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Into the second century of worldwide glacier monitoring-prospects and strategies

A contribution to the
International Hydrological Programme (IHP)
and the
Global Environment Monitoring System (GEMS)

Prepared by the World Glacier Monitoring Service
Edited by W. Haeberli, M. Hoelzle and S. Suter

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Preface

Although a total amount of water on Earth is generally assumed to have remained virtually constant, the rapid growth of population, together with the extension of irrigated agriculture and industrial development, are putting stress on the quality and quantity aspects of natural systems. Because of the increasing problems, society has begun to realize that it can no longer follow a 'use and discard' philosophy – either with water resources or any other natural resource. As a result, the need for a consistent policy of rational management of water resources has become evident.

Rational water management should be founded upon a thorough understanding of water availability and movement. Thus, as a contribution to the solution of the world's water problems, UNESCO, in 1965, began the first world-wide programme of studies of the hydrological cycle – the International Hydrological Decade (IHD). The research programme was complemented by a major effort in the field of hydrological education and training. The activities undertaken during the Decade proved to be of great interest and value to Member States. By the end of that period, a majority of UNESCO's Member States had formed IHD National Committees to carry out relevant national activities and to participate in regional and international cooperation within the IHD programme. The knowledge of the world's water resources had substantially improved. Hydrology became widely recognized as an independent professional option and facilities for training hydrologists had been developed.

Conscious of the need to expand upon the efforts initiated during the International Hydrological Decade and further to the recommendations of Member States, UNESCO launched a new long-term intergovernmental programme in 1975: the International Hydrological Programme (IHP).

Although the IHP is basically a scientific and educational programme, UNESCO has been aware from the beginning of a need to direct its activities towards the practical solutions of the world's very real water resources problems. Accordingly, and in line with the recommendations of the 1977 United Nations Water Conference, the objectives of the International Hydrological Programme have been gradually expanded in order to cover not only hydrological processes considered in interrelationship with the environment and human activities, but also the scientific aspects of multi-purpose utilization and conservation of water resources to meet the needs of economic and social development. Thus, while maintaining IHP's scientific concept, the objectives have shifted perceptibly towards a multi-disciplinary approach to the assessment, planning, and rational management of water resources.

As part of UNESCO's contribution to the objectives of the IHP, two publication series are issued: 'Studies and reports in hydrology' and 'Technical papers in hydrology'. In addition to these publications and in order to expedite the exchange of information in the areas in which it is most needed, works of a preliminary nature are issued in the form of technical documents.

The purpose of the continuing series 'Studies and reports in hydrology', to which this volume belongs, is to present data collected and the main results of hydrological studies, as well as to provide information on hydrological research techniques. The proceedings of symposia are also sometimes included. It is hoped that these volumes will furnish material of both practical and theoretical interest to water resources scientists and also to those involved in water resources assessment and planning for rational water resources management.

Contents

Foreword

International coordination of long-term glacier observations is a century-long tradition. It started with the establishment of the International Glacier Commission during the 6th International Geological Congress at Zurich, Switzerland in 1894. The goals of this worldwide glacier monitoring programme were defined by F.-A. Forel from Geneva, the first president of the Commission, in a remarkable article entitled *Les variations périodiques des glaciers. Discours préliminaire,* published on pp. 209–29 of the Archives des sciences physiques et naturelles (Geneva, Vol. 34).

Since 1894, the goals of internationally coordinated glacier monitoring have evolved and multiplied. Today, the evolution of glaciers and ice caps is recognized as one of the key variables relating to early detection strategies in view of possible man-induced climatic change. The general shrinkage of mountain glaciers during the 20th century is a major reflection of the fact that rapid secular change in the energy balance of the Earth's surface is taking place on a global scale.

As a contribution to the International Hydrological Programme (IHP) of the United Nations Educational, Scientific and Cultural Organization (UNESCO) and to the Global Environment Monitoring System (GEMS) of the United Nations Environment Programme (UNEP), the World Glacier Monitoring Service (WGMS) of the International Commission on Snow and Ice (ICSI/IAHS), as one of the permanent services of the Federation of Astronomical and Geophysical Data Analysis Services (FAGS/ICSU), collects and publishes standardized glacier data.

The present publication was prepared to mark the occasion of the centenary of worldwide glacier monitoring. Scientific review of the text was accomplished by the consultants and national correspondents of WGMS via correspondence. On 12-13 October 1995, a two-day workshop with invited experts from various countries and representatives of sponsoring agencies was held at the Federal Institute of Technology (ETH), Zurich, Switzerland, in order to complete the final editing of the manuscript and, especially, to formulate recommendations for the future.

The volume presented here opens with an original article written by F.-A. Forel and is followed by a selection of thematic and regional chapters. As complete coverage of all glacierized areas of the world was beyond practical possibilities, characteristic examples are given from all continents, including the special cases of the continental ice sheets. The conclusions and recommendations chapter was discussed, edited and agreed upon by the participants in the 1995 expert meeting at ETH Zurich. The Appendices contain lists of the consultants and national correspondents to WGMS as well as a list of the experts participating in the 1995 meeting and the programme.

Thanks are due to ETH Zurich and UNEP for funding the expert meeting, to UNESCO for enabling the present publication and to all authors and reviewers who contributed in one way or another. The expression of our special gratitude and admiration goes to all those individuals and national/international organizations who have initiated, planned, supported and carried through a unique task for ten decades, and hopefully into many more to come. . . .

Wilfried Haeberli, Martin Hoelzle, Stephan Suter
Zurich, November 1995

1 Les variations périodiques des glaciers

F. A. FOREL
PRÉSIDENT DE LA COMMISSION
INTERNATIONALE DES GLACIERS

Reprint of original article

COMMISSION INTERNATIONALE DES GLACIERS

LES

VARIATIONS PÉRIODIQUES

DES GLACIERS

DISCOURS PRÉLIMINAIRE

PAR

F.-A. FOREL
Président de la Commission.

Extrait des *Archives des sciences physiques et naturelles*,
t. XXXIV, p. 209.

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IMPRIMERIE AUBERT-SCHUCHARDT
REY & MALAVALLON, SUCCESSEURS
1895

LES

VARIATIONS PÉRIODIQUES DES GLACIERS

par F.-A. FOREL

président de la Commission internationale des glaciers.

Sur l'initiative de M. le capitaine Marshall Hall, F.G.S., à Parkstone, Dorset, Angleterre, le VIme Congrès international de Géologie, réuni à Zurich en août 1894, a décidé la création d'une commission chargée d'étudier les variations en grandeur des glaciers actuels, dans les diverses contrées de la terre.

Cette Commission est composée de :

MM. Dr F.-A. Forel, prof. à Morges. *président* (Suisse).

Dr Léon Du Pasquier, prof. à Neuchâtel, *secrétaire* (Suisse).

Dr Seb. Finsterwalder, prof, à Munich (Allemagne).

Dr Ed. Richter, prof. à Graz (Autriche).

Dr K.-J.-V. Steenstrup, géologue à Copenhague (Danemark et colonies).

Dr H.-F. Reid, prof. à Baltimore (États-Unis d'Amérique).

Prince Roland Bonaparte, à Paris (France).

Capitaine Marshall Hall, à Parkstone (Grande-Bretagne et colonies).

Dr Torquato Taramelli, prof. à Pavie (Italie).

P.-A. Oyen, géologue à Christiania (Norwège).

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Dr Iwan Mouchketow, géologue à St-Pétersbourg (Russie).
Dr F.-U. Svenonius, géologue à Stockholm (Suède).

La *Commission internationale des glaciers* a précisé le champ de son activité en formulant les principes suivants :

a) Chacun des membres de la Commission est compétent pour organiser, comme bon lui semble, et de la manière la plus utile, les études historiques et les observations actuelles et futures sur les glaciers, dans la région qu'il représente, et pour les publier en rapports originaux et détaillés dans une revue indigène.

b) La Commission internationale est l'organe de réception et de publication des rapports sommaires fournis par ses divers membres sur les variations en grandeur des glaciers dans les diverses contrées alpines du globe. Un rapport général sera publié chaque année dans les *Archives des sciences physiques et naturelles de Genève* par les soins du bureau de la Commission.

Pour servir d'introduction à ces rapports, le président de la Commission va exposer les faits principaux constatés dans les Alpes centrales d'Europe, qui sont la contrée glaciaire la mieux observée pendant le siècle actuel.

Et d'abord, quel est le phénomène à étudier ?

C'est l'une des apparitions les plus intéressantes et les plus grandioses que nous offre le monde des Alpes. Les glaciers varient de volume. Pendant cinq ans, dix ans, vingt ans ou plus, nous voyons, sans cause apparente, un glacier augmenter de longueur, dépasser ses limites, repousser ses moraines, parfois des moraines séculaires,

envahir des pâturages, renverser des forêts, démolir des chalets. Il semble que cette crue irrésistible, qui domine tout obstacle, va amener dans la vallée une nouvelle époque glaciaire. Mais, également sans cause apparente, nous voyons le glacier s'arrêter dans cette expansion étrange, puis diminuer, reculer, se raccourcir, et cela pendant dix ans, pendant vingt ans, pendant trente ans et plus, tellement que l'envahissement précédent étant oublié, on peut croire que le glacier va disparaître dans cette fusion progressive. Puis encore, au bout d'un certain nombre d'années, ou de lustres, cette décrue prend fin et le glacier recommence à s'allonger, et ainsi de suite. Variation périodique en longueur des glaciers d'écoulement, tel est le phénomène apparent [1].

Cette variation en longueur coïncide avec une variation de même sens dans les autres dimensions de la masse de glace; en même temps que le glacier s'allonge, il s'épaissit et il s'élargit. C'est donc une variation de volume et non pas seulement de forme. Pour simplifier, nous la désignons sous l'expression de *variation de grandeur*.

Cette variation périodique est irrégulière dans le temps et dans l'espace. Les maximums successifs sont diverse-

[1] En même temps, nous pouvons constater des variations dans l'étendue des névés, dans le nombre et l'importance des flaques de neige qui résistent à la chaleur de l'été, l'apparition ou la disparition de petits glaciers temporaires. Le phénomène que nous caractérisons par le mot *d'enneigement*, varie comme la grandeur des glaciers. Quelles sont les relations entre l'enneigement des montagnes et les variations de grandeur des glaciers ? ce sera certainement un point important à étudier; mais, pour le moment, ne compliquons pas le travail et occupons-nous seulement des glaciers proprement dits.

ment espacés; en ses crues successives, le glacier descend plus ou moins loin dans la vallée; la crue est parfois très rapide, parfois très lente; souvent un glacier reste pendant bien des années immobile et stationnaire. La variation en grandeur est parfois générale et s'étend à l'ensemble des glaciers d'une région, parfois elle n'est que partielle et n'atteint que quelques glaciers. Elle n'est pas nécessairement simultanée pour tous les glaciers; parfois une crue est bien marquée sur quelques glaciers, tandis que d'autres sont, ou stationnaires, ou en décrue. D'autres fois, tous les glaciers d'une vallée, d'un groupe de montagnes, d'une chaîne varient ensemble; tous ils s'accroissent et envahissent les vallées, ou bien tous ils s'amaigrissent et s'étiolent.

Au milieu de telles irrégularités, n'y a-t-il pas une loi, ou peut-être des lois dont l'enchevêtrement cause le désordre apparent des faits? Essayons de les dégager en contemplant non pas des exemples individuels, mais le tableau d'ensemble des variations glaciaires dans les Alpes suisses pendant le cours du siècle actuel.

En dépouillant et en critiquant les anciennes observations dispersées dans la littérature alpine des trois premiers quarts du siècle, et en y joignant les observations modernes contenues dans nos quinze rapports annuels [1], nous pouvons tracer, dans leurs grandes lignes, les allures des glaciers des Alpes centrales pendant le XIXme siècle.

[1] F.-A. Forel. Les variations périodiques des glaciers des Alpes. Rapports annuels publiés dans l'*Écho des Alpes*, XVII et XVIII, Genève, 1881 et 1882, et dans le *Jahrbuch des Schweizer Alpen Clubs*, XVIII à XXX, Berne, 1883 à 1895.

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Avant 1811, nous n'avons pas d'observations valables sur l'ensemble des glaciers suisses.

A partir de 1812, phase de crue générale qui amène, vers 1818, 1820 ou 1825, suivant les glaciers, à un état de maximum ; les glaciers ont atteint partout de très grandes dimensions; pour beaucoup, c'est la plus grande extension connue dans l'époque historique. D'après l'affirmation des auteurs, il semble que cette crue a été reconnue sur tous les glaciers ; aucune exception n'est signalée d'une manière authentique. Appelons cet épisode le maximum du premier quart du siècle.

Après ce maximum qui, ainsi que je viens de le dire, a eu lieu à des dates différentes suivant les glaciers, a commencé une décrue, mal marquée, peu générale, qui a été suivie par une crue tout aussi indécise, tellement qu'il est impossible de fixer l'époque du minimum, aussi bien pour les glaciers considérés individuellement que pour l'ensemble des glaciers des Alpes. Les variations de longueur ont abouti à un nouvel état de maximum pour bon nombre de glaciers vers 1840, 1850 et 1860, disons vers le milieu du siècle.

A partir de ce maximum, qui, pour beaucoup de glaciers, a été fixè à l'année 1855 ou 1856, phase de décrue générale très nette, très intense, très prolongée, décrue pour les glaciers qui ont eu un maximum authentique vers 1850, décrue pour ceux chez lesquels ce maximum ne s'est pas manifesté. Vers 1870, tous les glaciers des Alpes, sans exception certaine connue[1], étaient en décrue.

[1] D'après le dire d'un montagnard, le glacier de Sasso Nero, val Peccia, Tessin, ne se serait mis en décrue qu'en 1880 (F.-A. F., XIVe rapport).

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A partir de 1875, nous avons constaté les indices d'une nouvelle période. Les uns après les autres, un certain nombre de glaciers se sont mis successivement en phase de crue. Le glacier des Bossons (groupe du Mont-Blanc) a, le premier, commencé à s'allonger en 1875; en 1878, la Brenva; en 1879, le Trient et Zigiorenove, etc., etc. Le développement de cette phase de crue continue encore actuellement. Elle n'est pas générale; à côté de glaciers en crue manifeste, des glaciers, leurs voisins, continuent à décroître; tel groupe de montagnes a tous ses glaciers en crue, tel autre tous ses glaciers en décrue. Je puis caractériser cette inégalité dans la manifestation de la crue en ces termes : tous les glaciers du Mont-Blanc, la moitié de ceux du Valais, un quart de ceux de l'Oberland bernois, quelques glaciers des Alpes grisonnes et autrichiennes se sont mis en crue dans les vingt ans de 1875 à 1895; pour les autres, aucun indice de croissance n'est encore devenu apparent.

Enfin, dans les deux dernières années 1893 et 1894, quelques-uns des glaciers qui avaient fait cette poussée du dernier quart du siècle se sont mis positivement en décrue. Leur front a commencé à reculer, ils diminuent d'épaisseur; le glacier du Rhône qui, avant 1893, semblait être stationnaire et promettait une mise en crue probable, a recommencé à décroître avec une nouvelle ardeur. Il paraît que, pour ces glaciers, la phase de crue est terminée, et que la décrue a sérieusement commencé.

Ces variations s'expriment par le tableau suivant :

Crue générale, de 1811 à 1818.
Grand maximum du 1er quart du siècle, de 1818 à 1825.
Décrue ou état stationnaire, de 1818 à 1830 et 1840.
Minimum, vers.....

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Crue ou état stationnaire, de 1830 à 1850, 1860 et 1870.
Maximum du milieu du siècle, 1850, 1856, 1870.
Grande décrue générale, de 1850 et 1870, à.....
Minimum, vers.....
Petite crue de fin du siècle, de 1875 à 1893 et .....
Maximum pour quelques glaciers (?), 1893.
Petite décrue de fin du siècle (?), 1893 à .....

Tel est, résumé en quelques phrases, ce que nous savons de plus positif sur les variations des glaciers du pays montagneux le mieux étudié jusqu'à présent. Je ne crois pas qu'il soit possible, pour le moment, de faire une généralisation plus complète pour aucune autre contrée glaciaire. C'est peu de chose. Les traits de ce tableau sont peu précis. Nous pouvons cependant en tirer quelques grandes lignes.

1° Les variations des glaciers sont individuelles. Chaque glacier a ses allures spéciales ; ses phases de crue et de décrue, ses états de maximum et de minimum lui appartiennent en propre. Deux glaciers voisins, les divers glaciers d'une même vallée, d'un même groupe de montagnes n'ont pas nécessairement la même histoire.

Conclusion pratique : L'observation d'un seul glacier ne suffit pas à renseigner sur les variations de l'ensemble des glaciers d'un pays.

2° Au milieu des inégalités individuelles, des allures particulières des divers glaciers, on arrive cependant à démêler des allures générales, des variations d'ensemble des glaciers de la contrée. Cela est très bien marqué en certains temps : la grande crue du premier quart du XIXme siècle, le maximum de 1856, la grande décrue du

troisième quart du siècle, la crue locale des glaciers du Mont-Blanc dans le dernier quart du XIXme siècle. Quand tous les glaciers des Alpes suisses étaient en crue en 1818, quand tous les glaciers étaient en décrue en 1870, ils subissaient certainement des actions générales; il y avait là un phénomène d'ensemble.

Conclusion pratique : Il y a lieu d'étudier, par une généralisation convenable, les grandes allures des glaciers de l'ensemble de chaque contrée montagneuse.

3° Si j'analyse plus attentivement ces mouvements d'ensemble qui apparaissent au milieu de l'irrégularité des périodes de variation, voici comment je les apprécierai, toutes réserves faites sur la sûreté des conclusions, qui ne sont jusqu'ici appuyées que sur une ou deux répétitions du phénomène.

a) La phase de crue commence successivement, individuellement, pour chaque glacier. C'est l'un après l'autre que les divers glaciers d'un même groupe de montagnes entrent en phase d'allongement; c'est l'un après l'autre que lés divers groupes de glaciers d'une même chaîne de montagnes commencent leur période.

b) La phase de décrue, au contraire, paraît commencer avec plus de simultanéité. C'est en 1856 que la grande majorité des glaciers qui étaient en crue au milieu du siècle ont commencé à diminuer de longueur; c'est en 1893 que la petite crue de fin du siècle s'est terminée sur plusieurs glaciers.

Autrement dit : l'état de minimum semble être individuel, l'état de maximum semble présenter un caractère de simultanéité mieux marqué. (Les observations ultérieures confirmeront-elles ces indices de loi? l'avenir nous l'apprendra.)

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Conclusion pratique : C'est l'état de maximum dont il est le plus facile de préciser la date dans les périodes successives des variations glaciaires.

4° La durée des périodes est longue : elle se mesure par dizaines d'années. Des faits constatés en Suisse pendant le siècle actuel, il résulte qu'en cent ans certains glaciers ont présenté trois états de maximum, d'autres deux seulement, quelques-uns peut-être un seul. La durée moyenne d'une période (d'un état de minimum à un autre) serait, d'après cela, de plus de trente ans, de moins de cinquante. Cette durée très prolongée des périodes semble aussi résulter des observations historiques des glaciers de Grindelwald et du Vernagt, qui remontent à plusieurs siècles en arrière. Jusqu'à meilleur avis, ce sera une valeur de 30 à 50 ans que nous attribuerons à la durée de ce phénomène périodique. Une périodicité d'une telle amplitude est évidemment d'observation difficile : elle demande longueur de temps, persévérance et patience. Cette durée correspond à la durée moyenne d'une vie d'homme ; elle la dépasse peut-être. Qu'est-ce que les 15 années de nos observations méthodiques suisses pour étudier les caractères d'une oscillation dont les battements se succèdent à raison de deux à trois par siècle ? nous n'en avons eu que la moitié tout au plus d'une période.

Conclusion pratique : Préparons-nous à de la patience, de la persévérance, de la prudence dans nos conclusions.

5° Vu le petit nombre de périodes dont nous possédons des résultats positifs, il nous est impossible, pour le moment, de reconnaître s'il y a isochronisme des périodes successives, s'il y a succession identique du développe-

ment des phases de plusieurs glaciers du même groupe dans les diverses périodes. En fait de synchronisme, nous n'en avons reconnu des indices, et encore sont-ils bien faibles, que dans l'époque de maximum de quelques glaciers (maximums de 1856 et 1893).

