significant changes in the hydrology and sediment load of glacial meltwater streams that could result from glacier melting and shrinkage are discussed separately.

Outburst floods from glacier-dammed lakes

Trunk glaciers commonly impound lakes in ice-free tributary valleys, and a tributary glacier may impound a lake if it enters a trunk valley. In the latter case, a relatively large lake with great flood potential may be formed (Fig. 24). Lakes can also be held in embayments between two glaciers, between a glacier and the valley sides, and within, under, or on top of a glacier. New ice dams and new lakes are formed during both glacier advance and recession according to changes in glacier extent and configuration.

Ice dams are prone to sudden, catastrophic leakage, resulting in floods that can exceed normal snowmelt or rain-induced floods by more than an order of magnitude. Such floods are often referred to by the Icelandic word *jökulhlaup* (glacier flood) because they are relatively common in that country. Catastrophic drainage usually occurs in late summer when the water level in the impounded lake is relatively high. Tunnels through or under the ice dam enlarge rapidly due to several possible mechanisms, such as flotation of the ice, high water pressures causing ice deformation and tunnel enlargement, and melting due to heat energy supplied by the escaping lake water (see discussion and references in Young, 1977). In some cases, flotation and wholesale collapse of the ice dam may occur (Clague, 1987).

Outburst floods differ from normal floods in that discharge increases rapidly to a peak, and then ceases abruptly. After drainage, the ice dam reforms or reseals, and the lake

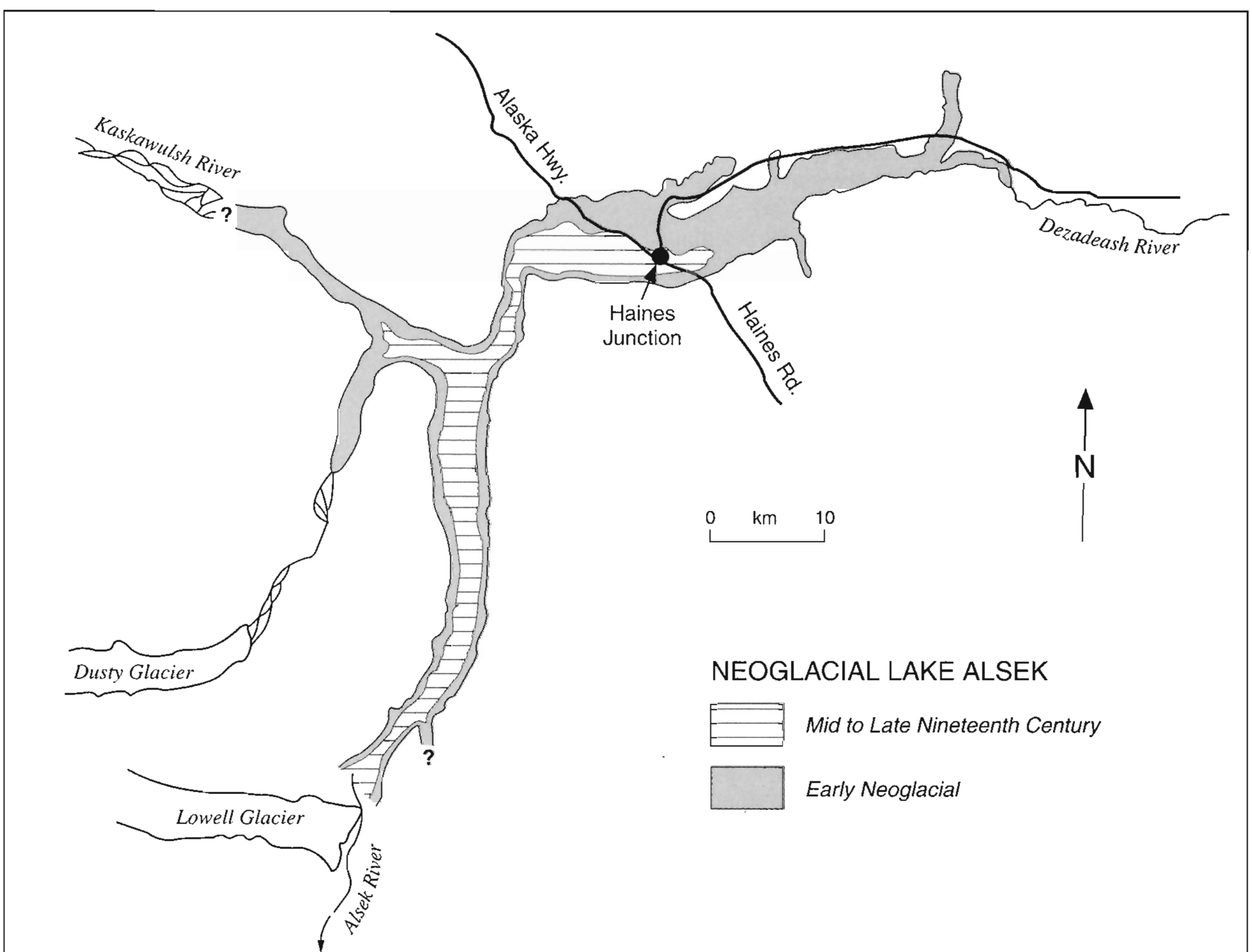


Figure 24. *Maximum extent of Lake Alsek during the last inundation of Dezadeash Valley in the middle of the nineteenth century and during early Neoglacial time. (Clague and Rampton, 1982, p. 98)*

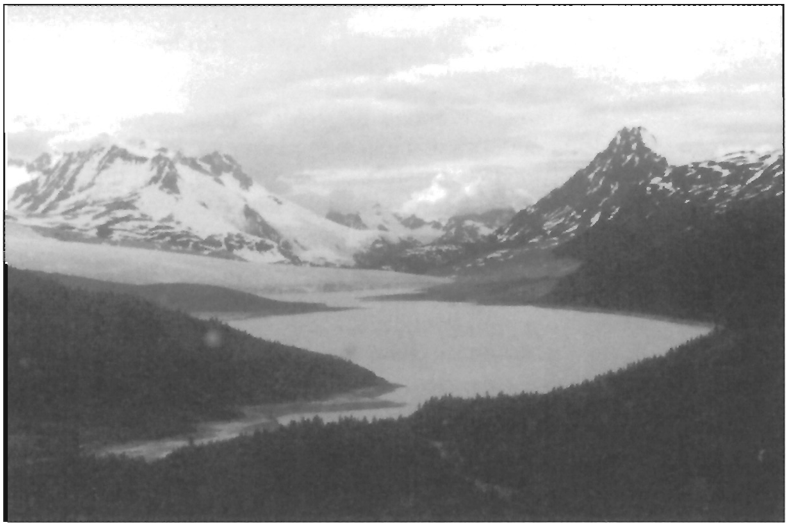


Figure 25.

Ape Lake, Coast Mountains near Bella Coola, British Columbia. The normal lake outlet is in the foreground, but the lake is dammed by Fyles Glacier, visible in the distance. Catastrophic drainage in 1984 and 1986 was through a tunnel in Fyles Glacier. GSC 1997-012U

slowly refills. This cycle may be repeated on an annual basis, or drainage may occur at irregular intervals of several years. The timing of drainage changes as a glacier dam becomes thicker or thinner.

Many "self-dumping" lakes are known in the Canadian Cordillera and some have been studied in detail. For example, a detailed history of Tulsequah Lake in the northern Coast Mountains of British Columbia was compiled by Marcus (1960): drainage occurred annually for many years during the 1940s and 1950s. Appropriately named, Flood Lake, also in the northern Coast Mountains, was monitored by the National Hydrology Research Institute (Mokievsky-Zubok, 1980). Summit Lake, which discharges via the Salmon River (Mathews, 1965; Gilbert, 1972; Mathews and Clague, 1993), is well known because of its proximity to the town of Stewart, British Columbia and a mine. Clague and Evans (1994) have described other examples.

In rugged terrain, glacier outburst floods can be transformed into rapidly moving debris flows if loose morainal or colluvial debris adjacent to the stream channel is incorporated into the floodwater. For example, drainage of a small ice-dammed lake on Cathedral Crags near Field, British Columbia, generated a series of debris flows in the vicinity of the well-known spiral tunnels on the Canadian Pacific Railway (Jackson, 1979). The most publicized flow, in 1978, resulted in partial burial of a freight train and blockage of the Trans-Canada Highway.