Conclusion pratique : Il y a encore beaucoup de faits non élucidés qui seront découverts par l'observation ultérieure.

Quelle est la cause de ces variations ?

Le glacier est une masse d'eau à l'état solide, provenant des précipitations atmosphériques, neige ou givre [1]. La glace étant une substance semi fluide, de très faible fluidité, le glacier se déforme et s'écoule dans la vallée, mais avec une prodigieuse lenteur ; le glacier en apparence immobile est une masse qui, nourrie dans ses hautes régions, tend à s'accroître indéfiniment en s'allongeant, en s'épaississant, en s'élargissant. D'une autre part l'attaque par la chaleur, dans les basses régions où son écoulement l'amène, transforme la glace en eau liquide, de fluidité parfaite, qui s'évacue facilement ; le glacier en fusion se débarrasse immédiatement de ses parties liquéfiées qui sont emportées par le torrent glaciaire. Il tend à diminuer par son extrémité terminale. Deux facteurs d'action opposée régissent donc le volume du glacier : le facteur d'alimentation, le facteur de fusion.

Alimentation du glacier. Le glacier est formé par l'accumulation des couches de neige tombées sur les sommets des montagnes, neiges qui constituent les nevés et qui, se

[1] Les modifications que subissent les cristaux de neige pour se transformer en grain du glacier n'entrent pas en jeu dans les phénomènes que nous considérons ici.

transformant en glace, s'écoulent lentement dans les vallées. Plus les chutes de neige sont fortes plus le névé acquiert de l'épaisseur, plus le débit du fleuve glacé est considérable et son écoulement rapide. Si dans les variations climatiques il se produit une variation dans l'abondance des précipitations neigeuses, elle se manifestera par une variation dans le volume du glacier par le fait de son alimentation plus ou moins grande. Le volume du glacier sera en fonction directe de l'abondance des précipitations neigeuses.

Liquéfaction du glacier. En s'écoulant dans la vallée le glacier arrive dans une région où l'été est assez chaud pour que la chaleur attaque notablement la glace. Chaque année une couche plus ou moins forte de sa surface extérieure, de ses bords, de son front est transformée en eau qui s'écoule dans le torrent glaciaire. Tandis qu'il se construit dans ses régions supérieures, le glacier se détruit dans ses régions inférieures, et son épaisseur diminuant chaque année, il arrive au point où cette épaisseur est réduite à zéro et où le glacier finit. Plus la chaleur de l'été est forte, plus épaisse est la couche de glace ainsi détruite, plus puissante est ce qu'on appelle l'*ablation*. Si dans les variations climatiques il se produit une variation dans la chaleur des étés, elles se manifestera par une variation dans le volume du glacier par le fait de sa liquéfaction plus ou moins rapide. Le volume du glacier est en fonction inverse de la chaleur estivale.

Or ces deux facteurs, humidité atmosphérique et chaleur qui régissent le volume du glacier, sont variables : sans parler des variations journalières et annuelles, ils présentent une périodicité cyclique ; la moyenne de l'humidité, la moyenne de la température d'une série d'an-

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nées est tantôt plus élevée, tantôt moins élevée que la normale ; les différences individuelles très variables d'une année à l'autre laissent apparaître, si on étudie le climat par des procédés convenables, des variations périodiques plus ou moins régulières. Brückner dans son beau livre des *Klimaschwankungen* a évalué ce cycle à 35 ans environ. Si les facteurs varient indépendamment l'un de l'autre le produit varie nécessairement ; si l'alimentation et si la destruction des glaciers sont variables, le volume du glacier doit l'être aussi.

Pour que la résultante soit variable il faut que les facteurs soient indépendants l'un de l'autre ; or il est incontestable que les faits météorologiques, chaleur et humidité atmosphérique, réagissent directement l'un sur l'autre.

L'abondance de neige dépend non seulement de l'humidité relative de l'air, mais encore de la température de celui-ci. La quantité de vapeur d'eau dont l'air est capable est fonction directe de sa température. D'une autre part l'état solide des précipitations aqueuses dépend directement de la température ; au-dessus du degré zéro des thermomètres de Celsius ou de Réaumur elles ont lieu sous forme de pluie. Enfin la variabilité de la température est une condition de l'abondance des précipitations ; quand la température est constante, la vapeur d'eau reste à l'état aériforme. Donc quand l'hiver est très froid, quand il est court, quand la température y est constante les neiges sont peu abondantes ; et vice-versa un hiver peu rigoureux prolongé, et à grande variabilité de température, donnera de grandes épaisseurs de neige. D'après cela, alors même que c'est l'humidité de l'air qui est le facteur décisif de l'alimentation du glacier, l'importance

des précipitations neigeuses est sous une dépendance indirecte des faits de température.

D'une autre part, la liquéfaction de la glace est due à la chaleur. Mais l'action efficace des rayons solaires et la température de l'air qui doit agir par contact dépendent directement de la nébulosité, fait d'humidité Quand le ciel est couvert, la radiation solaire est arrêtée par la couche des nuages et la température de l'air inférieur est moins élevée. La chaleur latente dégagée par la condensation directe de la vapeur d'eau sur le corps du glacier dépend aussi de l'état d'humidité de l'air. Par conséquent, alors même que c'est la chaleur qui est le facteur décisif de la liquéfaction du glacier, celle-ci est sous la dépendance indirecte de l'état d'humidité de l'air.

Enfin l'état anémométrique, le repos ou l'agitation de l'air qui agissent puissamment soit pour amener ou écarter les nuages chargés de neige, soit pour aggraver ou modérer les faits de liquéfaction du glacier, le régime des vents est intimément lié, comme cause et comme effet aux faits de chaleur et d'humidité atmosphérique.

Chaleur, humidité, vents, ces facteurs météorologiques se pénètrent mutuellement et réagissent les uns sur les autres. Il pourrait donc se faire que par une combinaison convenable, leurs actions opposées s'annulassent et que la résultante restât constante.

Mais si ces facteurs ont les relations intimes que nous venons d'indiquer, cependant, dans leurs effets sur le volume du glacier, ils fonctionnent d'une manière très indépendante.

Les deux facteurs dont la résultante se traduit par les dimensions du glacier ont leur action dominante dans les deux saisons opposées de l'année ; le facteur alimen-

tation est dû aux précipitations neigeuses de la saison froide, le facteur liquéfaction est dû aux chaleurs de la saison chaude.

Le lieu d'activité maximale de ces deux actions est de même aussi différent. L'alimentation du glacier se fait surtout dans les hautes régions, sur les sommets et dans les névés ; la liquéfaction du glacier, dans la partie terminale de la vallée d'écoulement.

Enfin il est encore une différence importante entre les deux facteurs au point de vue du développement des réactions dans le temps. L'alimentation du glacier se fait essentiellement dans les hauts névés ; le névé s'écoule lentement dans les vallées, et c'est au bout de longues dizaines d'années que la glace, partie des hauts sommets, arrive à l'extrémité terminale du glacier. Les variations du facteur d'alimentation devront donc probablement être recherchées dans le passé, dans un temps fort éloigné du moment présent où nous constatons leur effet sur la grandeur du glacier. La liquéfaction de la glace a lieu au contraire essentiellement à l'extrémité terminale, c'est-à-dire dans les parties qui arrivent actuellement au lieu où nous étudions la variation de grandeur. L'alimentation du glacier est donc, peut-être, de réaction lointaine dans le temps, la liquéfaction, de réaction immédiate ou actuelle.

A tous ces points de vue les deux facteurs opposés qui régissent les variations en volume du glacier sont donc essentiellement différents, par leur nature, par l'époque de leur origine, par le lieu de leur action maximale et par la saison de leur activité. Ils sont absolument indépendants l'un de l'autre, et il n'est pas étonnant que leur résultante présente des caractères de grande irrégularité.

DES GLACIERS.

Quoiqu'il en soit, les facteurs, chaleur et humidité atmosphérique sont l'un et l'autre des faits météorologiques.

Les causes des variations en grandeur des glaciers doivent donc être cherchées dans les variations des conditions météorologiques. La grandeur relative des glaciers est un indice de la variation du climat.

Nous possédons donc dans le phénomène tangible, tombant directement sous l'observation des variations en grandeur des glaciers, un moyen direct de constater les variations possibles des grands facteurs météorologiques. Cela légitime l'attention du monde savant pour le phénomène que nous étudions.

Les études que la Commission internationale espère obtenir sur l'ensemble des glaciers du globe offriront un grand intérêt.

Tout d'abord les faits observés sur les glaciers, si différents par leurs dimensions et les conditions de leur existence dans les diverses régions de la terre, permettront d'établir une théorie du phénomène des variations en grandeur des glaciers et de leurs relations avec les faits météorologiques. Nous savons que cette relation est incontestable ; mais quelle est-elle ? Est-ce la chaleur, est-ce l'humidité de l'air qui est le phénomène dominant ? Nous savons que ce sont les variations périodiques dans les précipitations neigeuses et dans la chaleur estivale qui sont la cause des variations glaciaires ; mais à quelle époque devons-nous rechercher les réactions de la cause sur l'effet ? Pour la liquéfaction du glacier c'est certainement les variations actuelles de la chaleur qui sont à considérer ; mais pour l'alimentation du glacier, pour les variations

de son débit et de sa vitesse d'écoulement, sont-ce de même les variations actuelles et celles des années immédiatement antécédentes ? Ou bien sont-ce des variations éloignées dans le temps, des variations qui se sont accomplies il y a bien des dizaines d'années alors que la glace qui aujourd'hui arrive au front du glacier tombait sous forme de neige sur les hauts névés ? Cette question est difficile et la réponse n'en sera donnée que lorsque nous et nos successeurs auront accumulé de nombreuses observations faites dans des conditions différentes et soigneusement critiquées.

En second lieu, ces variations glaciaires actuelles ont un grand intérêt pour le géologue. Lorsque nous les comprendrons mieux, elles nous expliqueront peut-être ces évènements considérables de l'histoire ancienne du globe que l'on appelle la période glaciaire ou les époques glaciaires; l'envahissement étrange, simultané ou successif, à une ou plusieurs reprises, de certaines régions alpines par des glaciers immenses dont nous ne possédons plus d'analogues que dans l'*Inlandsis* du Groenland. L'étude des périodes glaciaires actuelles élucidera certainement la compréhension des périodes glaciaires de l'ère quaternaire géologiques.

Au point de vue de la météorologie générale, de la climatologie, nos variations glaciaires ont aussi un très grand intérêt. Elles se manifestent aussi bien dans les glaciers de l'Himalaya et de la Nouvelle Zélande que dans ceux de l'Alaska, du Groenland, que dans le Caucase, les Alpes scandinaves, les Pyrénées et les Alpes du centre de l'Europe; mais ces manifestations sont-elles simultanées ou alternantes ? Y a-t-il coïncidence ou opposition, ou n'y a-t-il aucun rapport entr'elles ? Cette

question est de la plus haute importance, et elle aidera, quand nous pourrons y répondre, à résoudre le problème capital posé à la météorologie générale : les variations climatiques sont elles universelles, simultanées sur l'ensemble du globe, ou bien successives dans les diverses régions ? Ce qui revient à dire : sont elles de cause extérieure à la terre, de cause cosmique si elles apparaissent simultanément sur l'ensemble du globe ou bien de cause terrestre si elles alternent et se compensent dans les diverses régions du monde ? Quand nos études auront répondu à ces trois questions préliminaires :

« Les variations glaciaires sont-elles simultanées et de même signe, ou bien n'ont-elles pas de relations entre elles :

« *a*) dans les diverses chaînes de montagne d'un même continent ? (Alpes, Pyrénées, Alpes scandinaves, par exemple) ;

« *b*) dans les diverses régions du même hémisphère au nord de l'équateur ? (par exemple glaciers Européens, glaciers Nord-Américains, glaciers Asiatiques, glaciers polaires arctiques) ;

« *c*) dans les glaciers des deux hémisphères au nord et au sud de l'équateur, glaciers arctiques d'une part, glaciers antarctiques ? (Nouvelle Zélande, Sud-Amérique, régions polaires antarctiques). »

Quand nous aurons répondu à ces trois questions préliminaires, la météorologie générale et l'étude des variations de climat y auront certainement gagné une base importante pour des déductions d'un haut intérêt.

L'œuvre scientifique de la commission internationale qui aspire à embrasser dans son activité les glaciers des Alpes, des Pyrénées, du Caucase, de la haute Asie, de la

Scandinavie, de l'Islande, de l'Amérique du Nord, du Grönland, des régions polaires arctiques, de la Nouvelle Zélande, de l'Amérique du Sud, des régions polaires antarctiques, est donc de haute utilité et nous devons l'entreprendre avec courage, avec patience. avec persévérance.

Comment mener à bonne fin cette étude dans les conditions fort différentes représentées par les diverses régions glaciaires du globe? Il est difficile de donner des règles générales, et, pour le moment, nous ne croyons pas qu'une méthode unique et uniforme soit applicable. Voici quelques uns des procédés qui ont jusqu'à présent été mis en jeu pour l'étude des variations glaciaires; nous n'indiquerons pas les variantes de méthode qui peuvent différer suivant les conditions locales:

1. *Méthode du glacier du Rhône*, exécutée par les ingénieurs du Bureau topographique fédéral pour le compte du Club Alpin Suisse et de la Société helvétique des sciences naturelles. Chaque année, au commencement de septembre on lève le plan de la langue du glacier, et l'on mesure la superficie du terrain mis à nu par la retraite du glacier, ou envahie par celui-ci dans son avancement; cela donne les variations de la longueur. En même temps on fait un nivellement des profils en travers du glacier et du névé, suivant des alignements toujours les mêmes; cela indique les variations du volume du glacier. Enfin on mesure l'avancement annuel de repères placés chaque année sur les mêmes profils; cela donne les variations de la vitesse d'écoulement. Cette méthode est la plus complète; elle a l'inconvénient d'être fort dispendieuse.

2. *Méthode des forestiers suisses.* En avant du front du

glacier, deux repères fixes, placés sur les deux rives de la vallée, établissent une ligne de base, De cette ligne les distances de quelques points principaux situés sur le front du glacier sont mesurées chaque année au commencement de septembre et leur position est indiquée en abscisses et ordonnées. Un croquis à échelle convenable accompagne le rapport et indique les variations en longueur du glacier.

3. *Méthode photographique*, mise en jeu par M. Joseph Tairraz, à Chamonix. Chaque année, à la même saison (septembre ou octobre) une vue photographique du front du glacier est levée avec le même appareil, du même point de pose. La comparaison des vues successives montre les variations en grandeur du glacier. Ces variations sont en général trop peu accentuées pour apparaître facilement d'une année à l'autre sur des vues de front ; elles ne se constatent souvent qu'au bout de plusieurs années. Une série prolongée de ces vues de front est très instructive.

Des vues de profil de l'extrémité terminale du glacier montreraient bien plus facilement les variations de la longueur. Mais pour les glaciers à variations rapides le choix du lieu de pose serait souvent bien difficile.

La combinaison de vues de front et de vues de profil est certainement très recommandable.

4. *Cartes topographiques.* La comparaison des cartes topographiques levées à des époques différentes donne des renseignements précieux sur l'importance des variations. Malheureusement cette méthode (la seule utilisable jusqu'à présent pour les glaciers difficilement abordables, comme ceux des régions polaires) n'indique pas les dates du début et de la fin des phases, les dates du maximum ou du minimum de la longueur des glaciers. Or ce sont ces

dates qui ont le plus de valeur pour une comparaison utile du phénomène des variations considéré dans des pays différents.

5. *Observations de naturaliste.* L'aspect des moraines indique souvent avec netteté si un glacier est en crue ou en décrue. Si le glacier est en crue les moraines frontales sont refoulées, bousculées, les moraines latérales sont en contact avec le glacier; tout montre une activité croissante dans le transport des matériaux apportés par le glacier. Si le glacier est en décrue les moraines, aussi bien les frontales que les latérales, sont séparées de la glace par un espace libre plus ou moins large. A côté de ces symptômes les plus évidents de l'état du glacier, il est une foule de détails d'observation, qui aident à confirmer la certitude; ils varient avec chaque glacier et doivent être laissés à l'expérience et au tact du naturaliste.

6. *Témoignages.* En consultant les souvenirs des montagnards voisins du glacier, on obtient souvent des renseignements intéressants sur les dates critiques des variations de longueur, sur les époques du dernier maximum ou du dernier minimum. Une enquête intelligente donne souvent des résultats précieux. Il dépend du tact du naturaliste de critiquer ces témoignages, malheureusement trop souvent peu précis, de les appuyer les uns sur les autres, de les corriger les uns par les autres, et de tirer des conclusions justes et certaines de témoignages individuels qui ont tous leur part d'incertitude et d'erreur. J'ai moi-même pendant longtemps utilisé cette méthode et elle a donné des résultats certainement utiles et satisfaisants.

Ainsi que je l'ai dit, chacune de ces méthodes peut-être appliquée avec des variantes différentes suivant les temps

et les lieux. Les conditions de mensuration, d'observation et d'étude sont si différentes d'un pays à l'autre, d'un glacier à l'autre, que nous devons laisser à nos collaborateurs la plus grande indépendance pour agir pour le mieux des intérêts scientifiques qui leur sont confiés.

L'œuvre que la Commission internationale des glaciers a devant elle est grande et intéressante; elle est difficile. Abordons-la avec calme, courage et dévouement. Pour commencer, traitons le problème le plus simplement possible et bornons-nous à récolter tous les faits historiques qui peuvent nous faire connaître les variations glaciaires dans le passé[1], et à instituer des observations qui nous les fassent connaître dans le présent et dans l'avenir. Quand cette base aura été solidement établie, les questions subsidiaires de cause, d'effet, de relations avec d'autres phénomènes, les questions théoriques, etc., se présenteront tout naturellement à nos études, et nous, ou nos successeurs, les traiterons à mesure qu'elles se développeront devant nous.

Nous invoquons pour ces travaux la sympathie et la collaboration de tous les travailleurs, physiciens, naturalistes, alpinistes ou explorateurs, l'appui aussi des Académies et des gouvernements; leur concours nous est nécessaire pour mettre en train et pour mener à bonne fin la belle entreprise que nous avons reçu pour mission d'organiser. Ce concours ne nous fera pas défaut.

[1] Excellent exemple à suivre : E. Richter. Geschichte der Schwankungen der Alpengletscher. *Zeitschrift des D. u. Oe. Alpenvereins*, XXII, Wien 1891.

2 Historical evolution and operational aspects of worldwide glacier monitoring

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2.1 INTRODUCTION

Climatic trends are clearly reflected in mass and temperature changes in glaciers and permafrost. It is for this reason that perennial land-ice bodies are key parameters for climate system monitoring (Haeberli, 1990;Wood, 1988, 1990). This phenomenon, in fact, was recognized at the very beginning of internationally coordinated long-term glacier observations and is still today the main focus of corresponding programmes. In the meantime, however, the goals of international glacier monitoring have evolved and multiplied. Glacier signals as key elements for early detection of climatic change as induced by anthropogenic greenhouse forcing are receiving increased attention. Central aspects of this matter concern (Haeberli, 1995):

1. secular rates of change in energy fluxes at the Earth/atmosphere interface;
2. natural (pre-industrial) variability in these energy fluxes; and
3. possible acceleration trends of ongoing and potential future changes.

Observed glacier fluctuations contribute important information about all three aspects. In fact, glacier fluctuations in cold mountain areas result from changes in the mass and energy balance at the Earth's surface. Rates and ranges of such glacier changes can be determined quantitatively over various time intervals and expressed as corresponding energy fluxes with their long-term variability. This permits direct comparison with other effects of natural and estimated anthropogenic greenhouse forcing. In addition, glacier changes are linked to changing atmospheric conditions via important filters, such as pronounced memory and enhancement functions. As a consequence, glacier changes are among the clearest signals of ongoing warming trends existing in nature. Both indicators, glacier mass balance as the direct/undelayed signal and glacier length change as an indirect/delayed signal (Fig. 2.1), should be applied in combination for worldwide glacier and climate system monitoring (Haeberli *et al.*, 1989; Wood, 1990).