During the period of glacier recession and thinning that has occurred since the Little Ice Age, many glacier-dammed lakes have drained for the first time, releasing catastrophic flood discharges down valleys that may not have experienced a similar event for a thousand or more years. For example, Ape Lake, in the British Columbia Coast Mountains south of Bella Coola, drained suddenly for the first time in October 1984 (Fig. 25) (Desloges et al., 1989). In less than 24 hours, $46 \times 10^6 m^3$ of water were released, causing major erosion and damage to roads, bridges, and a logging camp downstream. Fortunately, few people were in the valley when the flood occurred and there were no injuries.

Clearly, the greatest potential hazard exists where glacier-dammed lakes are likely to drain for the first time. Extensive investigations at a settlement site downstream would find no evidence of catastrophic floods, and development could occur with no awareness of the hazard. Even where lakes have drained repeatedly in the past, new settlement can be placed in a hazard zone if the flood evidence is not recognized. Where outburst floods occur relatively frequently close to settled areas, flood hazard is minor because local people are aware of the danger and avoid developing the floodplain.

Glacier-dammed lakes are common around valley glaciers and more extensive ice masses (see Fig. 1 for generalized distribution of glaciers). Young (1977) indicated that possibly the highest concentration of lakes is in the St. Elias Mountains, and there are numerous lakes in the Coast Mountains.

As glacier recession occurs in accordance with climate warming, the behaviour of existing lakes will change. Catastrophic drainage will occur for the first time at lakes that have not previously drained, and other existing lakes will drain more frequently until they and their glacier dams disappear altogether (Clague and Evans, 1994). New lakes with potential for outburst floods will develop at sites that are presently ice covered.

The potential dangers of outburst floods can be offset if glacier-dammed lakes are identified and flood-prone channels designated as hazard zones. Sites where lakes may develop can be identified by air photo interpretation and monitored. For example, existing and potential glacier-dammed lakes in the Stikine and Iskut River basins were identified by Perchanok (1980) in order to estimate flood and sedimentation hazards to proposed hydroelectric dams (Fig. 26). Potential maximum flood discharges can be estimated by empirical and theoretical methods (Clague and Mathews, 1973; Clarke, 1982).

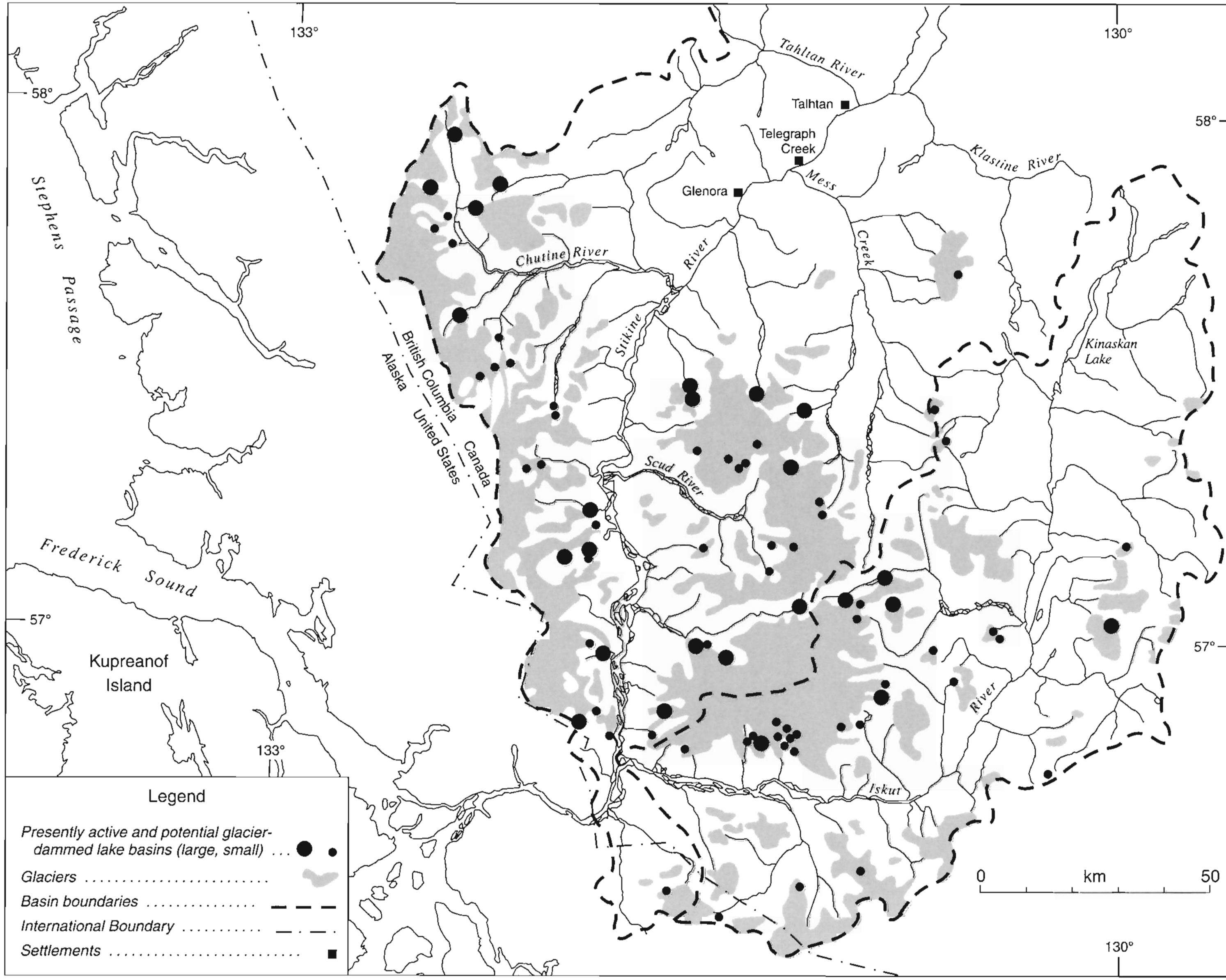


Figure 26. *Existing and potential glacier-dammed lakes in the Stikine-Iskut area (after Perchanok, 1980).*

Floods and debris flows from moraine-dammed lakes

Many of the morainal ridges that were constructed at glacier termini during the Little Ice Age subsequently became dams that impounded lakes between the moraine and the receding glacier. These dams consist of glacial till, which is a loose, porous, heterogeneous mixture of particles ranging from boulders to clay, by no means the ideal, watertight material. Failure of such dams is common, and results in catastrophic floods downstream. Once a dam has burst, lake level is lowered significantly, or the lake may drain completely. Thus most floods of this kind are single events, not reoccurring like glacier outburst floods.

Moraine-dammed lakes are numerous in all regions with glaciers. Even the moraines of small cirque glaciers commonly hold small lakes. Most lakes are impounded by Little Ice Age moraines, but a few older moraine dams survive.

Evidence recognized only a short time ago, together with some recent events, sheds light on the mechanisms of dam failure and suggests that catastrophic floods and debris flows commonly result from the failure of moraine dams. Two examples are illustrative. Blown and Church (1985) documented the spectacular breaching of a Little Ice Age moraine dam at Nostetuko Lakes in the Homathko River basin of the Coast Mountains (Fig. 27). Apparently, a large section of the receding glacier fell into the lake directly opposite its outlet,