The principles applied to worldwide glacier monitoring have to be physically correct and as simple as possible. The combined fulfilment of both requirements is not trivial but fundamentally important for the survival of long-term glacier observations. The present contribution aims to summarize the historical development, briefly describe the various types of presently available data and to give some hints as to their possible interpretation.

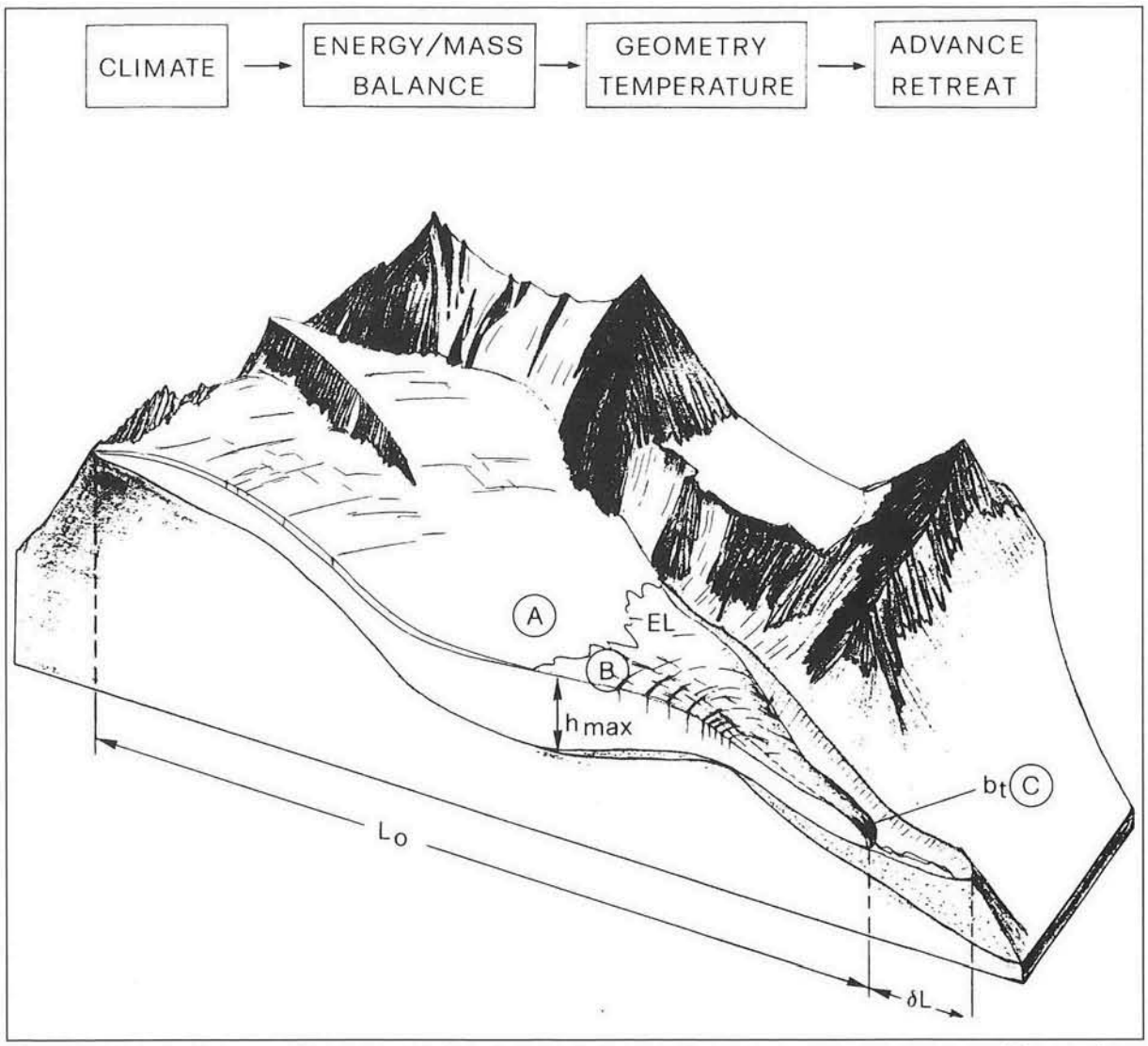

Figure 2.1 Schematic illustration of the processes linking climate and glaciers with the most important parameters used for quantifying long-term mass changes from cumulative length changes (advance/retreat of glacier tongues, cf. text for explanation of symbols; A, B, C = minimum array of stakes/pits for determination of specific mass balance with the linear balance model and repeated mapping).

2.2 HISTORICAL BACKGROUND

Worldwide collection of information about ongoing glacier changes was initiated in 1894 with the setting up of the International Glacier Commission at the 6th International Geological Congress in Zurich, Switzerland. It was hoped that long-term observation of glaciers would provide answers to two fundamental questions in climatology and earth science research at that time (Forel 1895), namely:

1. What are the mechanisms of modern variations in climate and glaciers; in particular, are glacier fluctuations globally uniform, synchronous and, hence, caused by some extra-terrestrial influences or are they regionally variable, asynchronous and therefore brought about by terrestrial conditions?
2. How should one understand the dramatic processes of the geologically most recent past, the Ice Age, the recognition of which significantly contributed to the breakthrough that represents the theory of evolution?

Over the years 1895 to 1913, the various presidents of the Commission published annual reports on glacier variations in several regions of the world (Forel and Du Pasquier, 1896, 1897; Richter, 1898, 1899, 1900; Finsterwalder and Muret, 1901, 1902, 1903; Reid and Muret, 1904, 1905, 1906; Brückner and Muret, 1907, 1909, 1910, 1911*a*, 1911*b*; Rabot and Muret, 1912; Rabot and Mercanton, 1913; Hamberg and Mercanton, 1913). The reports were written in French, German, Italian and English, and gave mainly qualitative information (advance/retreat). Results of quantitative measurements were presented for a limited number of cases only. The most detailed data were available for glaciers in Europe (Alps and Scandinavia), information from the southern hemisphere remaining extremely rare. A general tendency towards glacier shrinkage was recorded, as well as some spectacular changes, such as the rapid advance of Vernagtferner in the Austrian Alps or the catastrophic disappearance of ice in Glacier Bay (Alaska) – the largest retreat of glaciers ever directly observed by man.

During and between the two World Wars, the reports now prepared by the secretary of the newly-

formed Commission of Snow and Ice[1] (later the International Commission on Snow and Ice, ICSI) became thinner and were issued at longer intervals (Mercanton, 1930, 1934, 1936). They contained numerical data on length changes in glaciers in the European Alps and in Scandinavia, as well as references to various interesting national reports. Matthes (1934), for instance, gave a whole series of length variation measurements of Nisqually Glacier since 1857 in his US report for 1931–32. Signs of shrinking and glacier retreat clearly predominated, with the exception of a short, but marked, advance of glaciers in the European Alps around 1920. The reports which followed (Mercanton, 1948, 1952, 1954, 1958, 1961) continued to give numerical values on length changes in glaciers in the Alps, Scandinavia and Iceland. Glacier retreat still clearly predominated there during this period.

Within the framework of the International Hydrological Decade (IHD, 1965-1974) initiated by IAHS (ICSI), strong emphasis was placed on combined observations of water, ice and energy balances of glacierized basins. The specific role of glaciology and the methods of these investigations were demonstrated by Hoinkes (1968, 1970) and guidelines concerning data acquisition were published by the United Nations Educational, Scientific and Cultural Organization (UNESCO, 1970–73). When comparing the glaciological, hydrological and mapping methods of assessing glacier mass change, it was shown by Tangborn *et al.* (1975) that discharge from glacierized basins for a particular year might not reflect the actual glacier mass changes as a result of liquid storage changes within the glacier. When observations over longer periods are considered, however, it can be demonstrated that this combined effort of assessing the water, ice and energy balance provides a valuable contribution to the understanding of the hydrological cycle in high alpine regions (e.g., Glen, 1982; Reinwarth and Oerter, 1988).

In 1967, the Permanent Service on the Fluctuations of Glaciers (PSFG) was established as one of the services of the Federation of Astronomical and Geophysical Services (FAGS) of the International Council of Scientific Unions (ICSU). This resulted in publication of the *Fluctuations of Glaciers* series at 5-year intervals. In Volume I (IAHS(ICSI)/UNESCO, 1967), mass balance data from various countries, including the USSR, USA and Canada, were published for the first time, thus forming the essential link between climatic fluctuations and glacier-length changes. In the second volume (IAHS (ICSI)/UNESCO, 1973), length variation data – showing signs of intermittent glacier advance – from the USA and the USSR, as well as from other countries, complemented the corresponding records from the European Alps, Scandinavia and Iceland, where most glaciers continued to retreat. The third and fourth volumes (IAHS(ICSI)/UNESCO, 1977, IAHS (ICSI)/UNESCO, 1985) witnessed a major step towards standardization and computer-based processing of data. Length variation data from the southern hemisphere (Argentina, Kenya, New Zealand, Peru and others) were now also regularly included. Glacier readvances were reported from various parts of the world, especially from the European Alps where mass balances had been predominantly positive since the mid-1960s. For the first time, therefore, empirical information started to become available about glacier reactions to well-documented, strong signals in mass-balance history.

The *World Glacier Inventory* (WGI) was planned as a snapshot of ice conditions on Earth during the second half of the 20th century. Within the framework of the Global Environment Monitoring System (GEMS) of the United Nations Environment Programme (UNEP), a Temporary Technical Secretariat (TTS/WGI) started operations in 1976 as another service of ICSI. Instructions and guidelines for the compilation of standardized glacier inventory data were developed by UNESCO (1970) and later updated by IAHS(ICSI)/UNEP/UNESCO (1977, 1978, 1983). Detailed and preliminary regional inventories were made all over the world to update earlier compilations (especially Field, 1975; Mercer, 1967) and to form a modern statistical base for the geography of glaciers (Fig. 2.2). From the very beginning, inventory data were formatted to allow computerized data processing (cf. IAHS, 1980).

The year 1986 saw the start of the new World Glacier Monitoring Service (WGMS), combining the former two ICSI services (PSFG and TTS/WGI). The importance of glacier fluctuation and inventory data as key variables in climate system monitoring, as a basis for hydrological modelling with respect to possible CO_2- effects (US Department of Energy, 1985) and as fundamental information in glaciology, glacial geomorphology and quaternary geology had become indisputable. The tasks of the WGMS are to:

1. complete and continually upgrade a global inventory of perennial surface ice masses;
2. continue collecting and publishing standardized glacier fluctuation data at 5-year intervals;
3. publish results of mass balance measurements from selected reference glaciers at about 2-year intervals;
4. include satellite observation of remote glaciers in order to reach global coverage;
5. periodically assess ongoing changes.

This work is being carried out under the auspices of ICSI, FAGS, UNESCO and GEMS/UNEP. Data from WGMS flow into the World Data Center for Glaciology (WDC) and the Global Resources Information Database (GRID) of GEMS. Volumes V (1980–1985) and VI (1985–1990) of *Fluctuations of*

1. The Commission was set up by the International Association of Scientific Hydrology (IASH), later to be renamed the International Association of Hydrological Sciences (IAHS).

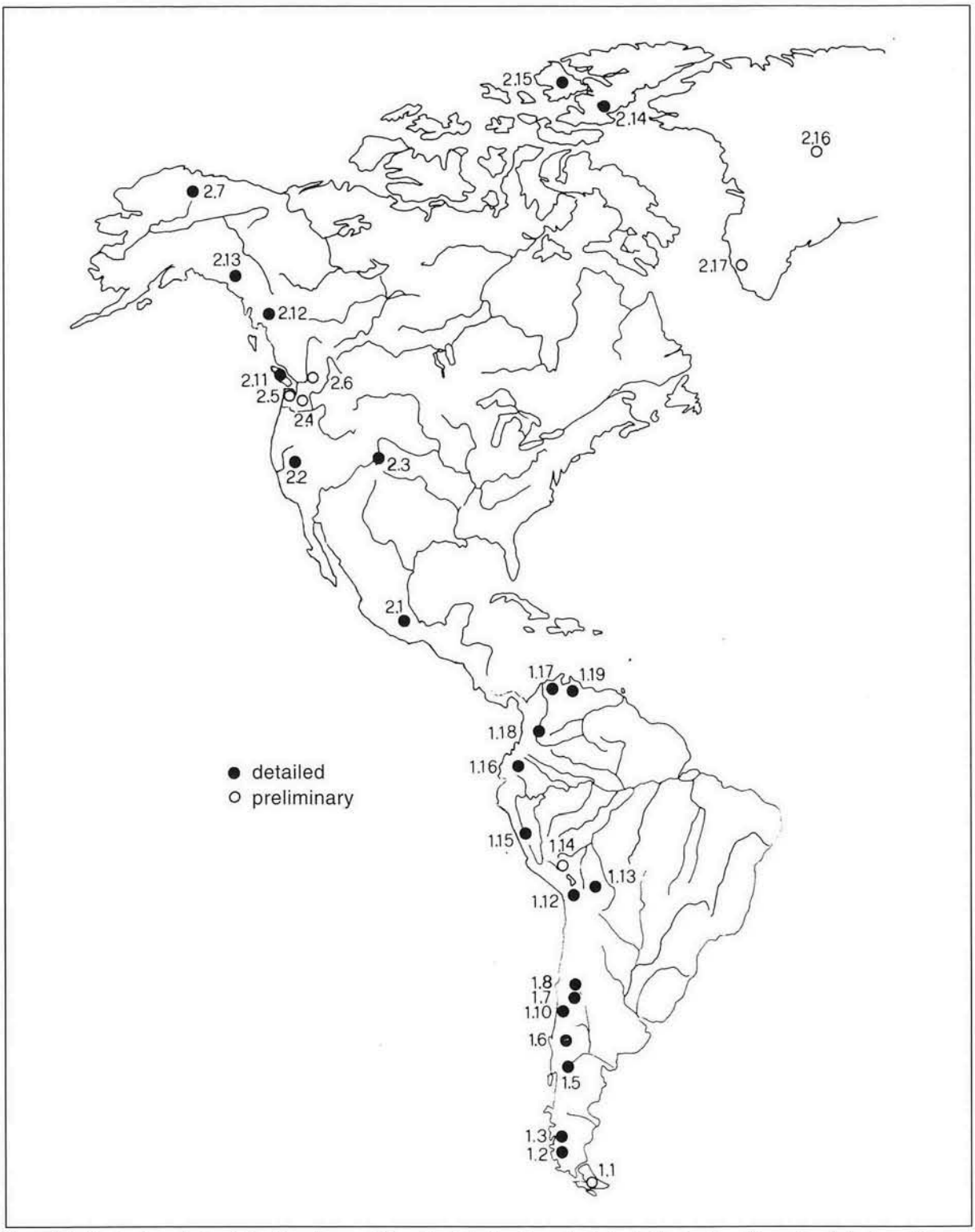

Figure 2.2 Glacier inventories of the Americas and Greenland. Full and empty circles indicate where the regional glacier catalogues are detailed or preliminary, respectively; numbers refer to the listings in the 1988 status report on the *World Glacier Inventory* (IAHS(ICSI)/UNEP/UNESCO, 1989).

Glaciers (IAHS(ICSI)/UNEP/UNESCO, 1988, 1993*a* and a *World Glacier Inventory* status report in 1988 (IAHS (ICSI)/UNEP/UNESCO, 1989) were published. In addition, a biennial *Glacier Mass Balance Bulletin* is being issued (IAHS(ICSI)/UNEP/UNESCO, 1991, 1993*b*, 1994) and a popular brochure has been prepared on glaciers and the environment (UNEP, 1992). For the fifth volume of *Fluctuations of Glaciers*, an effort was made to collect and publish internationally short abstracts on special events, such as glacier surges, ice avalanches, glacier floods or debris flows, drastic retreats of tidal glaciers, rock slides onto glaciers and glacier-volcano interactions. The biennial *Glacier Mass Balance Bulletin* was designed to speed up and facilitate access to information on mass balances of selected reference glaciers. By making the results more easily understandable through the use of graphic representation rather

than purely numerical data, the *Bulletin* complements the *Fluctuations* series where the full collection of digital data, including the more numerous observations of glacier length change, can be found.

2.3 OBSERVATIONAL PROGRAMME

Collection of standardized glacier fluctuation data today follows recommendations published by UNESCO (1969, 1970/73) and regularly updated instructions for submission of data for the *Fluctuations of Glaciers* series. Data include general information on observed glaciers, data on length variation, mass balance, area/ volume/thickness change, availability of hydrometeorological data and reports on special events. Large-scale glacier maps are also being systematically included.

Glacier *mass balance* is the direct, undelayed signal of climatic change. In ablation areas and areas of temperate firn (which predominate at lower latitudes/altitudes and in regions with humid climatic conditions), atmospheric warming mainly causes changes in mass and geometry of glaciers. An assumed step change (δ) in equilibrium line altitude (ELA) induces an immediate step change in specific mass balance (b = total mass change divided by glacier area). The resulting change in specific mass balance (δb) is the product of the shift in equilibrium line altitude (δELA) and the gradient of mass balance with altitude (db/dH) as weighed by the distribution of glacier surface area with altitude (hypsometry). The hypsometry represents the local/individual or topographic part of the glacier sensitivity, whereas the mass balance gradient mainly reflects the regional or climatic part. As the mass balance gradient tends to increase with increasing humidity (Golubev and Kotlyakov, 1978; Haeberli, 1983; Kuhn, 1981), the sensitivity of glacier mass balance with respect to changes in equilibrium line altitude is generally much higher in areas with humid/maritime climatic conditions than in areas with dry/continental conditions (Oerlemans, 1993*a*). It is therefore not astonishing that principal-component analysis of spatio-temporal distribution patterns classifies long-term glacier mass balances first of all according to 'continentality' (Letréguilly and Reynaud, 1990) and that glaciers can gain mass as a reaction to increased precipitation despite a simultaneously rising air temperature (Mayo and Trabant, 1984). Cumulative mass changes lead to ice thickness changes which, in turn, exert positive feedback on mass balance and at the same time influence the dynamic redistribution of mass by glacier flow.

Long-term mass balance measurements optimally combine the geodetic/photogrammetric with the direct glaciological method in order to determine changes in volume/mass of entire glaciers (repeated mapping) with high temporal resolution (annual measurements at stakes and pits). The primary goals of such systematic observations are usually to:

1. determine the annual specific balance as a regional signal;
2. understand in detail the processes of energy and mass exchange at glacier surfaces.

The annual specific balance as a regional signal can be obtained most economically using geodetic/photogrammetric volume change determinations repeated at intervals of from several years to a few decades (Table 2.1), with or without annual observations on a minimum of three strategically selected index stakes (Fig. 2.1): two stakes should be monitored near the equilibrium line and at some suitable site in the accumulation area where the surface area is most extended and one near the glacier front to determine ablation gradients and to quantitatively interpret length changes over extended time periods as explained below. Data interpretation can be made by applying a simplified version of the linear balance model (Reynaud *et al.*, 1986), which assumes the mass balance variation at each point of the glacier to be proportional to the mass balance variation of the entire glacier. This concept is an important working tool building on the basic experience that the spatial distribution of mass balance often remains highly similar from year to year: the temporal variability in db/dh remains small close to the average equilibrium line altitude where surface area and, hence, the influence on the overall mass balance of a glacier, is largest. The third stake recommended for a minimum stake network should be installed at the glacier terminus in order to check on the reliability of the linear balance model and to introduce adequate corrections if necessary (cf. Kuhn, 1984; Oerlemans and Hoogendorn, 1989).

TABLE 2.1 Geodetically/photogrammetrically determined secular mass balances of alpine glaciers

Glacier	*Observation Period*	*Coordinates*	*Median Elevation (m a.s.l.)*	*Surface Area (km²)*	*b (m w.e./a)*
Rhone	1882–1987	4637/0824	2,940	17.38	–0.25
Vernagt	1889–1979	4653/1049	3,228	9.55	–0.19
Guslar	1889–1979	4651/1048	3,143	3.01	–0.26
N. Schnee	1892–1979	4725/1059	2,690	0.39	–0.35
S. Schnee	1892–1979	4724/1058	2,604	0.18	–0.57
Hintereis	1894–1979	4648/1046	3,050	9.70	–0.41

Sources: Chen and Funk, 1990; Finsterwalder and Rentsch, 1980; IAHS(ICSI)/ UNEP/UNESCO, 1988.

Mass balance studies for improving understanding of the process with respect to energy and mass fluxes at glacier surfaces require extensive stake networks to be maintained and seasonally observed. Even with high densities of stakes and pits, the absolute values of volume/mass change must be carefully calibrated by repeated geodetic/photogrammetric mapping because the representativity of the monitored (stake-/pit-) network with respect to the entire glacier can otherwise not easily be assessed: especially crevassed areas, with their enlarged surfaces, tend to escape the direct glaciological analysis. Process-oriented mass balance observations are, thus, expensive and time-consuming. As a consequence, they should concentrate on characteristic effects of climatic variability. Mass balance gradients and their temporal changes under conditions of maritime/continental, tropical/polar climates, etc., as well as their long-term evolution with potential climatic changes, are of primary interest with respect to two-dimensional considerations and models (Oerlemans, 1993*b*). The three-dimensional distribution of mass balance patterns as a function of energy balance components such as snowfall, snow redistribution, solar radiation, sensible heat flux, etc. are investigated with digital terrain models and corresponding calculations of solar radiation, air temperature, etc. (Escher-Vetter, 1985; Funk, 1985). An ulti-

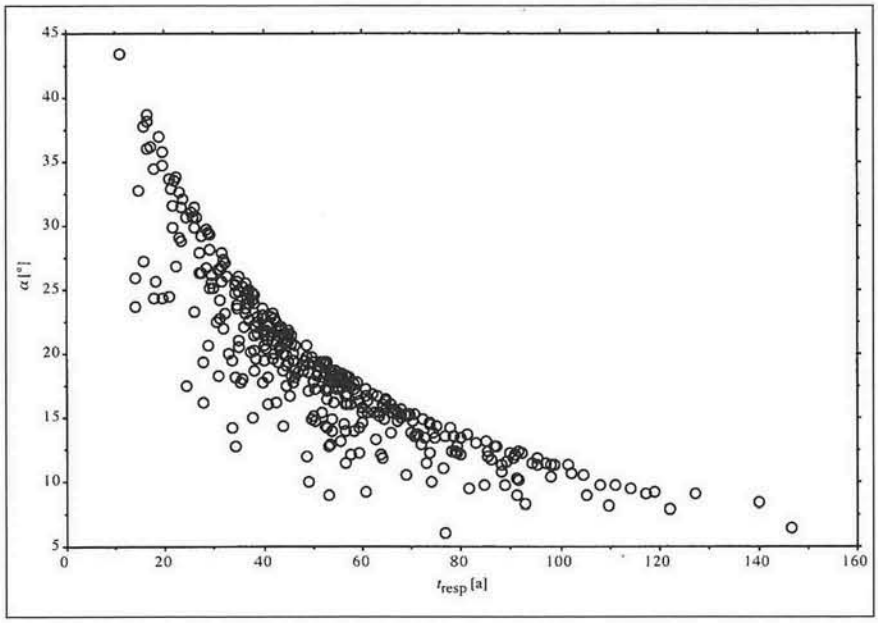

Figure 2.3 Response times t_{resp} as a function of average surface slope α for alpine glaciers longer than 2 km (from Haeberli and Hoelzle, 1995).

TABLE 2.2a Comparison of measured secular glacier mass changes (ca. 1890, 1920–1980) with estimates from cumulative length change

Glacier	*Rhone Glacier*	*Hintereisferner*
measured average b (m w.e./a)	−0.25	−0.41
total length today (km)	10.0	7.7
estimated maximum thickness today (m)	500	400
ablation at snout (m w.e./a)	5.5	5.0
estimated response time (a)	90	80
length change (km)	1.0	1.2
inferred balance change ∂b (m w.e./a)	−0.55	−0.78
inferred average b (m w.e./a)	−0.28	−0.39

TABLE 2.2b Secular glacier mass loss of Great Aletsch Glacier estimated from cumulative length change

maximum thickness today (m)	900
ablation at snout (m w.e./a)	12
estimated response time (a)	75
total length today (km)	24
length change (km, ca. 1915–1990)	1.5
inferred balance change db (m w.e./a)	−0.75
inferred average b (m w.e./a)	−0.38

Sources: Aellen, 1979; Funk, 1985; Haeberli and Hoelzle, 1995; IAHS(ICSI)/UNEP/UNESCO, 1988, 1991, 1993*b*; IAHS(ICSI)/UNESCO, 1985; and unpublished data of VAW/ETH, Zurich.

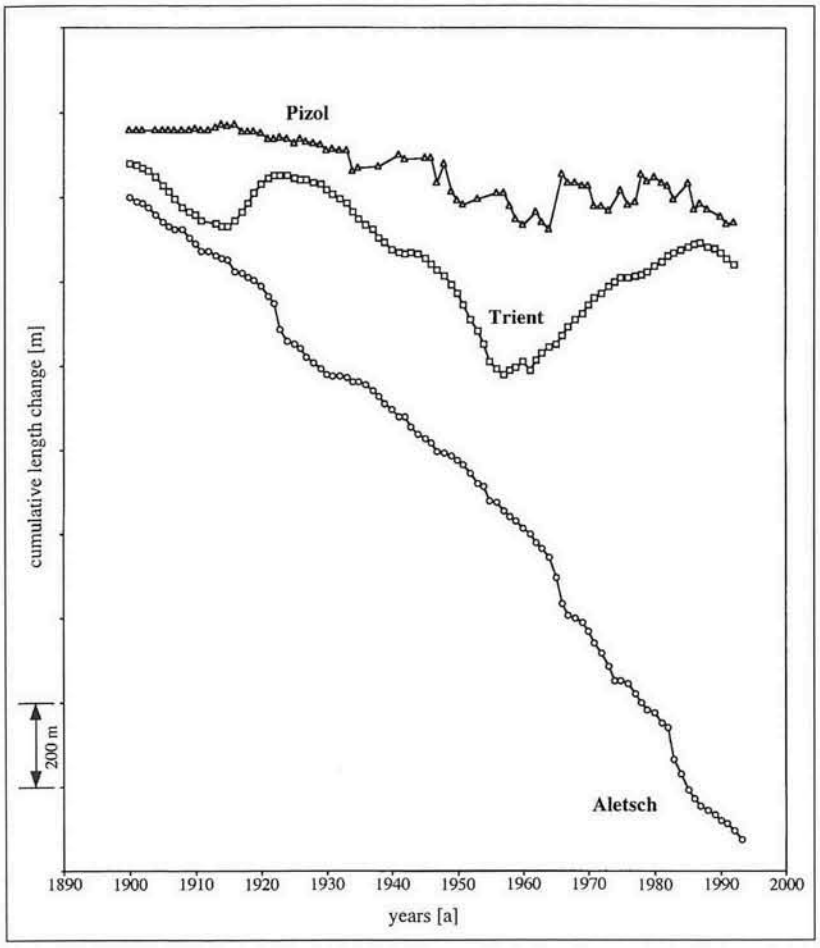

Figure 2.4 Cumulative length changes since 1900 of three characteristic glacier types in the Swiss Alps. Small cirque glaciers, such as Pizol Glacier, have low basal shear stresses and directly respond to annual mass balance and snow-line variability through deposition/melting of snow/firn at the glacier margin. Medium-sized mountain glaciers, such as Trient Glacier, flow under high basal shear stresses and dynamically react to decadal mass balance variations in a delayed and strongly smoothed manner. Large valley glaciers, such as Aletsch Glacier, may be too long to dynamically react to decadal mass balance variations but exhibit strong signals of secular developments. Taking into consideration the whole spectrum of glacier response characteristics gives the best information on secular, decadal and annual developments. Source: Kasser *et al.*, 1986, IAHS(ICSI)/UNEP/UNESCO, 1988, 1993*a*, Aellen and Herren, 1994 and Aellen, written communication.

mate goal of such investigations is to parametrize unmeasured glaciers and, thus, to better describe ongoing changes on a worldwide scale.

The complex chain of dynamic processes linking glacier mass balance and *length changes* is at present numerically simulated for only a few individual glaciers studied in great detail (cf., for instance, Kruss, 1983; Oerlemans, 1988; Oerlemans and Fortuin, 1992; Greuell, 1992). Most complications, however, disappear if the time intervals analysed are sufficiently long. After a certain reaction time (t_r) following a change in mass balance, the length of a glacier (L_o) will start changing and finally reach a new equilibrium ($L_o+\delta L$) after the response time (t_a). After full response, continuity requires that (Nye, 1960)

$$\delta L = L_o * \delta b / b_t \qquad (2\text{-}1)$$

with b_t = (annual) ablation at the glacier terminus. This means that, for a given change in mass balance, the length change is a function of the original length of a glacier and that the change in mass balance of a glacier can be quantitatively inferred from the easily observed length change and from estimates of b_t as a function of ELA and db/dH. The response time (t_a) of a glacier is related to the ratio between its maximum thickness (h_{max}) and its annual ablation at the terminus (Johannesson *et al.*, 1989)

$$t_a = h_{max} / b_t \qquad (2\text{-}2)$$

Corresponding values for alpine glaciers are typically several decades (Fig. 2.3). During the response time, the mass balance b will adjust to zero again so that the average mass balance <b> is close to 1/2 * δb. Secular glacier mass changes estimated in this way agree well with the few measured long-term mass balance series existing in the European Alps (Table 2.2) – cumulative glacier length change clearly is a key phenomenon for assessing the representativity in space and time of the few measured glacier mass balances. Mechanically unstable glaciers (surges, calving), as well as heavily debris-covered ice bodies, must be excluded from direct climatic interpretation. Scatter in δL-values due to orographic complications can be reduced by averaging measurements from more than one glacier and scale effects can be taken into account by classifying glaciers according to length (Haeberli, 1990). The remarkable signal characteristics of glacier length changes immediately appear by looking at cumulative values and different size categories (Fig. 2.4):

1. the smallest, somewhat static, low-shear-stress glaciers (cirque glaciers, *glaciers réservoirs*) reflect yearly changes in climate and mass balance almost without any delay;
2. larger, dynamic, high-stress glaciers (mountain glaciers, *glaciers évacuateurs*) dynamically react to decadal variations in climatic and mass balance forcing with an enhanced amplitude after a delay of several years;

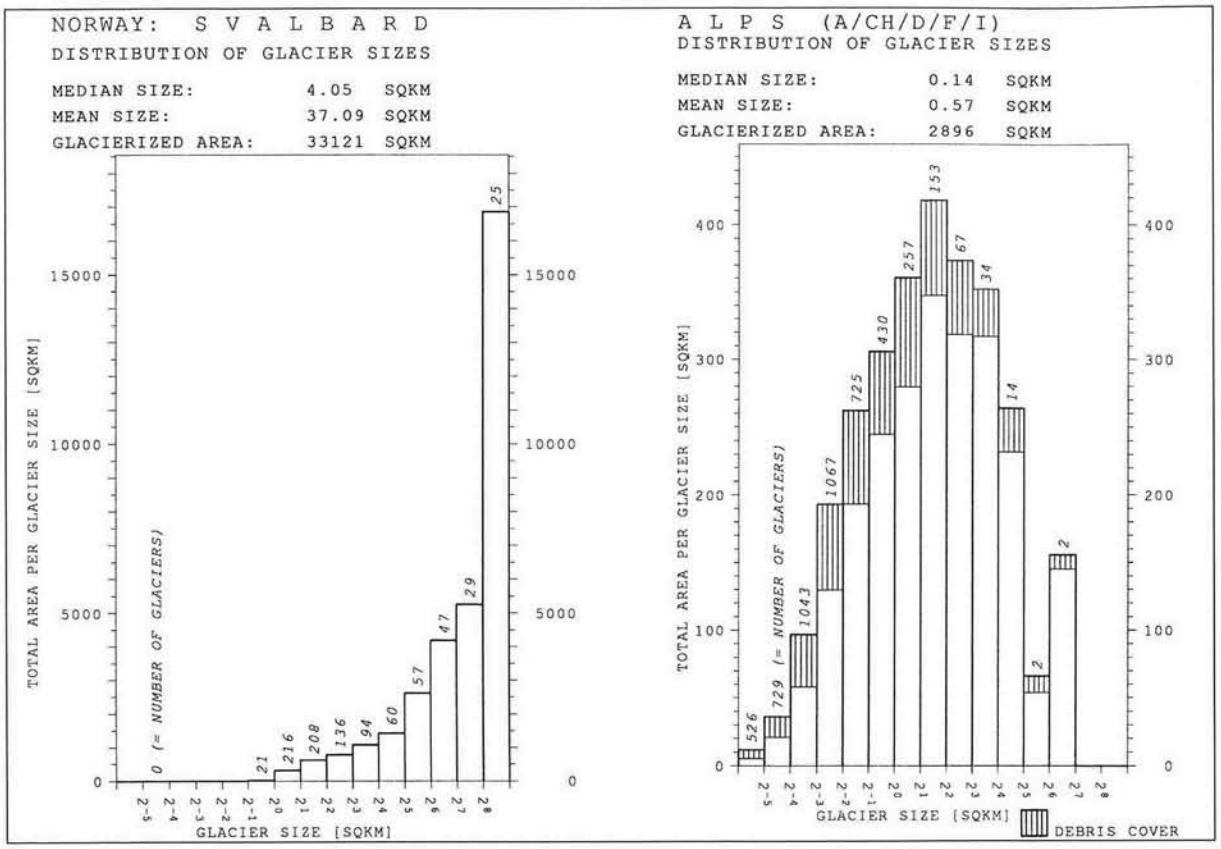

Figure 2.5 Size/area statistics from detailed inventories of glaciers on Svalbard and in the European Alps (from IAHS(ICSI)/UNEP/UNESCO, 1989).

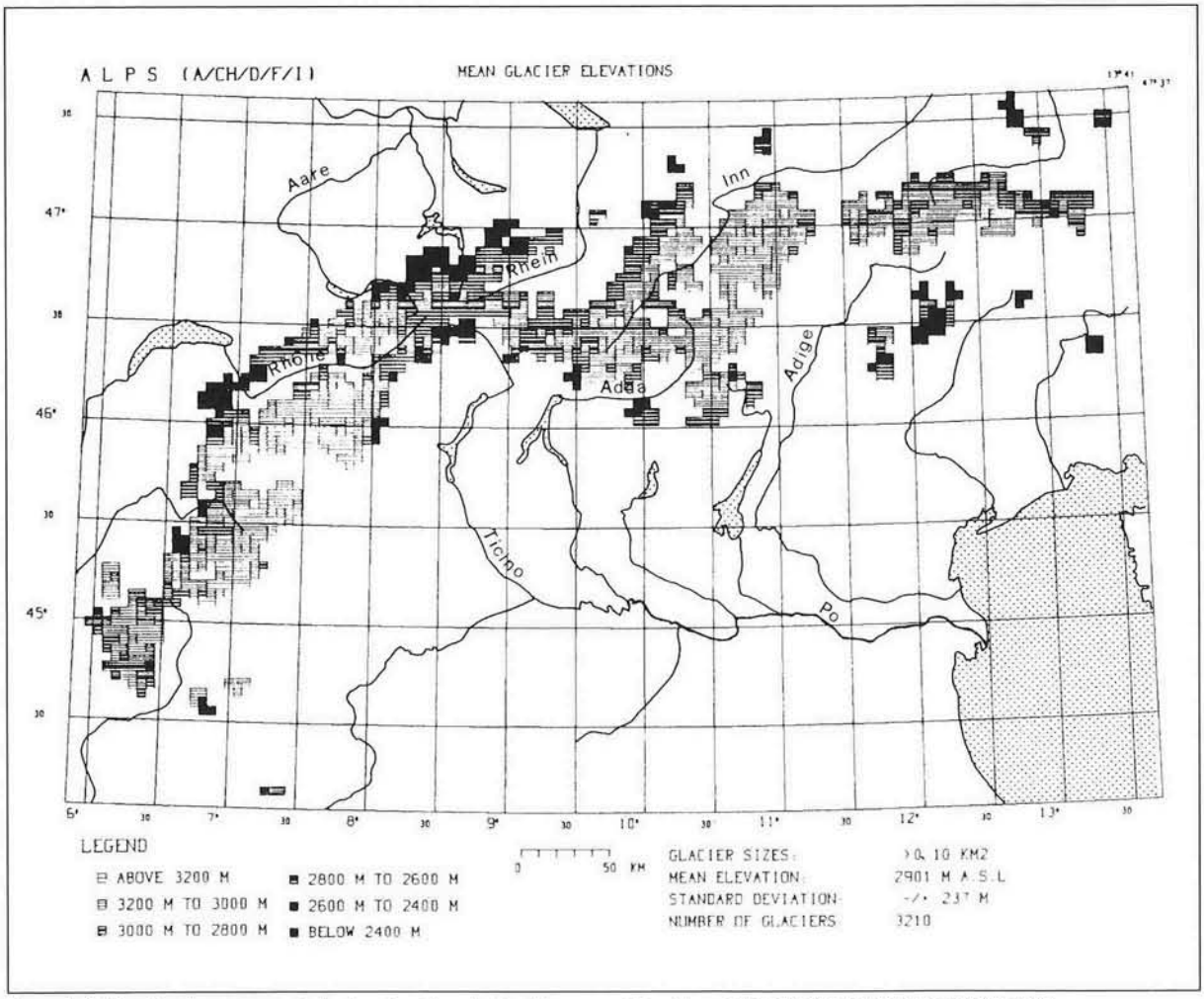

Figure 2.6 Distribution pattern of glacier elevations in the European Alps. From IAHS(ICSI)/UNEP/UNESCO, 1989.

3. the largest valley glaciers give strong and most efficiently smoothed signals of secular trends with a delay of several decades.

For the latter two size categories, the high-frequency (interannual) 'noise' is filtered out but the 'memory' of all major perennial ice bodies enables cumulation of effects for decades to centuries. The secular thickness change of a few tens of meters is thereby 'amplified' into a length change measured in hundreds to thousands of meters. The extreme clarity of this signal makes it possible to apply very simple observational methods such as, for instance, repeated tape-line readings. This, in turn, enables the cooperation of numerous non-specialists taking long-term measurements at several hundreds of glacier snouts all over the world. The quantitative and qualitative observations of secular glacier retreat in mountain ranges collected in this manner, especially at low latitudes, leave no room for doubt that climatic change causing glacier mass loss is, indeed, happening rapidly and that it is a global phenomenon. Analysis of information from remote sensing of glaciers in inaccessible regions has started in several places. (cf., for instance, Aniya *et al.*, 1992; Allison *et al.*, 1989; Kurter *et al.*, 1991; Østrem *et al.*, 1993; Swithinbank, 1988; Weidick, 1995, Zakharov, 1991). Intensive use of satellite imagery was made especially for compiling preliminary glacier inventories.

The *World Glacier Inventory - Status 1988* (IAHS (ICSI)/UNEP/UNESCO, 1989) is a guide to the existing statistical data base on the worldwide distribution and morphological characteristics of glaciers as documented in regional inventories (some detailed, others preliminary). Detailed glacier inventories include information on surface area, width, length, orientation, elevation, morphology, tongue activity, moraine characteristics and firnline positions. Repetition of this glacier inventory work is planned at time intervals comparable to characteristic dynamic response times of mountain glaciers (a few decades). This should help with analysing changes on a regional scale and with assessing the representativity of continuous measurements which can only be carried out on a small number of selected glaciers. Surface areas are basic to modelling energy balance and runoff over wide areas (Oerlemans, 1993*b*, 1994). Area statistics, for instance, show that the overall glacierized area in mountain regions with extended stream networks of valley glaciers is clearly dominated by the largest and thickest of the glaciers, whereas in the other regions it is dominated by much thinner and smaller mountain glaciers of medium size (Fig. 2.5). With regard to effects of a potential future global warming, this means that meltwater inflow to the sea from mountain glacierization of the first type – a main source of meltwater contributing to sea-level rise - would go on for a long time. In the second case, however, most glaciers could rapidly be reduced in size and many would even disappear completely within decades. Another example of glacier inventory data and their potential application is the information now available on mean glacier elevation (Fig. 2.6). This easily determined parameter is a rough approximation to equilibrium line altitude. As such, it is connected with continentality and, hence, with annual precipitation, mass balance gradient (activity index), mass turnover, englacial temperature and glacier/permafrost relations (Haeberli, 1983). Information on mean glacier elevation is, therefore, of basic importance for glaciological modelling and hydrological assessments (cf., for instance, Kotlyakov and Krenke, 1982).