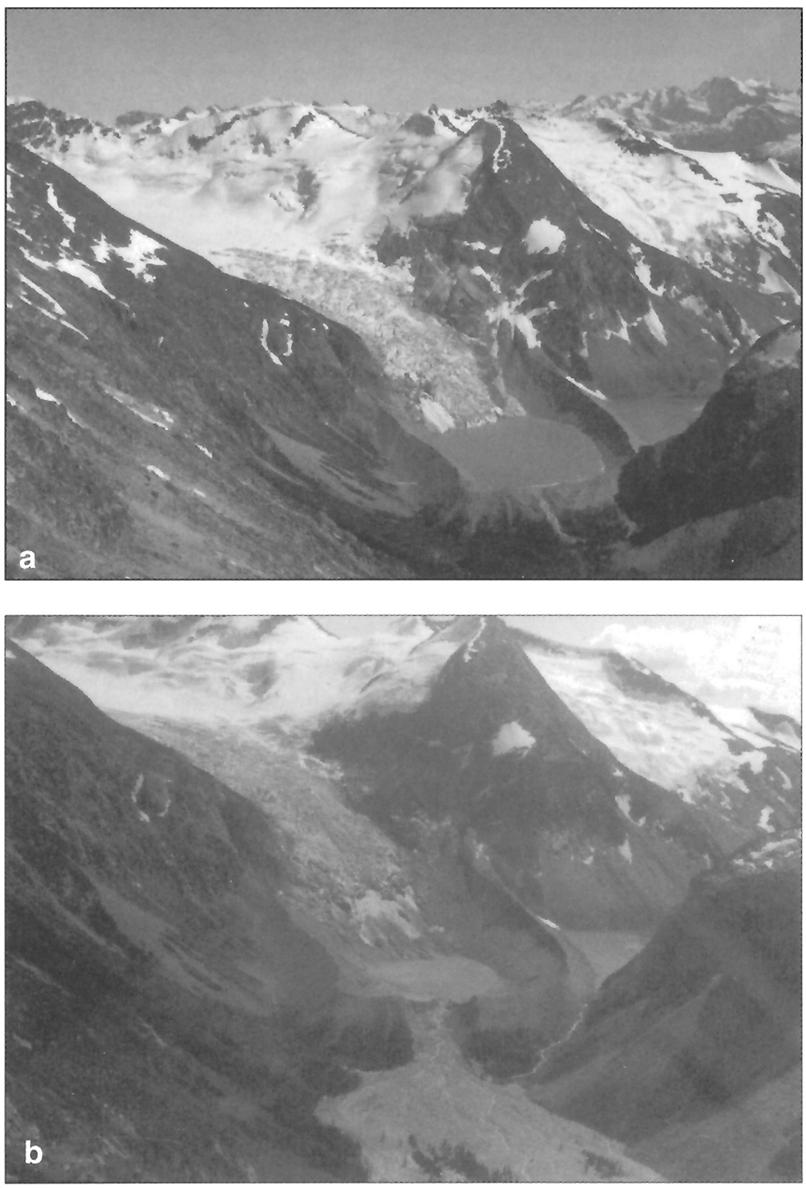


Figure 27. *Nostetuko Lakes before **a)** and after **b)** the moraine-burst flood of 1983, southern Coast Mountains, British Columbia. Photograph **a**: GSC 1997-012Y; Photograph **b** courtesy of M. Church.*

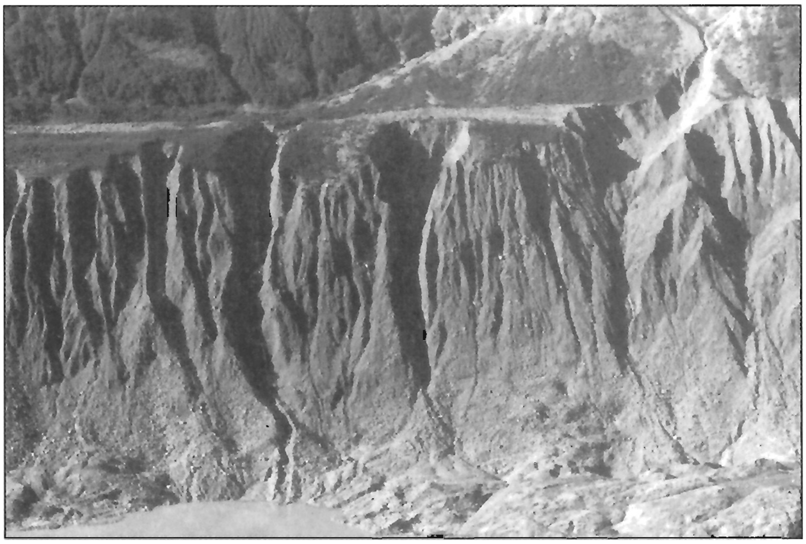


Figure 28.

View of a Little Ice Age lateral moraine that has been exposed by ice recession during the past century. Note gullies formed by rapid erosion due to running water and debris flows; several debris flow cones are visible at the foot of the steep slope; near Toba Inlet, southern Coast Mountains, British Columbia. GSC 1997-012V

generating a large wave that swept into the outlet channel. This caused so much erosion and channel lowering that most of the lake water (6.5 x $10^6 m^3$) escaped during the next 5 hours. The valley floor downstream was drastically modified by erosion and deposition. Fortunately, the area was uninhabited.

During the early 1970s, a massive debris flow was generated by the failure of a moraine dam near Klattasine Creek, another tributary of Homathko River (Clague et al., 1985). Water released from the lake incorporated so much loose debris along the steep valley downstream, that the flood was transformed into a debris flow. At Homathko River, 8 km from the lake, bouldery debris about 20 m thick was deposited, creating a dam which temporarily blocked this large river. Between 2 x 10^6 and 4 x $10^6 m^3$ of debris were moved by the main debris flow plus several subsidiary flows that followed.

Failure of moraine dams can be triggered by a variety of mechanisms. Dams may be overtopped or lake outlets eroded by waves generated by ice falls, rockfalls, and snow avalanches. Rapid runoff from snowmelt and summer rainstorms may raise lakes to unusually high levels, promoting overtopping or piping (subsurface erosion by percolating groundwater and development of conduits) and mass failure due to high pore water pressure. Earthquakes are another possible cause of failure. Clague and Evans (1994) and Costa and Schuster (1988) discuss further examples.

Climate warming, increased precipitation, and glacier recession provide appropriate conditions for the occurrence of catastrophic drainage from moraine-dammed lakes. It is hard to predict, however, whether or not the frequency of catastrophic failures will increase with climate warming. Some conditions that accompany rapid glacier recession, such as increased meltwater discharge, warming of basal ice in subzero glaciers, and melting of ice cores or permafrost within the moraines, promote dam failures. The number of moraine-dammed lakes will gradually diminish, however, unless new ones are created as a result of future glacier behaviour.

Although the time at which a moraine dam will fail cannot be predicted, the maximum discharge of the flood resulting from the collapse of any particular dam can be estimated using easily measurable parameters. On the basis of data gathered about historic moraine-burst floods, Costa and Schuster (1988) determined that maximum flood discharge is statistically related to the potential energy of the lake behind the moraine dam (computed as the product of dam height (metres), volume (cubic metres), and the specific weight of water). Similarly, Evans (1986) determined a relation between maximum flood discharge and the total volume of water that drains from the lake.

Other debris flows, landslides, and ice avalanches

As glacier termini recede up-valley and glacier surfaces are lowered, glacial deposits, chiefly till, are exposed in glacier forefields and on the steep proximal slopes of lateral moraines. This loose sediment is highly susceptible to erosion by running water and to mass movement by slumping and debris flows, particularly if water is added from melting permafrost or buried ice (cf. Mattson and Gardner, 1991). These processes are most intense immediately following deglaciation, and then decline as the most unstable sediment is transferred to lower elevations and slopes are slowly colonized by vegetation (Fig. 28). If rates of glacier recession increase due to climate warming, larger areas of unstable, erodible material will be exposed.

Glacier recession and thinning also result in the unloading of the toes of steep rock slopes, thereby increasing any inherent propensity of a slope for deep-seated failure (cf. Bovis 1990; Evans and Clague, 1992). Rockfalls involving relatively small volumes of material are common in such situations. Larger landslides are less common, except where rocks are particularly weak and unstable. Some of the many postglacial rock avalanches that have occurred in the unstable Quaternary volcanic rocks of the Garibaldi volcanic belt of the southwestern Coast Mountains, for example, can be at least partly attributed to glacier recession (Mokievsky-Zubok, 1977).

Most of the debris flows and landslides that are initiated on newly deglaciated terrain come to rest at the foot of steep slopes within the glacier forefield. Thus they are hazards to be avoided by mountaineers and others who venture close to the glaciers. In some instances, however, major debris flows and rock avalanches have travelled many kilometres down steep valleys, well beyond the glacier forefields, with devastating results. For example, in 1975, near Pemberton in the Garibaldi volcanic belt, a high-velocity debris flow triggered by glacier melting traveled 7 km downvalley, resulting in the death of four people who were temporarily working in a normally uninhabited area (Patton, 1976; Mokievsky-Zubok, 1977).

Ice avalanches (rapid downslope movement of blocks of ice) are commonly associated with melting glaciers on steep slopes. Rapid glacier recession coupled with the release of meltwater will encourage this process. The secondary effects of ice avalanches, such as the flood from Nostetuko Lakes, are potentially dangerous because they may propagate a long way downstream.

Glacier surges

Surges are periods of rapid glacier flow, during which a glacier may advance as much as several kilometres during an interval of a few months to two or three years, or active ice may advance into a terminal zone of stagnant ice, causing ice thickening rather than advance (Paterson, 1994). Eventually, the rapid flow ceases abruptly. Velocity declines to zero in the downvalley part of the glacier, which becomes stagnant and gradually buried by its own melted-out debris. Surges are restricted to specific glaciers, where they occur repeatedly, usually at intervals of several decades. In Canada, all surging glaciers (except one) are located in the St. Elias Mountains. Although all aspects of the rapid glacier flow are not fully understood, it is agreed that surges result from disruption of the normal subglacial drainage system (Paterson, 1994). Climate warming and increased snowfall could modify the timing of surges, and possibly initiate surging on glaciers that have not previously done so.

The rapid flow of a surging glacier is rarely dangerous, except perhaps to mountaineers, because the glacier usually readvances over its own stagnant ice or recent moraines. A severe hazard can arise, however, where a surging glacier advances from its own valley into that of a major river. The high rate of glacier flow permits an effective dam to form (more slowly advancing ice would be melted or washed away), resulting in inundation of the valley upstream. Subsequent failure of this dam causes catastrophic floods downstream, as described previously in this report.

Floods of this nature have resulted from surges of Lowell Glacier and impoundment of Alsek River during the Little Ice Age (Clague and Rampton, 1982) (Fig. 24). At its maximum, this lake was 200 m deep at the ice dam and extended over 100 km upstream to beyond the site of Haines Junction, which was inundated as recently as the middle to late nineteenth century. Flood terraces and giant ripples indicate that the lake emptied catastrophically when the ice dam failed.

Hydrological effects of glacier recession

Like the snowpack, glaciers function as reservoirs that store winter precipitation for later release during the warm season. Glacier meltwater, however, maintains streamflow throughout the summer and into the fall, well after the disappearance of the snowpack. The amount of water released by a glacier is dependent on summer weather, rather than the thickness of snow accumulated during the previous winter. Also, with the exception of areas that are mostly ice covered, such as the St. Elias Mountains, the volume of glacier meltwater is generally much less than that derived from the seasonal snowpack.

Glacier shrinkage or expansion due to climate change would significantly alter the capacity of the glacier reservoirs and the volumes of meltwater. Water yield may be increased in the short term as glaciers melt, but where glaciers disappear altogether, late summer stream discharge could be drastically reduced. Brugman (1992) warned that if present trends of glacier shrinkage continue, severe water shortages are to be expected in the future. In order to assess the overall hydrological impact of climate change, however, many other factors must be considered, such as the hydrology of the snowpack, noted previously in this report, and climatic and hydrological conditions that are not discussed in this paper.

MASS MOVEMENT

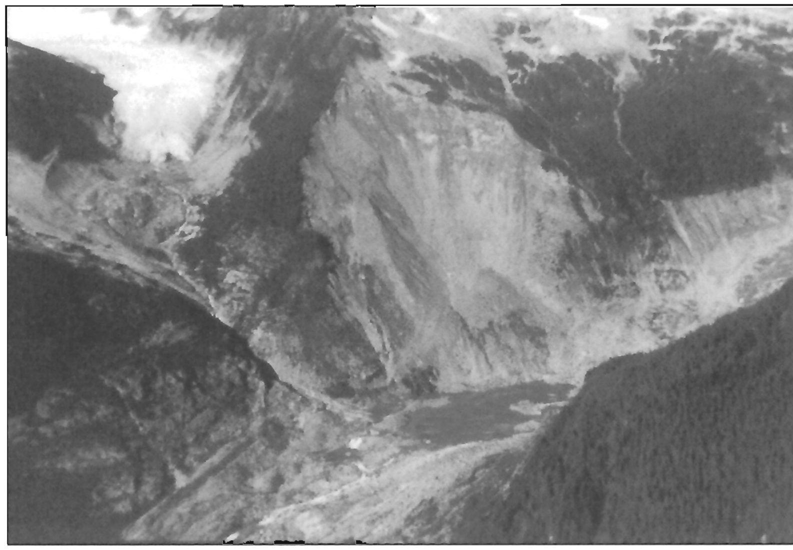
Processes and conditions

Mass movement refers to all types of direct, gravitational, downslope movement of earth materials, including rapidly moving falls, avalanches, slides, and flows of rock and debris, and the slow, imperceptible creep of soil and rock fragments (Fig. 29). Obviously, rapid mass movements are serious hazards. As in the case of snow avalanches, debris flows or landslides that start in the alpine zone may have impacts that extend to considerably lower elevations. Slower forms of mass movement, such as soil creep, pose a minor constraint to some forms of construction. Impacts of slow processes, in general, are sufficiently minor and their relation to weather and climate is insufficiently understood, to warrant no further discussion here.

The locations of sites where rapid mass movements occur are determined by a variety of factors. Many of these relate to the inherent condition of a slope, such as slope steepness, rock type and structure, and thickness and types of Quaternary sediments and soils, and are unrelated to climate or weather. For example, major rock avalanches, resulting from slope movements along inclined bedding planes, are common in regions of sedimentary rocks, such as the Rocky and Mackenzie mountains (Eisbacher, 1979). (A very general impression of the distribution of mass movement processes can be gained from the physiographic information in Fig. 1: slope processes are more intense in the higher mountains.)

Figure 29.

Scar (unvegetated area on mountainside) of a large rock avalanche that probably occurred during the past century near Knight Inlet, southern Coast Mountains, British Columbia. Rock debris fell onto a glacier and was then transported by the ice for about a kilometre (out of the photograph to the left) before being deposited on the valley floor as the glacier melted. Possible contributing causes of slope failure include undercutting and steepening of the slope by the glacier and unloading of the toe of the slope due to glacier recession. GSC 1997-012W



The timing of a sudden failure is often related to short-term weather conditions. For example, rockfalls are commonly triggered by thaw or by unusually intense rainstorms. The initiation of many landslides requires coincidence of an inherently unstable slope with a weather-related trigger. Thus for some processes, the frequency of events has been related to weather conditions. In other cases, additional factors may sometimes override the effects of weather. For example, debris flows at some sites in the Coast Mountains are triggered by intense rain *only* after a period of quiescence during which loose debris accumulates in the initiation zone (Church and Miles, 1987). In this case, even if the frequency of intense rainstorms were to increase, debris flow frequency might not rise. Also, many landslides are triggered by earthquakes. Thus the distribution of many types of mass movement events is extremely irregular in both space and time, and not amenable to either the usual magnitude/frequency methods of analysis or to simple correlation with weather conditions.

Effects of climate change and impact on human activities

Because landslides occur infrequently and are only partially controlled by weather, and because the weather modifications that will be associated with the predicted climate change have not been specified (e.g. intensity of rainstorms, number of freeze-thaw cycles, intensity of winter freezing), it is possible only to speculate on future changes in mass movement activity. If climate change were to result only in an increasing frequency of unusually high-intensity rainstorms, then there could be a related increase in the frequency of some kinds of landslides (e.g. Rapp, 1985). On the other hand, if the only effect of climate change is warmer summers and hence increased evaporation and lowering of groundwater tables, then some types of mass movement could be less frequent. In any event, it is likely that areas that should be identified as hazardous under present day conditions will remain so in the future.

EROSION AND DEPOSITION BY RUNNING WATER

Processes and conditions

Significant erosion and deposition by streams occur only during periods of high discharge, and the amount of work done by a stream depends on the duration and magnitude of the high discharge (Leopold et al., 1964).

In most alpine areas, high stream discharge occurs as a result of two processes: rapid melt of the winter snowpack during late spring and early summer, and runoff from combined rain and snowmelt at other times of the year. Maximum rates of late spring and early summer snowmelt result from direct absorption by the snowpack of solar energy on clear, sunny days (Church, 1988). The rate at which water is released depends primarily upon the solar energy received at the Earth's surface, which will not be appreciably modified by climate change. Runoff from combined rain and snowmelt generates the highest discharges in snowmelt streams, and is thus most effective with regard to stream erosion and deposition. In regions with winter precipitation maxima, such as the Coast Mountains, peak discharges occur during intense cyclonic storms in late fall and early winter. During such storms, the freezing level rises, causing melting of recently deposited snow, and so snowmelt combines with runoff from intense rainfall to produce major floods. In regions with summer precipitation maxima, such as the Rocky Mountains, peak discharges are generated by convectional storms which reach their maximum intensity during late summer and early fall.