2.4 REPRESENTATION AND INTERPRETATION OF COLLECTED DATA

Secular *mass balances* have been measured for six glaciers in the European Alps by repeated precision mapping since the late 19th century (Table 2.1). The average annual mass loss over the entire period varies between about 0.2–0.6 m water equivalent. Such values reflect an additional energy flux towards the Earth's surface of a few W/m^2 and, hence, roughly correspond to the estimated anthropogenic greenhouse forcing (IPCC, 1992; UNEP, 1994). The overall loss in alpine ice thickness since the end of the Little Ice Age is measured in tens of metres. Assuming typical alpine gradients of mass balance with altitude (db/dH, 0.5–1.0 m per year and 100 m altitude), the observed δb with respect to the 19th century corresponds to a shift in equilibrium line altitude (δELA) of roughly +50 m to +100 m. This result is further confirmed by directly comparing the glacier geometry at the maximum extent of the Little Ice Age with the current one (Maisch, 1988). Most, if not all, of this change can be explained by an air temperature increase of about 0.5–1°C (Kuhn, 1989; Oerlemans, 1994) as measured on average for the northern hemisphere (UNEP, 1991).

Both the annual and cumulative long-term glacier mass balances can be statistically analysed in a straightforward way. These analyses reveal considerable spatio-temporal variability over short time periods. Decadal to secular trends, on the other hand, are comparable beyond the scale of individual mountain ranges with continentality of the climate being the main classifying factor (Letréguilly and Reynaud, 1990) besides individual hypsometric effects (Furbish and Andrews, 1984; Tangborn *et al.*, 1990). Alpine glacier mass balances were strongly negative during the extremely warm decade 1980–1990. With an average value of –0.65 m water equivalent (Haeberli, 1994), the alpine ice cover may have lost about 10–20% of its total volume of about 200 km^3 as estimated for the 1970s (Haeberli and Hoelzle, 1995). The decadal average is also markedly higher than the secular average of some 0.3–0.4 m water equivalent and could possibly be an early indication of accelerating ice melt at non-polar latitudes. Extrapolating the above-mentioned average secular melt rate to all

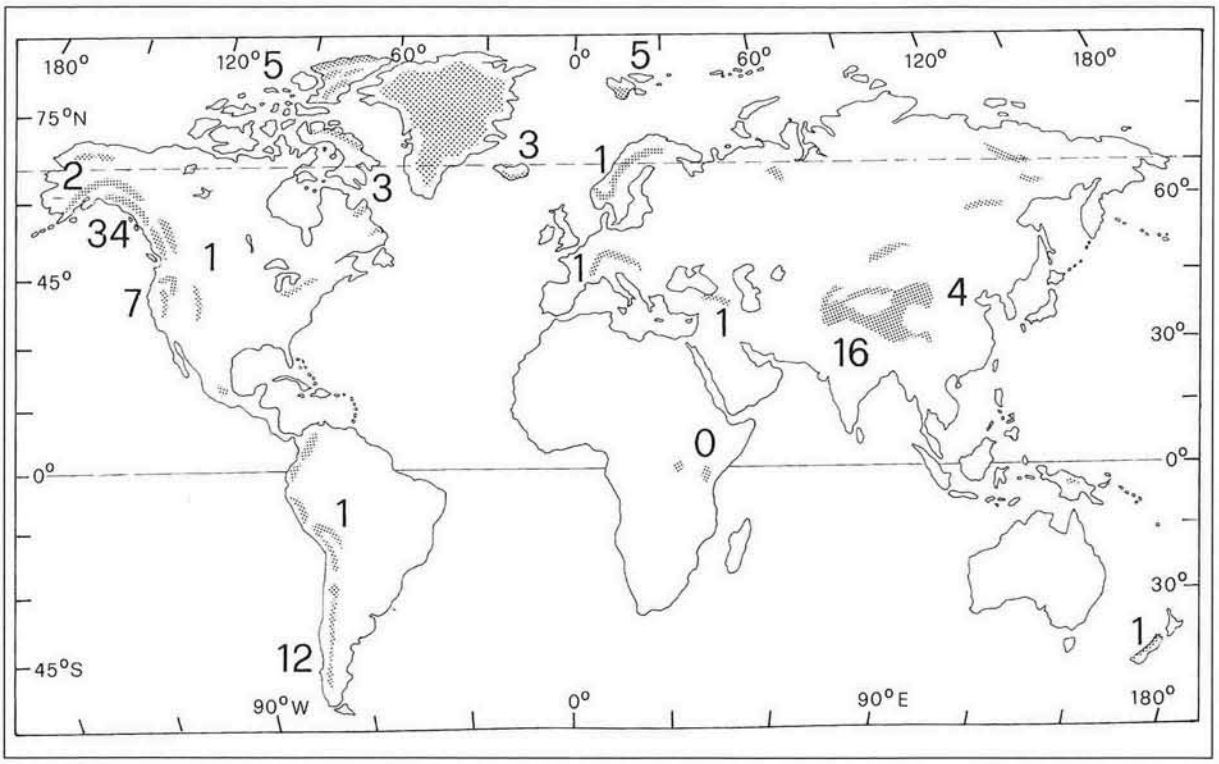

Figure 2.7 Location of glaciers (exclusive of Antarctica) showing, for each of 31 regions, the percentage-contribution to the total meltwater transferred to the ocean from glacier wastage. 0 indicates less than 0.5 percent. After USDOE (1985).

glaciers and ice caps outside the large polar ice sheets of Antarctica and Greenland gives a sea-level rise of some 5–6 cm during the 20th century. In fact, roughly one-third of the observed secular sea-level rise by 10–20 cm may be caused by the melting of mountain glaciers (IPCC, 1990). Owing to their size, high sensitivity and elevated englacial temperature, the large ice bodies around the Gulf of Alaska, in Patagonia, in the highest mountains of central Asia, in Svalbard and Iceland are the main contributors (Meier, 1984; cf. Fig. 2.7) and will continue producing meltwater throughout the 21st century (IPCC, 1992). The considerable uncertainty with such estimates relates to the problem of scaling a small number of observed values to large unmeasured areas. In firn of subzero temperatures, which predominates at polar latitudes, in regions of continental climate and at very high altitudes, atmospheric warming does not directly lead to mass loss through melting/runoff but to warming of firn layers and thereby produces corresponding signals in firn/ice temperature profiles with depth (Blatter, 1987; Haeberli and Funk, 1991; Robin, 1983). On the other hand, the most important (temperate) meltwater producers exist under very humid climatic conditions and, hence, react more sensitively to warming trends than the glaciers in the European Alps with their transitional climate. Energy balance modelling is an important tool for developing adequate scaling (Oerlemans, 1993*b*). In addition to energy balance effects at stable surface altitudes, cumulative lowering of glacier surfaces usually predating the delayed retreat of glacier tongues tends to reinforce mass losses through increased ablation. This mass-balance/altitude feedback is especially important on large and flat glaciers, which cannot dynamically adjust their length to climatic change within decadal or even secular time intervals but rather waste down in place.

Analysis of short-term variations in glacier length, i.e., for time periods shorter than the involved response times, is delicate because of the complex dynamics involved. Only glaciers with comparable geometries (especially with respect to length and slope) can be compared directly (Ding and Haeberli, in preparation). Averaging annual length changes for glaciers with highly variable geometry – either as yearly percentages of advancing and retreating glaciers or as mean annual length changes – has been historically popular. This approach, however, mixes together information from glaciers with highly variable response characteristics and suppresses the long-term memory function of glacier fluctuations (Haeberli *et al.*, 1989; Kuhn, 1978). Cumulative glacier length changes, on the other hand, can be interpreted in terms of the average mass balance during the considered time interval for time periods corresponding to the dynamic response time or to multiples of it. In the European Alps, for instance, such analyses confirm the representativity of the few secular glacier mass balances determined by repeated precision mapping (Table 2.2; cf. also Haeberli and Hoelzle, 1995). Modern cumulative glacier length changes can also be compared to ranges of pre-industrial and prehistoric variability, which are quite well documented by moraines and other geomorphic traces. Reconstruction of past glacier length changes from direct measurements, old paintings, written sources, moraines, pollen

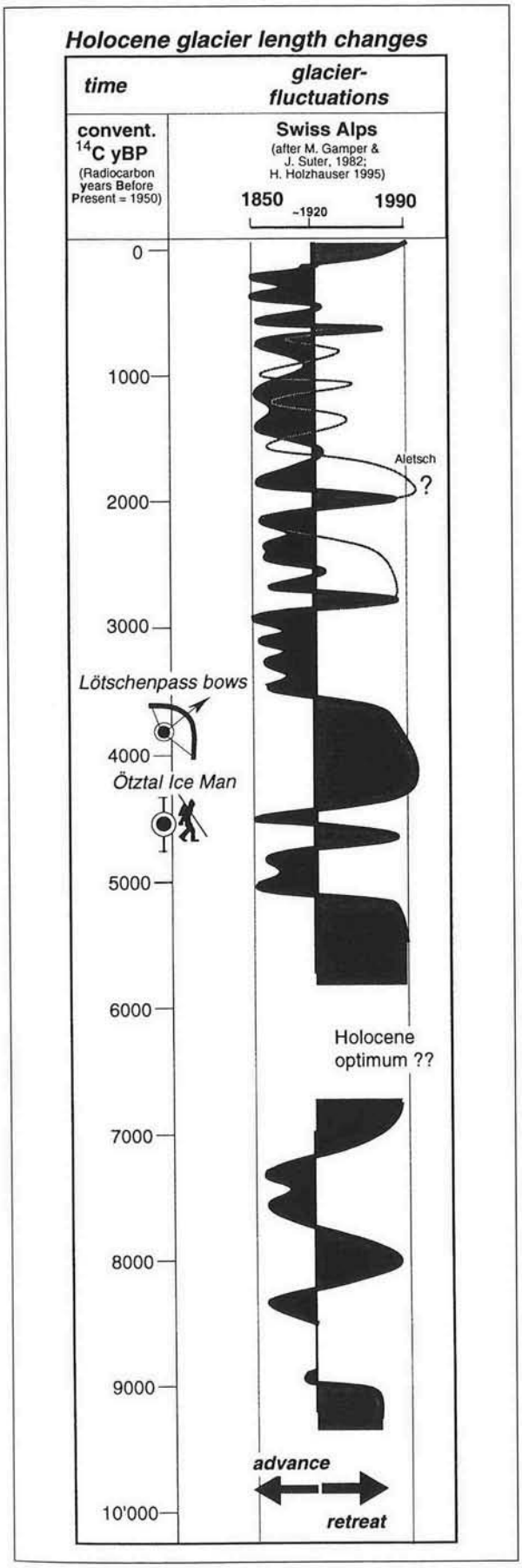

Figure 2.8 Holocene history of glacier length changes in the Alps and its relation to recent archaeological findings from melting ice in saddle configurations (Oetztal ice man/Hauslabjoch, three bows/Lötschenpass). Graph by M. Maisch, Zürich, modified from Gamper and Suter (1982) and Zumbühl and Holzhauser (1988).

analysis, tree-ring investigation, etc., indicates that, already in earlier times, glacier extent for certain periods had dwindled to today's levels and that the rates of change observed during the 20th century were probably not uncommon during the Holocene (Zumbühl and Holzhauser, 1988). On the other hand, Alpine glacier extent has varied over the past millennia within a range approximately defined by the extremes of the Little Ice Age maximum extent and today's reduced stage (Gamper and Suter, 1982; Fig. 2.8). This means that the situation seems to be evolving towards, or even beyond, the 'warm' limit of natural holocene variability.

Most recently, extraordinarily important evidence has also emerged from sites other than glacier snouts, i.e., from the top of glacier accumulation areas in the European Alps (VAW, 1993). Even at low altitudes, wind-exposed ice crests and firn/ice divides are not temperate but slightly cold and frozen to the underlying (permafrost) bedrock (Haeberli and Funk 1991). Such glaciological conditions at firn/ice divides and on permafrost (reduced heat flow through winter snow, no meltwater percolation through non-temperate ice, no basal sliding, low to zero basal shear stress) explain the perfect conservation of the 'Oetztal ice man', whose body had been buried by snow/ice in a small topographic bedrock depression on such a crest/saddle at Hauslabjoch (Austrian Alps; 3,200 m a.s.l.) more than 5,000 years ago and thereafter remained in place until it melted free in 1991. At an even lower altitude (2,700 m a.s.l.) but at a comparable site (Lötschenpass, Swiss Alps), three well-preserved wooden bows and a number of other archaeological objects were discovered as early as 1934 and 1944. Recent ^{14}C-AMS dating of the three bows gave dendro-chronologically corrected ages of around 4,000 years (Bellwald, 1992). The synchronous occurrences of mass minima on various sites of the glaciers (Fig. 2.8) confirm that glacier length variation and overall glacier mass change indeed occur simultaneously if considered on the secular time scale; the use of simple steady-state approaches is thus justified for climatically relevant time intervals. Warming periods comparable to the 20th century clearly have occurred before. The overall rate of melt since the 19th century is nevertheless very high and - as far as the European Alps are concerned - considerably exceeds average melt rates during the late Pleistocene vanishing of Ice Age glaciers (Table 2.3). The recent archaeological findings from melting ice in saddle configurations confirm that the extent of glaciers and permafrost in the Alps may be more limited today than ever before during the Upper Holocene.

Glacier inventory data can serve as a statistical basis for simulating regional aspects of effects of past and potential future climatic change. This latter application requires the introduction of a parametrization scheme using the four main geometric parameters contained in detailed inventories (length; maximum and minimum altitude along the central

flowline; surface area) and applying correspondingly simple algorithms for deriving such parameters as overall slope, mean and maximum thickness, equilibrium line altitude, mass balance at the glacier terminus, response time etc. (Fig. 2.3). A test study with the European Alps (Haeberli and Hoelzle, 1995) indicates a total Alpine glacier volume of some 130 km³ in the mid-1970s. Total loss in Alpine surface ice mass from 1850 to the mid-1970s can be estimated at about half the original value. Most of this change took place during the second half of the 19th century and the first half of the 20th century (Patzelt, 1985), i.e., in times of weak anthropogenic forcing. The short intervals of fast warming during this period may have been predominantly natural but could have included anthropogenic effects as well. An acceleration of this development, with annual mass losses of around 1 m per year or more as anticipated from IPCC scenario A for the 21st century, could eliminate major parts of the presently existing Alpine ice volume within decades. The striking sensitivity of glacierization in cold mountain areas with respect to trends in atmospheric warming is clearly revealed.

2.5 STRATEGIES AND PROSPECTS

Today's strategy of worldwide glacier monitoring is an attempt to combine various types of information. Observations of glacier length changes remain the essential key to the past and the most practicable possibility of reaching global coverage. Intercomparison of glacier length changes can be based on detailed and long-term standard curves for reasonable size categories of glaciers such as those built up in the European Alps (Haeberli, 1990). Quantitative interpretation of glacier length changes with respect to mass balance and climatic change relies on what is learned from the limited number of available detailed and long-term mass balance measurements in different climatic regions of the world (the key parameter being the mass balance gradient db/dH, cf. Kuhn, 1984). Such available long series of extensive mass balance measurements form the basis of our understanding of the climate/glacier relationship. More numerous mass balance measurements can be done using index stakes (linear balance model) combined with repeated mapping. This approach gives the mean mass balance and thus the direct glacier signal of climatic change on a regional scale. Glacier inventory data will help to assess the representativity with respect to glacier geometry and climatic sensitivity. The latter is reflected by the altitude of the equilibrium lines above a certain isotherm. Repeated mapping of glaciers as yet unexplored in various parts of the world is an aid in the documentation of the changing state of glaciers. At the same time, systematically collected reports about glacier hazards related to surges, instability of tidal glaciers, glacier floods, ice avalanches or eruptions of ice-clad volcanoes enable glaciologists to share their experiences of glacier-related hazards and natural catastrophes.

The quantitative relation between mass and length changes in glaciers over secular time scales opens up the possibility of better worldwide coverage through the application of remote-sensing techniques, ideally combined with energy balance models for more detailed quantitative analysis. Remote sensing could combine aerial photography, available in many regions since the 1950s, with high-resolution satellite imagery such as Spot, Thematic Mapper, etc. The results of energy balance modelling could be applied to mass balance gradients and ablation at the terminus for quantifying retreat and mass loss of unmeasured glaciers. In this way, (semi-) secular mass balances could be estimated for remote

Table 2.3 Alpine glacier mass changes. A comparison between the 20th century (presence P) and late-glacial time (L)

Parameter	*20th Century (P)*	*Late-glacial Time (L)*	*Ratio (P/L)*
Δt	100 a	10,000 a	1/100
$\delta T/\delta t$	+0.5 °C/100 a	+0.15 °C/100 a	3 1/3
$\delta P/\delta t$	+100 mm/100 a	+10 mm/100 a	10
L	10 km	100 km	1/10
h_{mean}	100 m	300 m	1/3
h_{max}	500 m	1000 m	1/2
u	100 m/a	10 m/a	10
b_t	5 m/a	0.5 m/a	10
t_r	15 a	1500 a	1/100
t_a	100 a	2000 a	1/20
t_r/t_a	0.15	0.75	1/5
δL	1 km	100 km	1/100
$\delta L/\delta t$	−10 m/a	−10 m/a	1
$(\delta L/\delta t)^*$	−10 m/a	−1 m/a	10
b	−0.3 m/a	−0.03 m/a	10

()* = corrected for length difference L_P/L_L

Δt = time interval considered, t = time, T = mean annual air temperature, P = annual precipitation, L = glacier length, h_{mean} = mean ice thickness, h_{max} = maximum ice thickness, u = surface flow velocity, b_t = ablation at the terminus, t_r = reaction time, t_a = response time, b = mean mass balance (from Haeberli, 1991). The indices L and P refer to Late-glacial Time and 20th Century, respectively.

areas and the global representativity of the few available direct measurements could be assessed. For this purpose, glaciers with optimal characteristics as 'climate signals' must be selected, i.e., relatively clean glaciers with adequate response times (decades), clearly defined geometry (firn/ice divide) and stable dynamics (no avalanching, surge or calving instabilities). With accelerated warming, larger glaciers would continue downwasting rather than retreating. Repeated mapping or profiling with a combination of laser altimetry and kinematic positioning using a global positioning system (GPS) would give important information in such cases, especially with regard to meltwater production and sea-level rise. Systematic application of advanced remote-sensing and modelling techniques indeed appears to be the main challenge for worldwide glacier monitoring into the 21st century.

It is now widely recognized that mass changes in glaciers are among the safest natural evidence of ongoing changes in the energy balance at the Earth's surface and, hence, that they can be considered key parameters for early detection of possible man-induced warming in the near future. Prospects would therefore appear to be good for maintaining and even expanding the existing observational network. However, problems and even dangers result from sometimes trivial and often overlooked but nevertheless fundamentally important sources, such as:

- the limited ability of university institutes to carry out long-term measurements and the understandable reluctance of governmental agencies to make long-term commitments;
- the growing financial problems, especially in developing countries and countries with changing socio-economic systems;
- the difficulty of finding scientists among the younger generation who are willing and able to carry through the work for a long time and with the necessary care;
- the still widespread overestimation with regard to the potential of modern methodology – especially satellite observation techniques – for replacing field measurements;
- the problems of applying earlier-developed techniques and terminology to low-latitude conditions, especially in tropical, monsoon-type and strongly continental areas.

There is, therefore, a most urgent need to:

- concentrate on existing, internationally coordinated long-term programmes of glacier monitoring;
- ensure continued measurement and reporting of high-quality data from glaciers with long measurement series;
- expand the existing network to the southern hemisphere;
- intensify the communication and mutual learning process between mass balance observers and analysts concerning conditions and processes in different climatic regions;
- include temperature information from deep boreholes in cold firn areas;
- develop and make operational new techniques for monitoring large glaciers;
- make clear to governmental agencies that long-term observations in glacierized regions are an important part of continuous environmental monitoring.

Glacier fluctuation data accumulated over time by generations of modest, patient and clear-sighted observers all over the world have become a real treasure of glaciology and give good reason to hope for a safe continuation of this important long-term commitment.

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3 Data management and application

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3.1 INTRODUCTION

The international database contains two different kinds of information about glaciers of the 20th century: (1) glacier inventory data describing the spatial variability and (2) glacier fluctuation data documenting changes over time.

Records on glacier fluctuations contain information about changing energy fluxes, rates of change and acceleration tendencies in connection with monitoring the evolution of the climate system. Mountain glaciers are highly sensitive, natural and globally representative indicators of the energy balance at the Earth's surface. Therefore, the measurements of glacier mass balances and length changes furnish valuable short- and long-term information about the result of energy exchange processes. As a consequence, glacier signals are good tools of early detection strategies for dealing with possible climatic change. The World Glacier Monitoring Service (WGMS) of the International Commission on Snow and Ice (ICSI/IAHS) collects and publishes worldwide standardized data as a contribution to the Global Environment Monitoring System (GEMS) of the United Nations Environment Programme (UNEP) and to the International Hydrological Programme (IHP) of the United Nations Educational, Scientific and Cultural Organization (UNESCO). These data are being made accessible in different ways, traditionally through publications and today also through electronic data access. The data appear in the following three publications of the WGMS:

3.1.1 The World Glacier Inventory

The need for a worldwide glacier inventory of existing perennial ice- and snow-masses was first considered during the International Hydrological Decade (IHD) declared by UNESCO for the period of 1965–1974. The International Commission on Snow and Ice (ICSI) of the International Association of Hydrological Sciences (IAHS) was requested by the Coordinating Council of IHD to prepare guidance material for the compilation of glacier inventory data (IAHS(ICSI)/UNEP/UNESCO, 1977). After speeding up the work in the 1980s by compiling the data sets from the different countries, an overview could be published (IAHS(ICSI)/UNEP/UNESCO, 1989). The *World Glacier Inventory* can be understood as a snapshot of the ice conditions on Earth during the second

half of the 20th century. It is planned to repeat the inventory on time scales corresponding to typical response times of the glaciers, in order to obtain valuable information about changes which can then be interpreted using parameterization models (Haeberli and Hoelzle, 1995).

3.1.2 The Fluctuations of Glaciers series

The Permanent Service on the Fluctuations of Glaciers (PSFG) was founded in 1959 (Kasser, 1970). Since 1959, mass balance and length change data have been published in the *Fluctuations of Glaciers* series. These data are compiled for the use of specialists in glaciology. In this database, long-term measurements on glaciers, mainly as time series, are stored in a standardized way. In the *Fluctuations of Glaciers* series, data can presently be found for 1,440 glaciers relating to mass balance, length variations, changes in volume, area and thickness (IAHS(ICSI)/UNESCO, 1967, 1973, 1977, 1985; IAHS(ICSI)/UNEP/UNESCO, 1988, 1993*a*).

3.1.3 The Glacier Mass Balance Bulletin

A third publication, the *Glacier Mass Balance Bulletin*, is being issued in order to (1) speed up and facilitate access to information concerning glacier mass balance and (2) make results of glacier mass balance measurements better understandable for non-specialists by using graphs, illustrations and brief conclusions/assessments relating to ongoing developments. In the *Glacier Mass Balance Bulletin*, summary information is presented for about 50 glaciers and extensive information for 10 glaciers (IAHS(ICSI)/UNEP/UNESCO, 1991, 1993*b*, 1994).

3.2 DESCRIPTION OF THE DATABASE

A new mode of data accessibility based on a database system is described on the following pages. This is not a final report on the database and its management. Rather, it is a progress report which seeks to complete a system of worldwide glacier data collection and distribution.

The World Glacier Monitoring Service (WGMS) has two databases storing two different data sources: *World Glacier Inventory* data and; *Fluctuations of Glaciers* and *Mass Balance Bulletin* data. Both data sources are now available in a database system at the ETH in Zurich on a VAX 9000 in the database system ORACLE. In the same database system, the global energy balance archive is stored at the Geographical Institute at the ETH in Zurich (Ohmura *et al.*, 1989; Ohmura and Gilgen, 1991).

In 1989, a first attempt was made to build a glaciological database with the WGMS data. With the help of this database, it was possible to publish Volume VI of *Fluctuations of Glaciers*. The database was subsequently increased by adding more tables and loading older data. In the next sections, a short description of the database structure is given.

3.2.1 Objectives of the implementation of a database system

The main objective of the database is to hold data from the different WGMS publications. Data previously stored in different filing systems and computers have been migrated from the files into the newly-built databases. The main advantages of grouping data in a database system are that it:

- reduces redundancy and inconsistency;
- increases the availability of data for all scientific users in machine-readable form;
- allows more flexibility in accessing data by selective data extraction;
- enables comfortable user handling;
- releases the user from responsibility for data managment;
- centralizes the data in one physical system;
- improves the possibilities of data checking;
- facilitates the interpretation of large data sets within a short space of time.

3.2.2 Relational database modelling

A database consists of different tables holding rows and columns. Every row contains information about an object of the real or abstract world. All the information stored in one row is called a tupel. A column (attribute) always describes the same characteristic of an object.

Since the column always describes the same characteristic, objects with different characteristics have to be in different tables. Nevertheless, one can combine information about objects from different tables if both these tables contain an attribute describing the same characteristic. Two objects are related if they have the same value for this attribute.

An object should be stored only once in a table. A double entry would be redundant: in the worst case, it would be contradictory.

The database containing glacier data consists of a table *Glacier*, which stores a single value attribute (e.g., name, latitude, longitude) for each glacier. Other attributes, for which there could be several values, have to be stored in another table. The lowest altitude of a glacier, for example, may change considerably over time owing to climatic fluctuations. Several values, therefore, exist for this attribute *lowest_elev* (= lowest elevation), which are stored in the table *Glacier_State*. This second table contains as many rows per object as there are measurements available for that object.

Tables storing such multiple occurrences of attribute values are called *dependent* tables. They only contain information about objects which exist in the object-table (there is no sense in storing information about an *unknown* object). To ensure the link to the

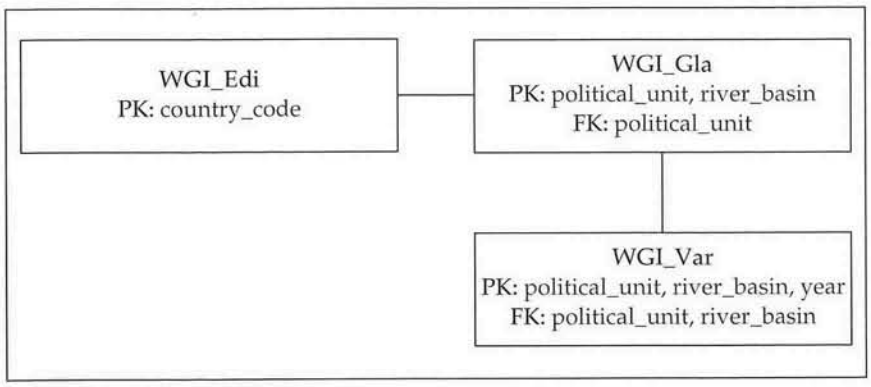

Figure 3.1 Database scheme for the World Glacier Inventory data. PK: means primary key and FK: means foreign key.

object-table, the dependent table also contains an attribute which has values uniquely identifying the appropriate object. The *Glacier_State* table, for example, contains the attribute *glacier* (= which is a number). With the help of this number, the glacier can be found in the table *Glacier* to which the lowest elevation values (*lowest_elev*) belong.

To obtain a better overview of a large amount of values stored in different tables, it is possible to create queries to the database, which delivers new tables, so-called *views*. These tables may combine information from different tables.

3.2.3 Structure of the Database

World Glacier Inventory (WGI)

The WGI database contains glacier data describing the spatial variability of the world's ice masses. This database presently consists of three tables (see Fig. 3.1). The *WGI_Gla* table contains static information about the glaciers such as location (latitude and longitude) and names. A second table *WGI_Var* stores variable information about the glaciers, such as glacier length, area and the photo-, map- or satellite-image year (see detailed list of attributes in Appendix A1). The third table contains information about the investigators of the different countries.

Fluctuations of Glaciers (FoG)

The FoG database presently consists of six tables (see Fig. 3.2). The main data from the FoG publications and the *Glacier Mass Balance Bulletin (MBB)* are stored in these tables and reference is made to the tabular data in the publication series. The data model follows a thematic grouping of the data in order to get a better overview.

The table *Glacier* contains basic data on each glacier, such as name and location (latitude and longitude). These data seldom change over time. Further on, the main differentiation is made by yearly and altitudinal changes. For example, the table *Mass_Balance_Overview* contains yearly variable mass balance information such as values for the accumulation and ablation area or the equilibrium line altitude (ELA). Mass balance data themselves can relate to the whole glacier but also to specific altitudinal ranges. Therefore, a new table *Mass_Balance* was created storing data which depend on time and on altitude.

The same idea is behind the other two tables *Glacier_State* and *Glacierpart_Variation*, too. The table *Glacier_State* contains yearly information, such as length variation (qualitative and quantitative) or highest, median and lowest elevation. In the table *Glacierpart_Variation*, altitudinal information, such as changes in volume, area and thickness, is stored

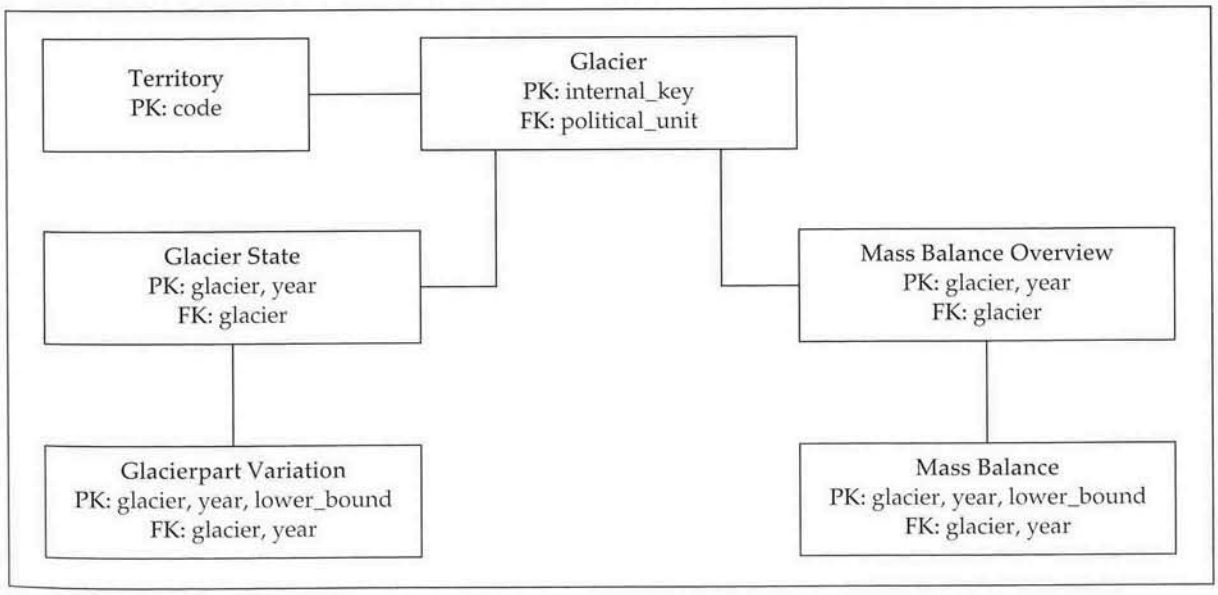

Figure 3.2 Database scheme for the Fluctuations of Glaciers data. PK: means primary key and FK: means foreign key.

additionally. The table territory contains information about country names and codes. In Appendix A1, each attribute and its meaning are given in tabular form.

3.3 EXTENT AND QUALITY OF DATA

3.3.1 Extent of data

At present, the FoG database contains data from the last four volumes of the IAHS *Fluctuations of Glaciers* series (IAHS(ICSI)/UNESCO, 1977, 1985; IAHS(ICSI)/UNEP/UNESCO, 1988, 1993*a*). The data are stored according to general information on the observed glaciers, variations in front position (glacier length changes), mass balance, mass balance versus altitude, changes in volume, area and thickness. In addition, all mass balance data from the *Glacier Mass Balance Bulletin* are available in the same database (IAHS(ICSI)/UNEP/UNESCO, 1991, 1993*b*, 1994).

The WGI database contains data on the spatial distribution of the glaciers. Only data having undergone a plausibility check (cf. Haeberli and Hoelzle, 1995) are loaded into the database. Table 3.1 shows the count of glaciers – in the WGI column for the data – which have been entered up to now. The fields marked with Y indicate that data exist but are not yet available in the database.

In addition, Table 3.1 contains the count of glaciers for each country stored in the database for the FoG and *Glacier Mass Balance Bulletin* (MBB) data.

Table 3.1 Count of glaciers for each publication series. 'Y' means that data exist but have not yet been entered into the database.

Country	*FoG (count of all data in the database)*	*MBB3 (count of the data in the publication) (IAHS(ICSI)/UNEP/UNESCO 1994)*	*WGI (count of all data in the database)*
Antarctica	39		
Argentina	10		Y
Australia	16		
Austria	142	8	925
Bolivia	2	2	
Canada	107	2	Y
Chile	36		Y
China	35	1	Y
Colombia	1		Y
France	11	2	1,130
Germany	5		5
Greenland	14		45
CIS	176	10	Y
Iceland	62	3	Y
India	5		
Indonesia	6		Y
Italy	273	2	1,376
Japan	2		Y
Kenya	13	1	Y
Mexico	2		Y
Nepal	12		130
New Zealand	4		3,154
Norway	55	17	2,998
Pakistan	37		69
Peru	10		Y
Poland	3		Y
Spain	25	1	31
Sweden	21	4	303
Switzerland	118	2	1,828
Uganda	1		Y
United Kingdom	33		
United States	197	3	Y
Venezuela	1		Y
Total	**1,474**	**58**	**11,994**

3.3.2 Data quality

The quality of the reported measurements depends on:

- the method applied
- the aim of the measurements
- the plausibility check before publication.

The applied techniques for mass balance measurements include (1) glaciological; (2) photogrammetrical; (3) hydrological and (4) index methods; and for measurements of the front variations (1) aerial or terrestrial photogrammetry and (2) geodetic ground survey (theodolite, tape, etc). The choice of the appropriate method is mainly a function of the aim of the measurement. It is therefore necessary to differentiate between process studies (local balance as a function of energy balance and topography) and long-term monitoring of regional effects (specific balance) or estimating mass balance changes in the glaciers on the basis of a measured length change together with the glacier length and estimated or measured ablation at the tongue.

The glaciological method of mass balance measurements with a dense network of stakes is the most appropriate application for process studies. In contrast, long-term monitoring of regional effects first requires a careful selection of the glacier in connection with its climatic and spatial representativity. The measurement can predominantly be based on repeated photogrammetry/surveying, which furnishes reliable data over extended periods of time. In order to increase time resolution to yearly periods, the linear balance model can be combined with repeated photogrammetry/surveying by using a set of a few index stakes: one in the accumulation area, one in the ablation area (both of them near the equilibrium line where the surface area is greatest) and one at the lower margin of the glacier. The stake at the glacier margin enables the mass balance gradient and its changes over time to be roughly assessed. Moreover, the long-term length change of the observed glacier can be quantitatively interpreted on this basis with respect to the secular mass balance which, in turn, provides information of fundamental importance for judging secular rates of change and possible acceleration tendencies.

Consistency

Checking the data before publication is of vital importance. For *Fluctuations of Glaciers* and *World Glacier Inventory* data, the use of the database now available considerably improves the possibilities for checking the consistency of the submitted data. This new tool will help to introduce a future rating system for the different data sets, enabling the scientific user to estimate the accuracy of the information.

In their current state, the databases are not free of inconsistencies. These inconsistencies can be traced to the history of the data collection and management system. The data grew in a slow and sometimes uncontrolled way and were stored in the 1970s for the first time in computer systems (at that time on punch cards).

The main problem is the identification code. Although the so-called WGI-number is a clear and good form of identification, many problems exist, especially concerning double codes. Double codes contradict the rules of the database design, which requires that data be redundancy-free, especially when the code is used as a primary key. The WGI-code is based on twelve digits. Positions 1 and 2 are reserved for the designation of the political unit. Position 3 designates the continent according to the number given in IAHS(ICSI)/UNEP/UNESCO (1989). Positions 4 to 7 are reserved for the drainage code. Positions 8 and 9 are free positions and positions 10 to 12 are reserved for local numbering of the glaciers. The now loaded WGI-data will use the WGI-code for the primary key and all double codes have to be changed therefore into a unique code. For a later loading of new WGI-data it is essential to always use the same WGI-number, otherwise it is impossible to connect the new database to the older one. Bearing in mind that the following should be read as an unbroken sequence, an example is given here of how the WGI-number should look:

CH	4	L001
Political Unit	Continent Code	Drainage Code

85	006
Free Position Code	Local Code

The FoG database also contains the WGI-code as one of the different attributes but this information still has to be adapted to the code from the WGI database in order to couple the two databases.

Homogeneity

A further problem is the homogeneity of the collected WGMS data. This problem is caused mainly by variable space and time scales of the data. For example, the same database has to contain both data from the large ice sheets in Antarctica and Greenland and data from very small perennial ice patches in Japan. It should also be noted that many time series in the FoG database are incomplete or relate to highly variable measurement intervals. Therefore, users of the database should be aware of these inhomogeneities and use the data properly, for example, by grouping the data to avoid mixing different data sources and their physical meaning.

Security

The term security refers to the data security itself and access to the database. The database receives regular back-ups on the computer at the ETH in Zurich and the data are periodically printed and widely distributed as hardcopies. The published versions of *Fluctuations of Glaciers* and the *World Glacier*

Inventory should continue to be issued at regular intervals.

Access to the database for public users is limited to the processing of queries and downloading data. Only the WGMS administrator has the right to insert and update the data in the database.

Loading of data

All data from the FoG-series (Volumes III to VI), once available on file, are now stored in the database. Volumes I and II will have to be typed in by hand in the near future.

For WGI data, plausibility checks are carried out while entering the data into the database. These checks will be made for all glaciers of each country. After this check, the data will be printed out and sent to the national correspondents with a list of the discovered errors or uncertainties for review.

3.4 APPLICATIONS

In this sub-chapter, some ideas are presented for possible applications of FoG and WGI data. These applications are based on simple parameterization schemes and on cumulative time series.

3.4.1 World Glacier Inventory

As a means of estimating basic glaciological characteristics of the inventoried ice bodies and simulating potential effects of climatic change on mountain glaciers, a parameterization scheme using simple algorithms for unmeasured glaciers (Hoelzle, 1994) can be applied to glacier inventory data. Concerning past and potential future climate scenarios, glacier changes for assumed mass balance changes are calculated as step functions between steady-state conditions for time intervals which approximately correspond to the characteristic dynamic response time (a few decades) of the glaciers concerned.

In order to test the procedure, a pilot study was carried out in the European Alps (Haeberli and Hoelzle, 1995), where detailed glacier inventories had been compiled in the mid-1970s. Glacier length changes for given disturbances in mass balance are calculated with respect to the characteristic dynamic response time (Jòhannesson, 1989) in the sense of step functions between steady-state conditions. Total glacier volume in the Alps is estimated at some 130 km^3 for the mid-1970s; strongly negative mass balances are likely to have caused a loss of about 10-20% of this total volume during the decade 1980–1990. Backward calculation of glacier length changes using a mean annual mass balance of 0.25 m water equivalent per year since the end of the Little Ice Age around 1850 AD gives considerable scatter but overall satisfactory results compared with long-term observations. The total loss in alpine surface ice mass since 1850 can be estimated at about half the original value. An acceleration of this development, with annual mass losses of around 1 metre per year or more as anticipated from IPCC scenario A for the 21st century, could eliminate major parts of the presently existing alpine ice volume within decades.