Effects of climate change and impact on human activities

It is likely that increased storminess, i.e. increased frequency and magnitude of storms, will accompany climate warming. Thus the storm-induced peak discharges of streams draining from alpine areas will likely increase, resulting in more erosion and sediment transport and deposition by streams. In

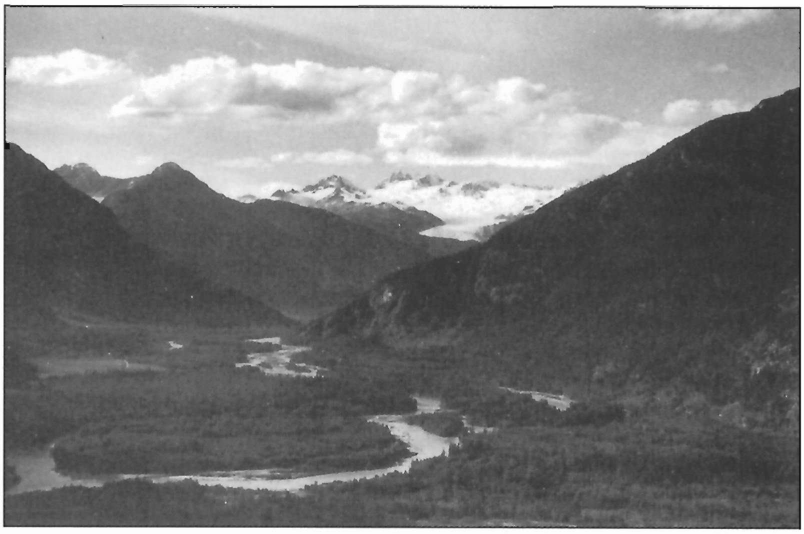


Figure 30.

Lillooet Glacier and upper Lillooet River, southern Coast Mountains, British Columbia. The braided river is laterally unstable and may be slowly aggrading (building up its bed) due to supply of sediment from recently deglaciated terrain. GSC 1997-012X

particular, the frequency of potentially damaging floods is likely to increase, along with the extent of flood-prone land on floodplains. Figure 1 shows streams draining from alpine areas.

An additional effect, possibly with more far-reaching consequences than direct climate-induced changes, will occur along streams draining from glaciers. Ice recession will expose large areas of unvegetated glacier forefields and moraines consisting of highly erodible glacial till (Fig. 28). Erosion and mass wasting of this material could result in locally increased sediment transport in glacial meltwater streams, and, more significantly, deposition of this material at points downstream. Deposition would occur along relatively gently sloping reaches of a stream, probably on floodplains where deposition is occurring at present, and would result in a general rise in the elevation of the floodplain (aggradation) (Fig. 30). This would be accompanied by increased flood-water levels, rapid shifts of channel positions (avulsion), and increased bank erosion. These effects would occur rapidly, during periods of high stream discharge. Aggradation of floodplains could occur within the alpine zone and also at considerable distances downstream. Roads, railroads, and other structures on or close to floodplains could be severely damaged by this process, with related increases in maintenance costs.

Significant changes in the hydrology of streams that drain from alpine areas will also occur as a result of climate change. (A full discussion of this is beyond the scope of this report.) Snow and ice retained in the alpine zone is a significant form of water storage, with runoff being eventually delivered to adjacent areas during the warm season, which is commonly a dry season in intermontane regions. Reduction in snowfall and snowpacks implies reduction in the total water yield from the uplands, which, when coupled with warmer temperatures and increased evaporation, could result in significant water shortages in developed areas.

EOLIAN PROCESSES: THE EFFECTS OF WIND

Process and conditions

Alpine soils commonly include fine particles (chiefly silt-sized) that have been transported to their present sites by wind. In many areas this silty material is intermixed with soil particles weathered from local rock or derived from Pleistocene glacial materials, but in some places, eolian material has accumulated to form a silty, stone-free veneer up to a few decimetres thick. Eolian particles are initially eroded by wind from sites where there is little vegetation, such as Little Ice Age moraines, rocky areas (where silt is released by weathering), bare soil exposed on steep slopes, and ground that has been trampled by people or animals.

Effects of climate change and impact on human activities

In as much as climate change is likely to result in increased dryness in alpine areas and reduction of vegetation density at dry sites, increased erosion and redeposition by wind may occur. Increased long distance transport of very small particles from more distant sources, including areas at much lower elevations, is also likely.

The potential impact on human activities of an increase in wind erosion and deposition appears to be minor. At present, eolian geomorphic activity is unnoticed by most visitors to the alpine zone; blowing dust may be an occasional nuisance, but is likely to be noticed less than the actual wind itself. Changes in the rates of accumulation (in the order of 1 mm per century) of eolian deposits and equivalent rates of erosion are likely to be insignificant when compared to potential human impacts on the alpine environment.

CONCLUSIONS

Geomorphological processes are related to climate in a complex manner. Estimates of how processes will react to anticipated climate modifications, such as warming and increased precipitation, are necessarily tentative and qualitative. They indicate the kinds of changes that may occur in the alpine landscapes, but not the magnitude or precise rates of such changes. Even though the discussion is limited to the area above timberline, the effect of many of the anticipated changes will extend a long way beyond the limits of the alpine zone.

If climate change proceeds as forecast by Global Circulation Models, the following might be expected:

1. Increased frequency of slides and debris flows in some areas due to melting of permafrost and more intense rainfall.
2. More floods and debris flows due to the failure of glacier and moraine dams caused by glacier retreat.
3. Increased avalanche danger due to heavier snowfall.
4. Increased channel instability, erosion, and flooding in glacier-fed rivers caused by increased sediment input due to glacier recession.
5. Reduced seasonal and longer term water storage capacity of alpine areas due to thinner winter snowpacks, rapid spring runoff, and a reduced extent of permanent snow and ice. This could lead to serious water shortages in areas of the plains to the east and dry regions of the interior of the Cordillera that rely on summer meltwater to maintain stream flow.
6. A rise in timberline which will result in a decrease in the extent of the alpine zone, altering the appearance of many mountain areas and their potential for recreation and tourism.
7. A detrimental effect on ski resorts at low elevations, but improve winter access for other activities, due to reduced snow and increased winter rain at elevations close to the present winter snowline, and a positive impact on higher elevation ski resorts due to increased snow fall on higher and colder mountains.
8. A change in scenery in southern mountain areas, such as Banff and Jasper national parks caused by significant loss of glaciers due to warmer summer temperatures.
9. A slow reduction in the extent and thickness of permafrost due to warming of ground temperatures. This will have noticeable effects only where soils contain a high proportion of ice, conditions that are probably not extensive in most alpine areas.

ACKNOWLEDGMENTS

I would like to thank the following people for reviews of various parts of the text and useful comments and suggestions: Margaret North, Dave McLung, and Michael Church of the Department of Geography, University of British Columbia; and Chris Burn of the Department of Geography, Carleton University. The Department of Geography at the University of British Columbia provided laboratory space and access to its map collection during compilation of Figure 1, and I am most grateful to department head, Tim Oke, and map curator Rosemary Cann, for facilitating this arrangement. Betty Hanson assisted in compiling Figure 1 and Alan Paige constructed some of the figures in this report. John Clague, Geological Survey of Canada, and Brian Luckman, University of Western Ontario, provided constructive reviews of the final draft of the manuscript. Also my sincere thanks to Bob Fulton of the Geological Survey of Canada for his advice and patience.

The work was done through Supply and Services Contract 23397-2-1507/01-SS with J.M. Ryder and Associates, Terrain Analysis Inc., Vancouver, British Columbia.

REFERENCES

Arno, S.F. and Hammerly, R.P.
1984: Timberline, mountain and arctic forest frontiers; The Mountaineers, Seattle, Washington, 304 p.

Barry, R.G.
1992: Mountain climatology and past and potential future climatic changes in mountain regions: a review; Mountain Research and Development, v. 12, p. 71-86.

Beaudoin, A.B.
1989: The response of the timberline zone to climatic change in the Canadian Rocky Mountains; The Canadian Geographer, v. 33, p. 83-86.

Blown, I. and Church, M.
1985: Catastrophic lake drainage within Homathko River basin, British Columbia; Canadian Geotechnical Journal, v. 22, p. 551-563.

Bostock, H.S.
1948a: Physiographic subdivisions of the Canadian Cordillera north of the fifty-fifth parallel; Geological Survey of Canada, Map 922A, scale 1:2 534 400.
1948b: Physiography of the Canadian Cordillera, with special reference to the area north of the fifty-fifth parallel; Geological Survey of Canada, Memoir 247, 106 p.
1970: Physiographic regions of Canada; Geological Survey of Canada, Map 1254A, scale 1:5 000 000.

Boudreau, F. and Payette, S.
1974: Le mont Jacques-Cartier, Parc de la Gaspésie; De toute urgence, v. 5, p. 3-18.

Bovis, M.J.
1990: Rock-slope deformation at Affliction Creek, southern Coast Mountains, British Columbia; Canadian Journal of Earth Sciences, v. 27, p. 243-254.