Altogether, the database from national inventories in the European Alps contains a total of 5,050 perennial surface ice bodies and refers to the time of the mid-1970s (Austria 1969, France 1967–1971, Germany 1979, Italy 1975–1984, Switzerland 1973). Only 1763, or 35%, of this total number are 'real glaciers' larger than 0.2 km^2 with complete information available on surface area, total length, maximum and minimum altitude. All these data were extracted from the WGI database. The parameterization scheme is being applied to this latter part of the sample. The remaining 3,287 (65%) ice bodies are perennial ice patches and glacierets smaller than 0.2 km^2 and are treated separately. Total surface area of all 5,050 inventoried surface ice bodies is 2,909 km^2 (IAHS(ICSI)/UNEP/UNESCO, 1989). The surface area of the 1,763 'eal glaciers' is 2,533 km^2 or 88% of the total surface area. The total volume of these 'real glaciers' is calculated as 126 km^3 and that of the 3,287 ice bodies $\leq$ 0.2km^2 as 2.6 km^3. The latter value is globally estimated by taking half (7.5 m) the mean value of h_F (=max. thickness at the flow line, 15 $\pm$ 4 m) for glaciers with 0.2 < F (area) < 0.4 km^2. Thus, the overall volume of perennial surface ice existing in the Alps around 1970 is about 100-150 km^3 with the best guess being 130 km^3. This volume corresponds to a sea-level equivalent of about 0.35 mm. Such a small value points to the limited importance for sea-level rise of glaciers in comparable high mountain areas with predominantly small glaciers but it also points to the vulnerability of these glaciers to the effects of climate: the decade 1980–1990, with a mean annual mass balance of –0.65 m water equivalent as measured on 8 regularly observed glaciers in the Alps (Careser, Gries, Hintereis, Kesselwand, Saint-Sorlin, Sarennes, Silvretta, Sonnblick; IAHS(ICSI)/UNEP/UNESCO, 1988, 1993*a*; Haeberli, 1994), may have seen a loss of surface ice volume of up to 20 km^3 or about 10-29% of the total volume existing around 1970. It is reasonable to assume that the total volume of perennial alpine surface ice at present is not much higher than 100 km^3. Comparably low total glacier volumes may have existed before 1950 AD, around 4,000 BP and around 5,000 BP; even smaller ice volumes are assumed for the early Holocene, i.e., the time period around 7,000 to 6,000 BP (Haeberli, 1994).

3.4.2 Fluctuations of Glaciers

The next example discusses some different time series of cumulative mass balances data extracted from the FoG database. The mass balance, which is the direct undelayed signal of the glacier response, gives valuable information on changes in the energy balance at the glaciers' surface. For this reason, cumulative mass balance data were extracted from

different mountain ranges all over the world where mass balance data were collected regularly. The data were then grouped by mountain ranges and graphically displayed (cf. Fig. 3.3 to 3.10). Seven of the larger and most intensively investigated mountain ranges could be used for comparison, i.e., the Altai (4 glaciers), the European Alps (11), the Arctic (6), the Caucasus (4), Scandinavia (13), the Tien Shan (11), Western America (16) and some other mountain ranges (9). On the basis of this comparison, the following statements can be made: first of all, a general tendency towards mass loss is recognizable over the past 30 to 40 years but the information looks quite different in each region. In the Altai and Caucasus, there is no remarkable trend towards a large mass loss. The Tien Shan shows a continued and undisturbed mass loss for all glaciers for which observations are available. In the European Alps, a general trend towards mass loss, with some interruptions in the mid-1960s, late 1970s and early 1980s, is observed. In Scandinavia, glaciers close to the sea have seen a very strong mass gain since the beginning of the 1970s but mass losses have occurred with the more continental glaciers. The mass loss of the latter continues up to the 1970s before being replaced by a more or less zero mass balance, with slightly positive trends in recent years. The Arctic shows a regular mass loss for the two long-term monitored glaciers in Svalbard. Western America shows a general mass loss near the coast and in the Cascade Mountains. Other glaciers in different mountain regions, such as the Rocky Mountains, in the Kamchatka and in the Pamirs, all show a decreasing tendency in cumulative mass balance. It is interesting to note the strong difference in tendency between maritime and continental glaciers in Scandinavia. That there should be mass gain in western Scandinavia could be explained by an increase in precipitation, which more than compensates for an increase in air temperature. Therefore, the principal classifying factor is the distance from the humidity source (continentality). Statistical analysis of these mass balance records confirms that the same type of spatio-temporal distribution pattern can be found for glaciers within individual mountain ranges. Secular trends are comparable beyond the scale of mountain range (Letréguilly and Reynaud, 1989, 1990).

3.5 FUTURE PERSPECTIVES

The data from the last 20 years of the FoG series and a large part of WGI data are now stored in the database at the WGMS Center in Switzerland. This database will be updated continually with new data and supplemented with older data from earlier volumes. As a further step, information on the availability of hydrometeorological data and on special events will be introduced into the database as well.

Electronic and selective access to the database greatly facilitates handling and utilization for all scientists, especially for researchers interested in

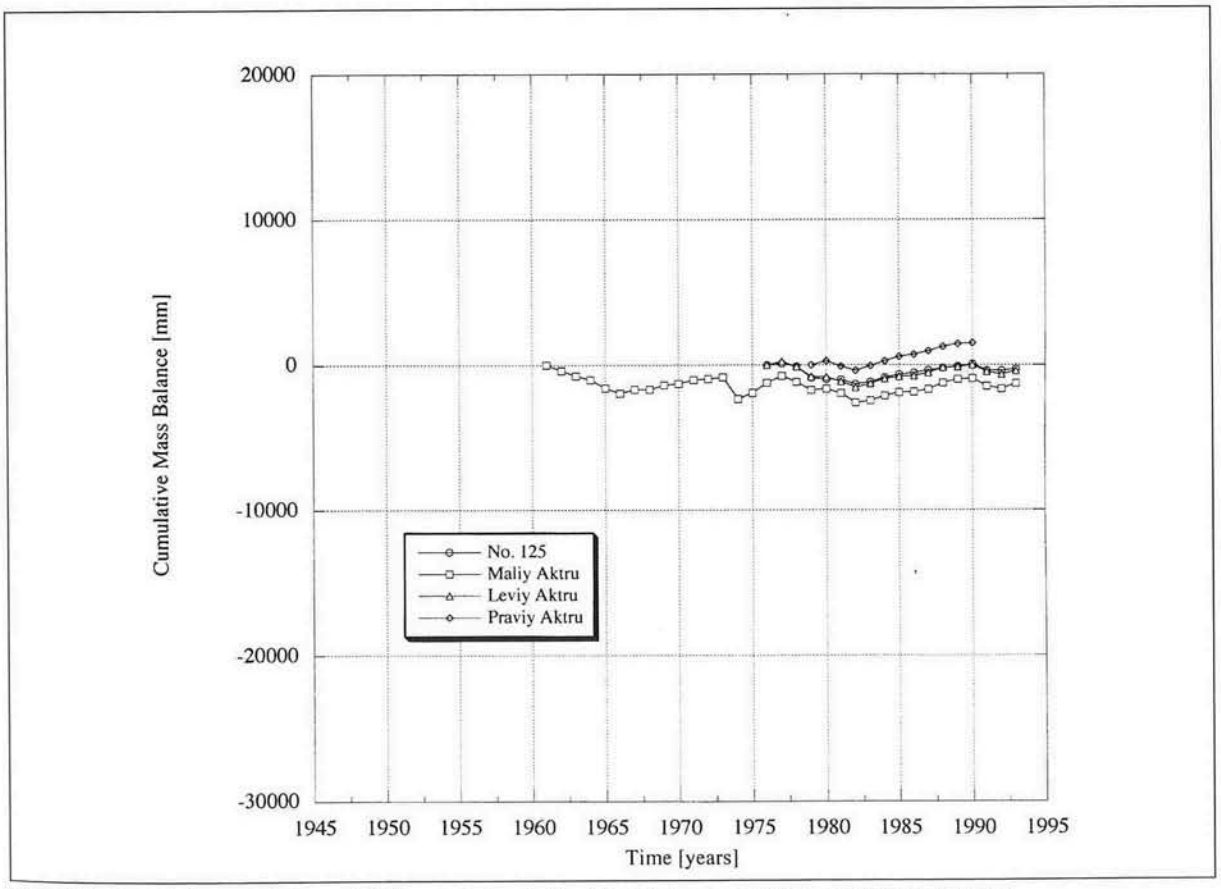

Figure 3.3 Cumulative mass balance of different glaciers in the Altai. Data extracted from the WGMS database.

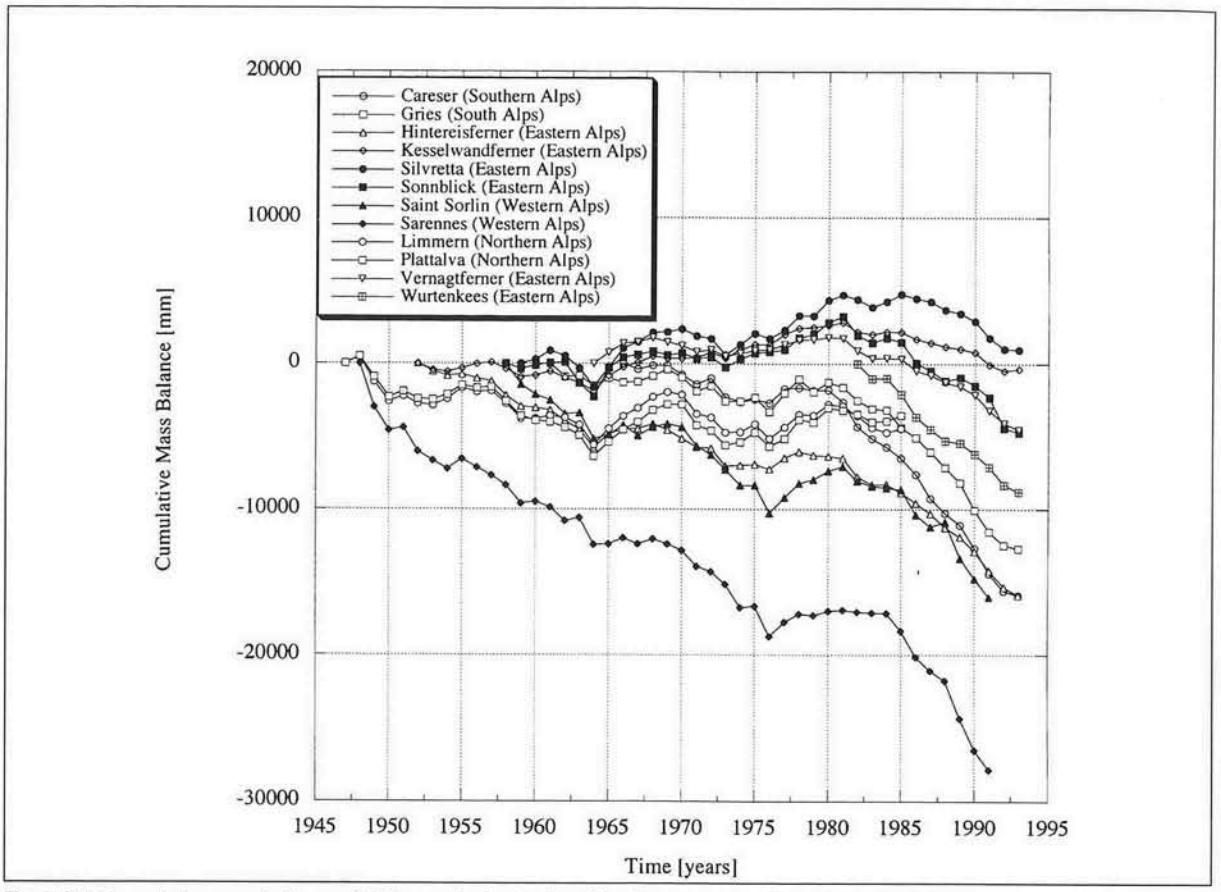

Figure 3.4 Cumulative mass balance of different glaciers in the Alps. Data extracted from the WGMS database.

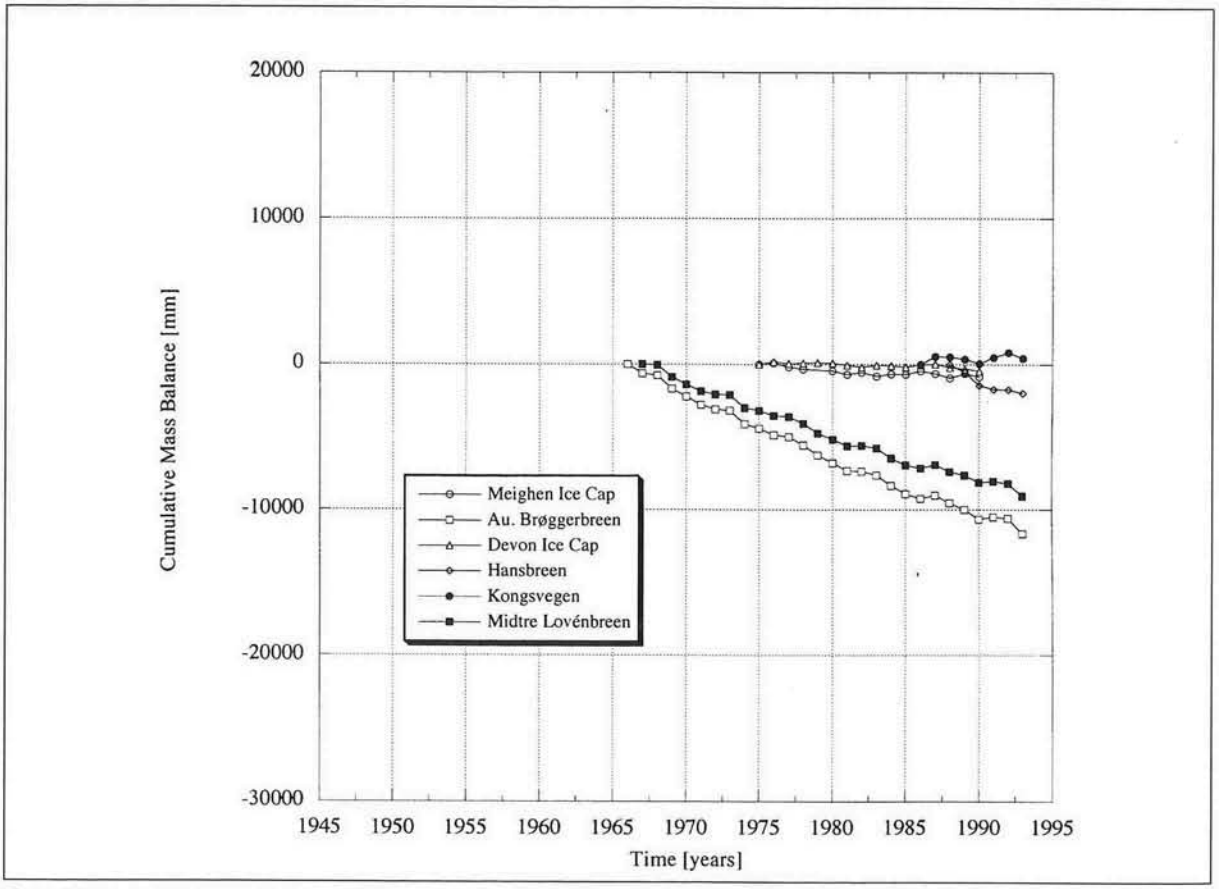

Figure 3.5 Cumulative mass balance of different glaciers in the Arctic. Data extracted from the WGMS database.

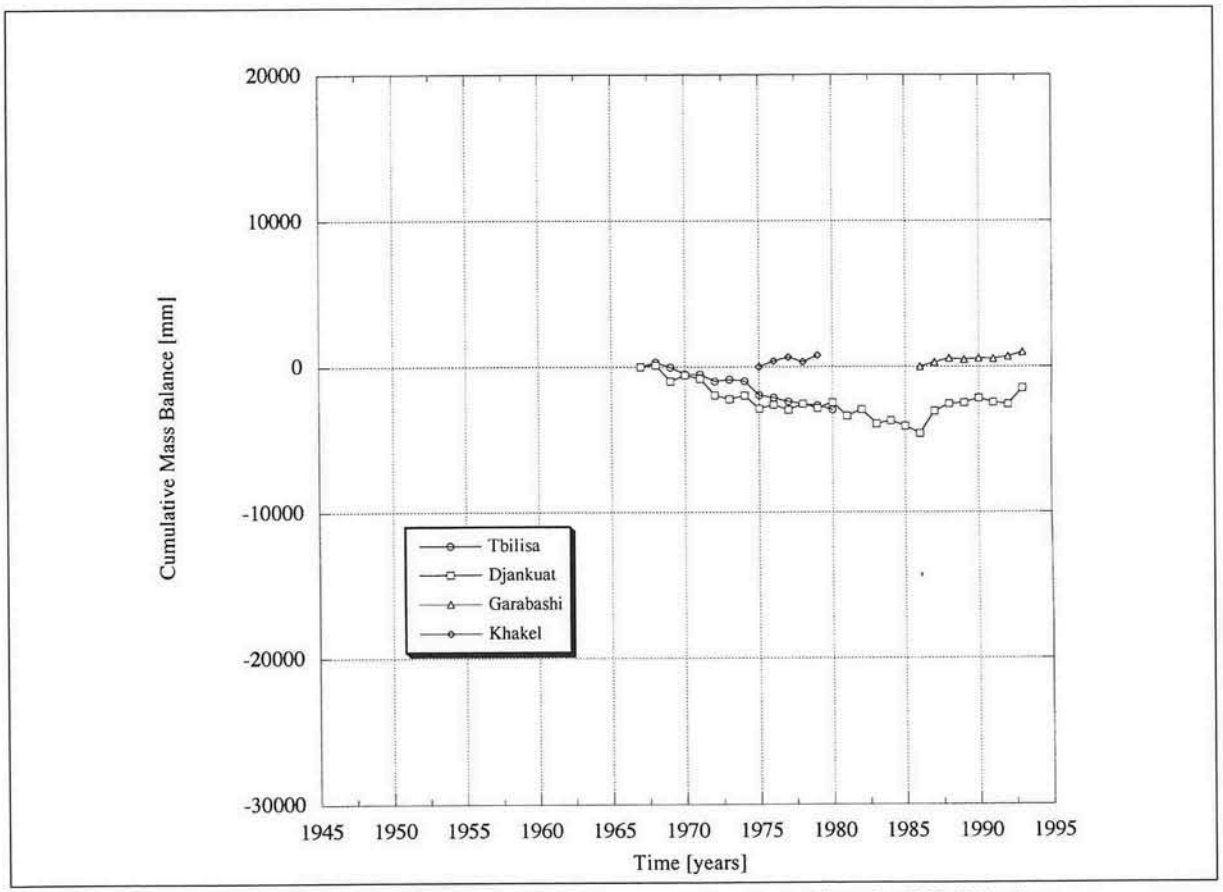

Figure 3.6 Cumulative mass balance of different glaciers in the Caucasus. Data extracted from the WGMS database.