Brink, V.C.
1959: A directional change in the subalpine forest-heath ecotone in Garibaldi Park, British Columbia; Ecology, v. 40, p. 10-16.
1964: Plant establishment in the high snowfall alpine and subalpine regions of British Columbia; Ecology, v. 45, p. 431-438.

British Columbia Ministry of Forests
1988: Biogeoclimatic zones of British Columbia; British Columbia Ministry of Forests, Victoria, British Columbia, scale 1:2 000 000.

Brown, R.J.E.
1967: Permafrost in Canada; Geological Survey of Canada, Map 1246A.
1974: Some aspects of air photo interpretation of permafrost in Canada; National Research Council of Canada, Division of Building Research, Technical Paper 409, 20 p. plus figures.
1979: Permafrost distribution in the southern part of the discontinuous zone in Nouveau-Québec and Labrador; Géographie physique et Quaternaire, v. 33, p. 279-289.

Brugman, M.M.
1992: Search for trends in glacier mass balance from western Canada; in Using Hydrometric Data to Detect and Monitor Climatic Change; Proceedings of NHRI Symposium No. 8, April 1992; National Hydrology Research Institute, Saskatoon.

Burn, C.R. and Lewkowicz, A.G.
1990: Retrogressive thaw slumps; Canadian Geographer, v. 34, p. 273-276.

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, v. 1, p. 161-176.

Canada Pêches et environnement
1977: Classification écologique du Térritoire du Labrador; Canada Pêches et environnement, Service de la gestion de l'environnement, Direction General des Terres, Région de l'Atlantique, échelle 1/1 000 000.

Carson, M.A. and Bovis, M.J.
1989: Slope processes; in Chapter 9 of Quaternary Geology of Canada and Greenland, (ed.) R.J. Fulton; Geological Survey of Canada, Geology of Canada, no.1, p. 583-594 (also Geological Society of America, The Geology of North America, v. K-1).

Chilton, R.H.
1981: A summary of climatic regimes of British Columbia; British Columbia Ministry of Environment, Assessment and Planning Division, Victoria, British Columbia, 46 p.

Church, M.
1988: Floods in cold climates; in Flood Geomorphology, (ed.) V.R. Baker, R.C. Kochel, and P.C. Patton; John Wiley and Sons, New York, p. 205-230.

Church, M. and Miles, M.J.
1987: Meteorological antecedents to debris flow in southwestern British Columbia: some case studies; Geological Society of America, Reviews in Engineering Geology, v. VII, p. 63-79.

Church. M., Stock, R.F., and Ryder, J.M.
1979: Contemporary sedimentary environments on Baffin Island, N.W.T., Canada: debris slope accumulations; Arctic and Alpine Research, v. 11, p. 371-402.

Clague, J.J.
1987: Catastrophic outburst floods; Geos, v. 16, no. 2, p. 18-21.

Clague, J.J. and Evans, S.G.
1994: Formation and failure of natural dams in the Canadian Cordillera; Geological Survey of Canada, Bulletin 464, 35 p.

Clague, J.J. and Mathews, W.H.
1973: The magnitude of jökulhlaups; Journal of Glaciology, v. 12, p. 501-504.

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

Clague, J.J., Evans, S.G., and Blown, I.
1985: A debris flow triggered by the breaching of a moraine-dammed lake, Klattasine Creek, British Columbia; Canadian Journal of Earth Sciences, v. 22, p. 1492-1502.

Clarke, G.K.C.
1982: Glacier outburst floods from "Hazard Lake", Yukon Territory, and the problem of flood magnitude prediction; Journal of Glaciology, v. 28, p. 3-21.

Costa, J.E. and Schuster, R.L.
1988: The formation and failure of natural dams; Geological Society of America Bulletin, v. 100, p. 1054-1068.

Desloges, J.R., Jones, D.P., and Ricker, K.E.
1989: Estimates of peak discharge from the drainage of ice-dammed Ape Lake, British Columbia, Canada; Journal of Glaciology, v. 35, p. 349-354.

Duk-Rodkin, A. and Hughes, O.L.
1992: Pleistocene montane glaciation in the Mackenzie Mountains, Northwest Territories; Géographie physique et Quaternaire, v. 46, p. 69-84.

Eisbacher, G.H.
1979: First-order regionalization of landslide characteristics in the Canadian Cordillera; Geoscience Canada, v. 6, p. 69-79.

Eisbacher, G.H. and Clague, J.J.
1984: Destructive mass movements in high mountains: hazard and management; Geological Survey of Canada, Paper 84-16, 230 p.

Evans, I.S.
1990: Climatic effects on glacier distribution across the southern Coast Mountains, B.C., Canada; Annals of Glaciology, v. 14, p. 58-64.

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

Evans, S.G. and Clague, J.J.
1992: Glacier-related hazards and climatic change; in The World at Risk: Natural Hazards and Climate Change, American Institute of Physics, Conference Proceedings 277, Cambridge, Massachusetts, p. 48-60.

Fahn, C.
1975: Northern Labrador (1969); in Mountain Glaciers of the Northern Hemisphere, (ed.) W.O. Field, U.S. Army Corps of Engineers, Cold Regions Research and Engineering Laboratory, Hanover, New Hampshire, 1640 p.

Falconer, G., Henoch, W.S., and Ostrem, G.M.
1965: Glacier Map of southern British Columbia and Alberta; Department of Mines and Technical Surveys, Geographical Branch, Ottawa, Ontario, scale 1:1 000 000.

Field, W.O.
1975: Mountain glaciers of the Northern Hemisphere; United States Army Corps of Engineers, Cold Regions Research and Engineering Laboratory, Hanover, New Hampshire, 1640 p.
1990: Glaciers of Alaska and adjacent Yukon Territory and British Columbia; American Alpine Journal, v. 32, p. 79-149.

Fisheries and Environment Canada
1978: Hydrologic Atlas of Canada; Canada Department of Fisheries and Environment, Ottawa, 34 maps.

French, H.M.
1976: The periglacial environment; Longman, London, 309 p.

Gardner, J.
1970: Rockfall: a geomorphic process in high mountain terrain; Albertan Geographer, v. 6, p. 16-20.

Gilbert, R.
1972: Drainings of ice-dammed Summit Lake, British Columbia; Canada Department of Environment, Inland Waters Directorate, Scientific Series, no. 20, 17 p.

Goodrich, L.E.
1982: The influence of snow cover on the ground thermal regime; Canadian Geotechnical Journal, v. 19, p. 421-432.

Government of Canada
1994: Canada's National Report on Climate Change; Government of Canada, Ottawa, 144 p.

Granberg, H.B.
1973: Indirect mapping of the snowcover for permafrost prediction at Schefferville, Quebec; Proceedings of the Second International Conference on Permafrost, Yakutsk, USSR, North American Contribution, Washington, D.C., National Academy of Science, p. 113-120.

Gray, J.T. and Brown, R.J.E.
1979: Permafrost presence and distribution in the Chic Chocs Mountains, Gaspésie, Québec; Géographie physique et Quaternaire, v. 33, p. 299-316.

Griggs, B.F.
1938: Timberlines in the northern Rocky Mountains; Ecology, v. 19, p. 548-564.

Haeberli, W.
1990: Glacier and permafrost signals of 20th-century warming; Annals of Glaciology, v. 14, p. 99-101.

Hall, K.
1986: Rock moisture content in the field and in the laboratory and its relationship to mechanical weathering studies; Earth Surface Processes and Landforms, v. 11, p. 131-142.

Harding, K.A.
1985: Glacier inventory of the Premier Range, British Columbia; B.A. thesis, University of Western Ontario, London, Ontario, 101 p.

Hare, F.K. and Thomas, M.K.
1979: Climate Canada; Wiley Publishers of Canada Limited, Toronto, Ontario, 230 p.

Harris, S.A.
1983: Comparison of the climatic and geomorphic methods of predicting permafrost distribution in western Yukon Territory; Proceedings of the 4th International Conference on Permafrost, Fairbanks, Alaska; National Academy of Sciences – National Academic Press Washington, D.C., p. 450-455.
1986: The permafrost environment; Croom Helm, London, 276 p.

Harris, S.A. and Brown, R.J.E.
1982: Permafrost distribution along the Rocky Mountains in Alberta; in Proceedings of the 4th Canadian Permafrost Conference, Calgary, Alberta, National Research Council of Canada, Ottawa, p. 59-67.

Henoch, W.E. and Stanley, A.
1967a: Glacier map of northern British Columbia and southeastern Alaska; Canada Department of Energy Mines and Resources, Inland Waters Branch, Glaciological Section, Ottawa, Ontario, scale 1:1 000 000.

1967b: Glacier map of Yukon Territory and District of Mackenzie; Canada Department of Energy, Mines and Resources, Inland Waters Branch, Glaciological Section, Ottawa, Ontario, scale 1:1 000 000.

Holland, S.S.
1964a: British Columbia physiographic subdivisions; British Columbia Department of Mines and Petroleum Resources, Victoria, British Columbia, scale 1:1 900 800.

1964b: Landforms of British Columbia, a physiographic outline; British Columbia Department of Mines and Petroleum Resources, Victoria, British Columbia, 138 p.

Ives, J.D.
1962: Iron mining in permafrost, central Labrador-Ungava; Geographical Bulletin, v. 17, p. 66-77.

1974: Permafrost; Chapter 4a in Arctic and Alpine Environments; (ed.) J.D. Ives and R.G. Barry; Methuen, London, p.159-194.

1979: A proposed history of permafrost development in Labrador-Ungava; Géographie physique et Quaternaire, v. 33, p. 233-244.

Jackson, L.E., Jr.
1979: A catastrophic glacial outburst flood (jökulhlaup) mechanism for debris flow generation at the Spiral Tunnels, Kicking Horse River basin, British Columbia; Canadian Journal of Earth Sciences, v. 16, p. 806-813.

Kearney, M.S.
1982: Recent seedling establishment at timberline in Jasper National Park; Canadian Journal of Botany, v. 60, p. 2282-2287.

Kershaw, G.P. and Gill, D.
1979: Growth and decay of palsas and peat plateau in the Macmillan Pass - Tsichu River areas, N.W.T., Canada; Canadian Journal of Earth Sciences, v. 16, p. 1362-1374.

Kuhn, M.
1985: Reactions of mid-latitude glacier mass balance to predicted climatic changes; in Glaciers, Ice Sheets and Sea Level: Effect of a CO_2 Induced Climatic Change; United States Department of Commerce, National Technical Information Service, Springfield, Virginia, p. 248-254.

Lamb, H.F.
1985: Palynological evidence for postglacial change in the position of tree limit in Labrador; Ecological Monographs, v. 55, p. 241-58.

Leopold, L.B., Wolman, M.G., and Miller, J.P.
1964: Fluvial processes in geomorphology; W.H. Freeman and Company, San Francisco, California, 522 p.

Lewkowicz, A.G.
1992: Factors influencing the distribution and initiation of active-layer detachment slides on Ellesmere Island, Arctic Canada; in Periglacial Geomorphology, (ed.) J.C. Dixon and A.D. Abrahams; John Wiley and Sons Ltd., Chichester, Great Britain, p. 223-250.

Luckman, B.H.
1976: Rockfalls and rockfall inventory data: some observations from Surprise Valley, Jasper National Park, Canada; Earth Surface Processes, v. 1, p. 287-298.

1978: Geomorphic work of snow avalanches in the Canadian Rocky Mountains; Arctic and Alpine Research, v. 10, p. 261-276.

1990: Mountain areas and global change: a view from the Canadian Rockies; Mountain Research and Development, v. 10, p. 183-195.

1994: Evidence for climate conditions between 900-1300 A.D. in the southern Canadian Rockies; Climatic Change, v. 26, p. 171-182.

Luckman, B.H. and Kearney, M.S.
1986: Reconstruction of Holocene changes in alpine vegetation and climate in the Maligne Range, Jasper National Park, Alberta; Quaternary Research, v. 26, p. 244-261.

Luckman, B.H., Harding, K.A., and Hamilton, J.P.
1987: Recent glacier advances in the Premier Range, British Columbia; Canadian Journal of Earth Sciences, v. 24, p. 1149-1161.

Mackay, J.R.
1975: The stability of permafrost and recent climatic change in the Mackenzie Valley, N.W.T.; in Report of Activities, Part B; Geological Survey of Canada, Paper 75-1B, p. 173-175.

Marcus, M.G.
1960: Periodic drainage of glacier-dammed Tulsequah Lake, British Columbia; Geographical Review, v. 50, p. 89-106.

Mathews, W.H.
1951: Historic and prehistoric fluctuations of alpine glaciers in the Mount Garibaldi map-area, southwestern British Columbia; Journal of Geology, v. 59, p. 357-380.

1955: Permafrost and its occurrence in the southern Coast Mountains of British Columbia; Canadian Alpine Journal, v. 38, p. 94-98.

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

Mathews, W.H. (comp.)
1986: Physiography of the Canadian Cordillera; Geological Survey of Canada, Map 1701A, scale 1:5 000 000.

Mathews, W.H. and Clague, J.J.
1993: The record of jökulhlaups from Summit Lake, northwestern British Columbia; Canadian Journal of Earth Sciences, v. 30, p. 499-508.

Mattson, L.E. and Gardner, J.S.
1991: Mass wasting on valley-side ice-cored moraines, Boundary Glacier, Alberta, Canada; Geografiska Annaler, v. 73A, p. 123-128.

McCarthy, D.P. and Smith, D.J.
1994: Historical glacier activity in the vicinity of Peter Lougheed Provincial Park, Canadian Rocky Mountains; Western Geography, v. 4, p. 94-109.

McClung, D. and Schaerer, P.
1993: The Avalanche Handbook; The Mountaineers, Seattle, Washington, 271 p.

Meier, M.F.
1965: Glaciers and climate; in The Quaternary of the United States, (ed.) H.E. Wright and D.G. Frey; Princeton University Press, Princeton, New Jersey, p. 795-805.

Mitchell, J.F.B., Manabe, S., Meleshko, V., and Tokioka, T.
1990: Equilibrium climate change - and its implications for the future; in Climate Change; the IPCC Scientific Assessment, (ed.) J.T. Houghton, G.J. Jenkins, and J.J. Ephraums; Cambridge University Press, Cambridge, p. 135-164.

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

1980: Sudden discharge of glacier-dammed Flood Lake, August 1979; Canada Department of Environment, National Hydrology Research Institute, Summary Report, 17 p.

Natural Resources Canada
1995: Canada Permafrost, The National Atlas of Canada 5th edition; National Atlas Information Service, Canada Centre for Mapping; Geomatics Canada; Natural Resources Canada, Ottawa, scale 1:7 500 000.

North, M.
1965: Vegetation of Alberta; Province of Alberta and University of Alberta, Edmonton, Alberta.

Ostrem, G.
1966: The height of the glaciation limit in southern British Columbia and Alberta; Geografiska Annaler, v. 48A, p. 126-138.

1972: Height of the glaciation level in northern British Columbia and southeastern Alaska; Geografiska Annaler, v. 54A, p. 76-84.

Paterson, W.S.B.
1994: The physics of glaciers (3rd edition); Pergamon Press, New York, 480 p.

Patton, F.D.
1976: The Devastation Glacier slide, Pemberton, B.C.; in Programme and Abstracts; Geological Association of Canada, Cordilleran Section, Annual Meeting, Vancouver, British Columbia, p. 26.

Payette, S. and Filion, L.
1984: White spruce expansion at the tree line and recent climatic change; Canadian Journal of Forest Research, v. 15, p. 241-251.

Payette, S., Filion, L., Delwaide, A., and Bégin, C.
1989: Reconstruction of tree-line vegetation response to long-term climatic change; Nature, v. 341, p. 429-432.

Perchanok, M.S.
1980: Reconnaissance of glacier-dammed lakes in the Stikine and Iskut River basins, British Columbia; Canada Department of Environment, Inland Waters Directorate, National Hydrology Research Institute, Ottawa, Ontario, 29 p.

Perla, R.I. and Martinelli, M., Jr.
1976: Avalanche Handbook; United States Department of Agriculture, Forest Service, Fort Collins, Colorado, Agriculture Handbook 489, 238 p.

Péwé, T.L.
1983: Alpine permafrost in the contiguous United States; Arctic and Alpine Research, v. 15, p. 145-156.

Prest, V.K., Grant, D.R., and Rampton, V.N.
1967: Glacial map of Canada; Geological Survey of Canada, Map 1253A, scale 1:5 000 000.

Price, L.W.
1970: Morphology and ecology of solifluction lobe development, Ruby Range, Yukon Territory; Ph.D. thesis, University of Illinois, Urbana, Illinois, 325 p.

Rampton, V.N.
1973: The influence of ground ice and thermokarst upon the geomorphology of the Mackenzie Beaufort region; in Research in Polar and Alpine Geomorphology; Proceedings, 3rd Guelph Symposium on Geomorphology, Guelph, Ontario, p. 43-60.

1982: Quaternary geology of the Yukon coastal plain; Geological Survey of Canada, Bulletin 317, 49 p.

Rapp, A.
1985: Extreme rainfall and rapid snowmelt as causes of mass movement in high latitude mountains; in Field and Theory: Lectures in Geocryology, (ed.) M. Church and H.O. Slaymaker; University of British Columbia Press, Vancouver, British Columbia, 213 p.

Sanguin, A-L. and Gill, A.M.
1990: Le Columbia Icefield et le Glacier Athabasca (Rocheuses Canadiennes); Geographica Helvetica, v. 45, p. 85-104.

Schaerer, P.A.
1977: Analysis of snow avalanche terrain; Canadian Geotechnical Journal, v. 14, p. 281-287.

Seppälä, M.
1986: The origin of palsas; Geografiska Annaler, v. 68A, p. 141-147.

Smith, D.J.
1987: Late Holocene solifluction lobe activity in the Mount Rae area, southern Canadian Rocky Mountains; Canadian Journal of Earth Sciences, v. 24, p. 1634-1642.

Smith, M.W.
1986: The significance of climate change for the permafrost environment; in Climate Change Impacts in the Canadian Arctic, (ed.) H.M. French; Proceedings of a Canadian Climate Program Workshop, Geneva Park, Ontario, p. 67-81.

1990: Potential responses of permafrost to climate change; Journal of Cold Regions Engineering, v. 4, p. 29-37.

Smith, M.W. and Riseborough, D.W.
1983: Permafrost sensitivity to climatic change; Proceedings, Fourth International Conference on Permafrost, Fairbanks, Alaska, p. 1178-1183.

Thompson, W.F.
1990: Climate related landscapes in world mountains: criteria and map; Zeitschrift fur Geomorphologie, Supplementband 78, 92 p.

Tranquillini, W.
1979: Physiological Ecology of the Alpine Timberline: tree existence at high altitudes with special reference to the European Alps; Springer-Verlag, New York, 137 p.

Walder, J. and Hallett, B.
1985: A theoretical model of rock fracture during freezing; Geological Society of America Bulletin, v. 96, p. 481-504.

Washburn, A.L.
1979: Geocryology; Edward Arnold, 406 p.

White, S.E.
1976: Is frost action really only hydration shattering? A review; Arctic and Alpine Research, v. 8, p. 1-6.

Williams, P.J. and Smith, M.W.
1989: The Frozen Earth, Fundamentals of Geocryology; Cambridge University Press, Cambridge, 306 p.

Young, G.J.
1977: Glacier outburst floods; in Canadian Hydrology Symposium, 77; National Research Council of Canada, Associate Committee on Hydrology, Proceedings, p. 1-12.

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, p. 59-72.

APPENDIX A

Preparation of Figure 1: characteristics and sensitivity of Alpine regions in Canada to climate change

Draft maps of British Columbia-Alberta, Yukon-District of Mackenzie, and Labrador-Newfoundland regions were compiled at a scale of 1:2 000 000, and subsequently reduced and combined at the presentation scale of 1:7 500 000.

DELIMITATION OF ALPINE AREAS

Method

Alpine areas in British Columbia and Alberta have been previously defined on 1:2 000 000 scale maps of biogeoclimatic zones (British Columbia Ministry of Forests, 1988) and vegetation (North, 1965). Alpine zone boundaries from these maps were redrawn and generalized to a level of detail suitable for presentation at 1:7 500 000.

Alpine areas elsewhere were derived by interpretation and generalization of timberline locations shown on 1:250 000 scale maps of the National Topographic System. Most white areas above the upper limit of forested land (green on these maps) were assumed to be alpine terrain. Little difficulty was encountered in identifying nonalpine treeless areas, such as bogs, bare rock, talus, and avalanche tracks, and in any case, many of these features are insignificantly small at the presentation scale. Where necessary, local timberline elevation was determined from the map and used to exclude nonalpine areas. Alpine "islands" smaller than about 7.5 x 15 km were ignored or grouped into larger units, and the outlines of small irregularities were smoothed. As a result, many small enclaves of alpine terrain are omitted from the final figure, and many narrow (less than about 7.5 km wide) forested valleys are included in the alpine zone. In general, islands of alpine terrain were grouped and included with intervening valleys where the resulting map unit consisted of at least 50% alpine terrain. Measurement from Figure 1 (at the publication scale of 1:7 500 000), however, would give only a rough approximation of the total alpine area.

The delimitation of generalized alpine areas was carried out by hand and eye. Modified boundary lines from the 1:2 000 000 scale vegetation maps were traced directly onto the 1:2 000 000 scale compilation maps. Generalized data from the 1:250 000 maps were reduced to the scale of the compilation map by the use of grids.

Difficulties encountered

a) Additional criteria for the delimitation of alpine areas were employed in far northern areas to distinguish alpine from arctic terrain. Along the west side of the Labrador uplands in northern Quebec and in northern Yukon Territory and District of Mackenzie, the 300 m contour was adopted as the boundary between alpine areas and arctic coastal lowlands. In terrain with high relief, subdivision of distinct physiographic units (e.g. mountain ranges) was avoided. Thus although no timberline is present (i.e. there are no trees) in the northern Torngat Mountains of Labrador, these mountains are included as an alpine area because the southern parts of the same range do rise above a timberline and hence fit the alpine definition.

b) Along the coasts of Labrador and Newfoundland, timberline is commonly depressed locally due to the effects of coastal exposure, and in many places, treeless areas extend down to sea level. These relatively low elevation zones are included as alpine areas because geomorphological processes here, in the absence of trees, are similar to those at higher elevations.

c) Timberline information is not provided on the 1:250 000 scale topographic maps for 23 O and 24 C, E, F, and L NTS areas. In these parts of Quebec, timberline was delimited by determining timberline elevation from adjacent maps.

d) In a few cases, there were obvious discrepancies in the elevation of timberline between adjacent maps.

e) Difficulties were encountered in the precise transfer of generalized boundary lines from the biogeoclimatic zones map of British Columbia to the 1:2 000 000 scale base map because the former map has no geographic grid and appears to have no consistent scale. Adjustments were made by using the local coastline or road networks as match lines for many small subdivisions of the map, and by more or less continuously shifting the relative position of the two maps during the transfer. Many alpine areas are slightly misplaced as a result.

DESIGNATION OF ALPINE AREA DESCRIPTORS: PHYSIOGRAPHY AND MODERN GLACIERS

Alpine areas that were delimited as described above were subdivided on the basis of physiography and modern glaciation only where significant differences exist that can be shown at the presentation scale of Figure 1. Physiographic descriptors are based on information provided by maps of physiographic regions (Bostock, 1948a, 1970; Holland, 1964a; Mathews, 1986) and ecological regions (Canada Peches et environment, 1977), topographic maps at scales of 1:1 000 000 and 1:250 000, and descriptive accounts of the physiography of the western Cordillera (Bostock 1948b; Holland 1964b).

Glacier maps of British Columbia and Alberta (Falconer et al., 1965; Henoch and Stanley, 1967a) and Yukon Territory and District of Mackenzie (Henoch and Stanley, 1967b), descriptions and maps in Field (1975), and 1:250 000 scale topographic maps of Labrador were used to identify the extent of modern glaciers. Glacier types and the areal proportions of ice cover were estimated visually from the glacier maps.

DESIGNATION OF REGIONS FOR CLIMATE AND PERMAFROST

Climatic regions were defined qualitatively and subjectively on the basis of three types of information. Previously defined climatic regions (cf. Hare and Thomas, 1979; Chilton, 1981), which are based on data derived from lowland observation sites, were used to provide a broad framework for alpine climate classification. Then climatic parameters known to have a strong influence on geomorphic processes (e.g. snowfall) and climatic differences reflected by landforms in alpine areas (e.g. presence or absence of rock glaciers) were considered. A full and interesting discussion of these and other concepts pertinent to the nonquantitative climatic classification of mountainous areas is provided by Thompson (1990).

Regionalization of alpine climate cannot be based on quantitative data due to the scarcity of standard weather observations above timberline. Also, quantitative data cannot be easily estimated from low-level weather records because of the great variability in both space and time of alpine weather, and the lack of clearly defined relations for trends of precipitation and temperature with elevation.

Maps in the Hydrological Atlas of Canada (Fisheries and Environment Canada, 1978) were used to assess the regional distribution of the climatic values used for climate classification: precipitation, snowfall, seasonality of precipitation, January temperatures, and annual temperature range. Regional data on the variables of most direct relevance to geomorphic processes, such as precipitation intensity, are not available for most alpine areas.

The distribution of permafrost is described in the table according to references noted in the text.