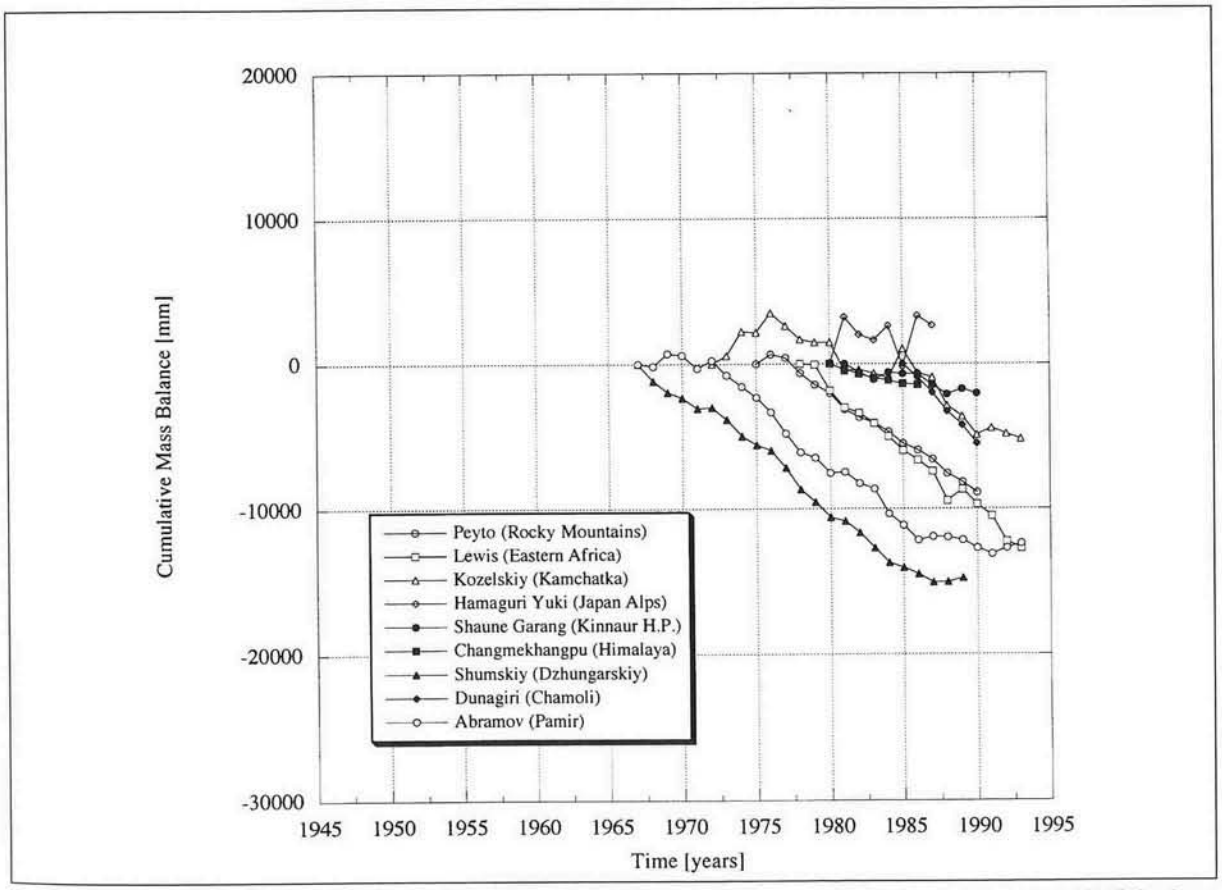

Figure 3.7 Cumulative mass balance of different glaciers in the different mountain ranges. Data extracted from the WGMS database.

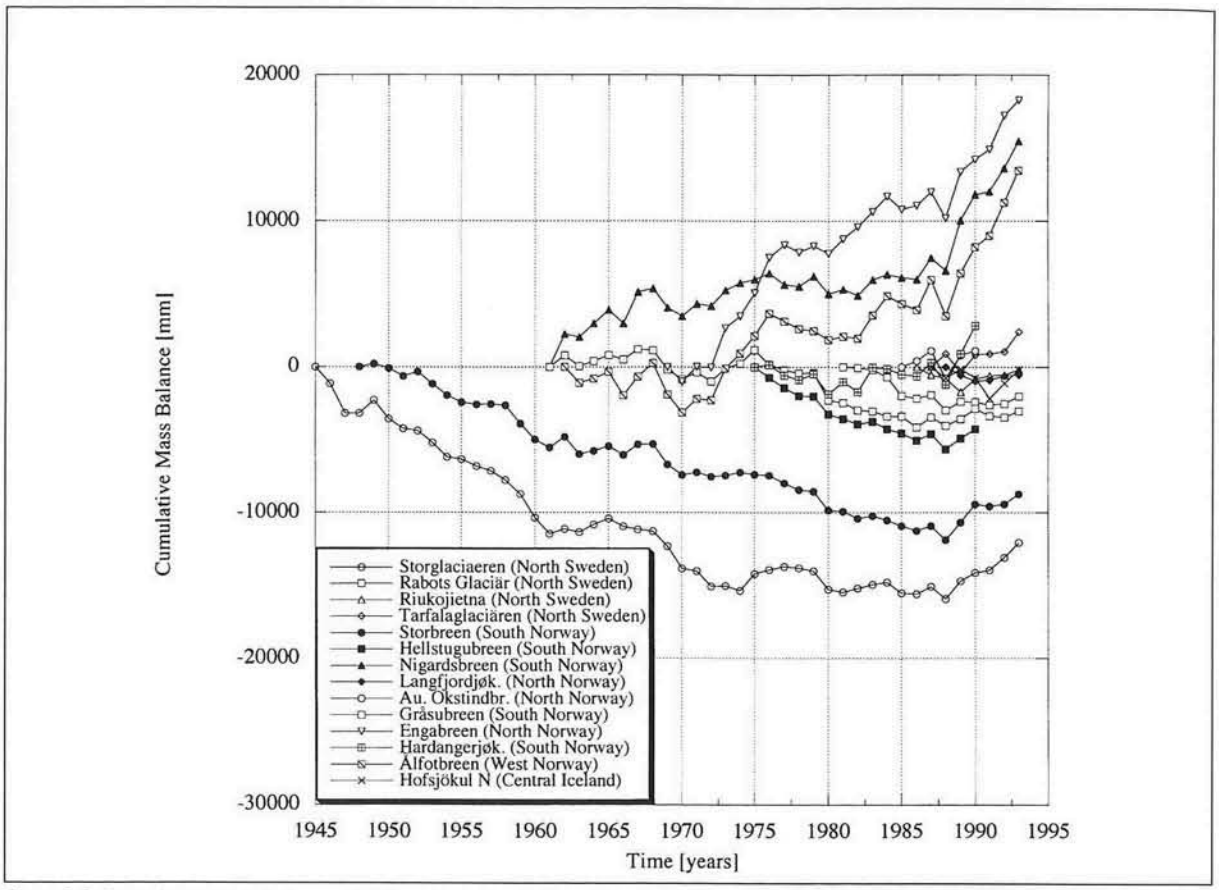

Figure 3.8 Cumulative mass balance of different glaciers in Scandinavia. Data extracted from the WGMS database.

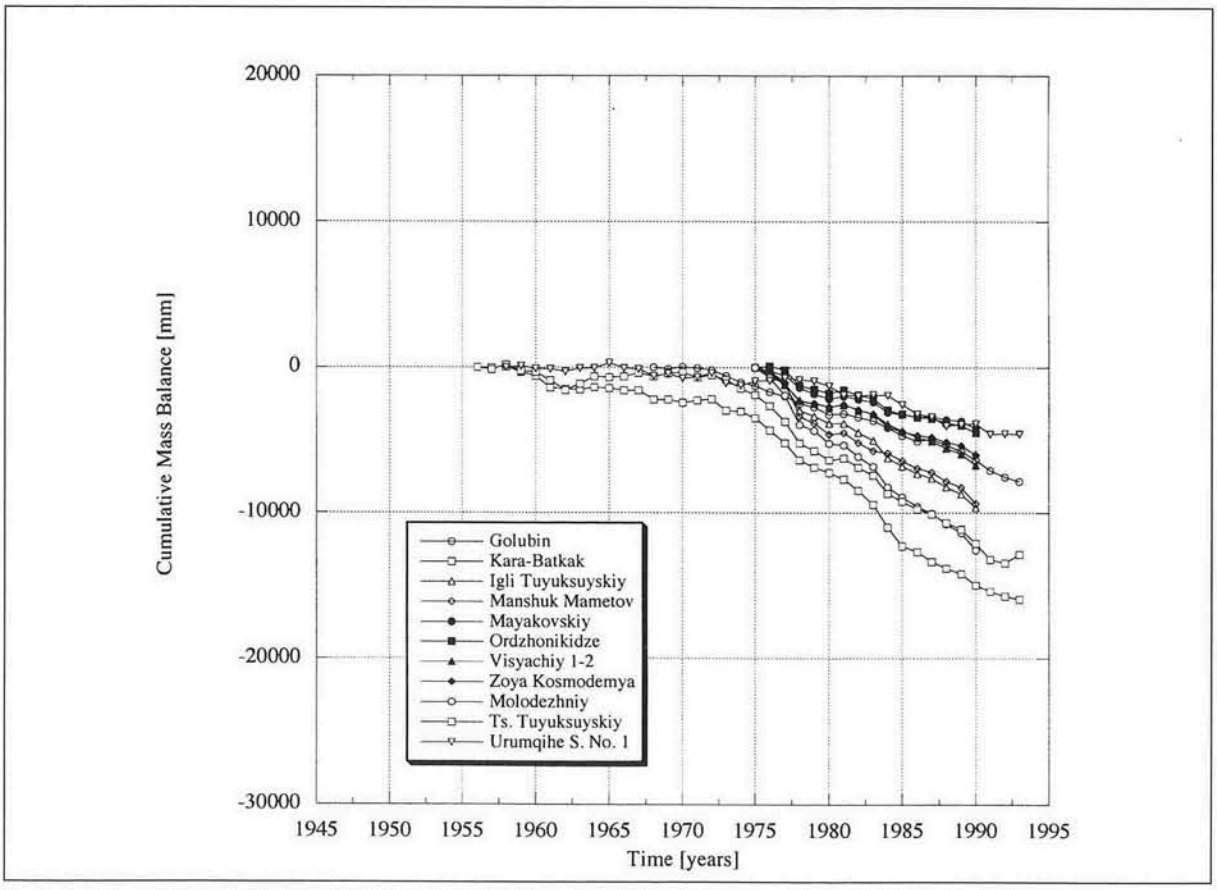

Figure 3.9 Cumulative mass balance of different glaciers in the Tien Shan. Data extracted from the WGMS database.

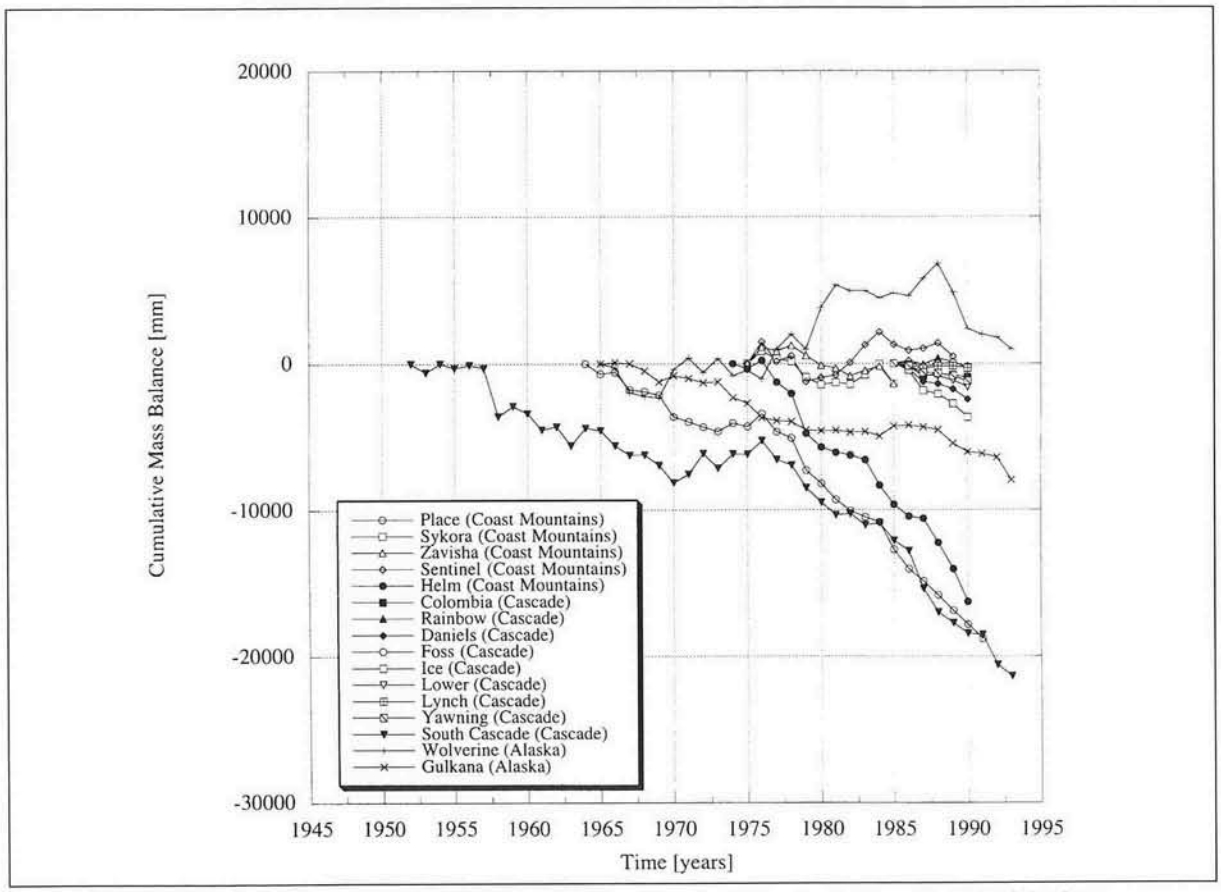

Figure 3.10 Cumulative mass balance of different glaciers in Western America. Data extracted from the WGMS database.

modelling. For this reason, it is planned to establish a World-Wide-Web site for the database, where users will easily be able to download, via Internet, the desired glacier data, such as mass-balance time series or inventories of different countries. A first step towards improved data accessibility via the World-Wide-Web was the Russian inventory by the World Data Center (WDC) in Boulder.

It is planned to improve data checking and to introduce a quality-rating system for data, especially with respect to the mass balance, length change and volume/area/thickness change data from the FoG. Better use should also be made of Geographical Information Systems for spatial representation, visualization and analysis of the database.

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A1 Appendix

(Attribute information of the database tables)
The table and attribute list for the Fluctuations of Glaciers database

Table A1.1 Glacier

Internal_Key	Number (5)	Not Null/PK	Internal Key
Political_Unit	Varchar2 (2)	Not Null/FK	Political Unit part of the WGI-number (Position 1 and 2)
River_Basin	Varchar2 (5)		Code for the determination of the hydrological catchment areas (Position 3 designates the continent, positions 4 to 7 are for the drainage code)
Free_Position	Varchar2 (2)		Freely chosen number (Positions 8 and 9)
Local_Code	Varchar2 (3)		Local Code helps to make the WGI-number clear (Positions 10 to 12)
Local_PSFG	Varchar2 (5)		This glacier number should be unique for each country.
Name	Varchar2 (15)		Name of the glacier
Gen_Location	Varchar2 (15)		Geographical location (general)
Spec_Location	Varchar2 (15)		Geographical location (more specific)
Latitude	Number (4,2)		Latitude in degrees and minutes
Card_Point_Lat	Varchar2 (1)		Cardinal point (N or S)
Longitude	Number (5,2)		Longitude in degrees and minutes
Card_Point_Lon	Varchar2 (1)		Cardinal point (E or W)
Prim_Classific	Varchar2 (1)		Primary Classification
Form	Varchar2 (1)		Form
Frontal_Chars	Varchar2 (1)		Frontal characteristics
First_Survey_FR	Date		Year of the first quantitative survey
First_Mass_Bal	Date		Year of the first mass balance survey
Expos_Acc_Area	Varchar2 (2)		Exposition of accumulation area
Expos_Abl_Area	Varchar2 (2)		Exposition of the ablation area
To_be_published	Number (4)		Publication number (Vol. I to VI)
MBB	Varchar2 (1)		Mass Balance Bulletin Key
Remarks	Varchar2 (500)		Remarks

Table A1.2 Glacier_State

Glacier	Number (5)	Not Null/PK	Reference to the table Glacier Internal_Key
Year	Number (4)	Not Null/PK	Year is together with 'Glacier' the Primary Key
Highest_elev	Number (4)		Highest Elevation
Median_elev	Number (4)		Median Elevation
Lowest_elev	Number (4)		Lowest Elevation
Snout_point_alt	Number (4)		Altitude of the snout
Error_Altitude	Number (4,1)		Estimated maximum error in altitude
Length	Number (5,2)		Length of the glacier
Variation_horiz	Number (6,1)		Variation quantitative between previous and present survey
Qualitative_var	Varchar2 (2)		Variation qualitative between previous and present survey
Error_variation	Number (4,1)		Estimated maximum error in variation
Ref_Date_State	Date		Reference Date is like Date of initial survey but will be automatically set for years between
Date_Survey	Date		Date of the survey
Method_Survey	Varchar2 (1)		Method of the survey
Ref_date_vartn	Date		
State_Published	Number (4)		
Vartn_published	Number (4)		

Table A1.3 Glacierpart_Variation

Glacier	Number (5)	Not Null/PK	Reference on table Glacier_State
Year	Number (4)	Not Null/PK	Reference on table Glacier_State
Lower_Bound	Number (4)	Not Null/PK	Lower Bound is together with Glacier and Year the Primary Key
Upper_Bound	Number (4)		
Area	Number (8,3)		Area for each altitude interval
Area_Change	Number (6)		Area change for each altitude interval
Thickness_Chg	Number (6)		Thickness change for each altitude interval
Volume_Change	Number (12)		Volume change for each altitude interval

Table A1.4 Mass_Balance_Overview

Glacier	Number (5)	Not Null/PK	Reference to table Glacier
Year	Number (4)	Not Null/PK	Year is together with 'Glacier' the Primary Key
Time_System	Varchar2 (1)		Time system
Beginn_Period	Date		Begin of survey period
End_Winter	Date		End of winter season
End_Period	Date		End of survey period
Equilibr_Ln_Alt	Number (4)		Equilibrium line altitude
Min_Sites_Acc	Number (3)		Number of minimum measurement sites in the accumulation area
Max_Sites_Acc	Number (3)		Number of maximum measurement sites in the accumulation area
Min_Sites_Abl	Number (3)		Number of minimum measurement sites in the ablation area
Max_Sites_Abl	Number (3)		Number of maximum measurement sites in the ablation area
Acc_Area	Number (8,3)		Accumulation area
Abl_Area	Number (8,3)		Ablation area
Aar	Number (4,1)		Accumulation area ratio
Rating_Ela	Varchar2 (1)		Rating of the ELA
Rating_Aar	Varchar2 (1)		Rating of the AAR
Sumry_Published	Number (4)		Year of publication (specific net balance)
Items_Published	Number (4)		Year of publication (mass balance *vs.* altitude)

Table A1.5 Mass_Balance

Glacier	Number (5)	Not Null/PK	Reference to table Mass Balance Overview
Year	Number (4)	Not Null/PK	Reference to table Mass Balance Overview
Lower_Bound	Number (4)	Not Null/PK	Lower Bound is together with Glacier and Year the Primary Key
Upper_Bound	Number (4)		
Area	Number (8,3)		Area for each altitude interval
Winter_Balance	Number (5)		Winter Balance
Summer_Balance	Number (5)		Summer Balance
Net_Acc	Number (5)		Net Accumulation
Net_Abl	Number (5)		Net Ablation
Net_Balance	Number (5)		Net Balance

Table A1.6 Territory

Code	Varchar2 (2)	Not Null/PK	Country Code
Name	Varchar2 (20)		Name of the Country
Report_Position	Number (2)		Position in the FoG report

A2 Appendix

(Attribute information of the database tables) The table and attribute list for the World Glacier Inventory database

Table A2.1 WGI_Edi

Ctr_Code	Varchar2 (2)	Not Null/PK	Country Code
Edi_Year	Number (4)		Year of the Edition
Edi_Name	Varchar2 (20)		Editor names
Edi_Adr	Varchar2 (30)		Editor addresses

Table A2.2 WGI_Gla

Political_Unit	Varchar2 (2)	Not Null/PK	Political Unit part of the WGI-number (Position 1 and 2)
River_Basin	Varchar2 (5)	Not Null/PK	Code for the determination of the hydrological catchment areas (Position 3 designates the continent, positions 4 to 7 are for the drainage code)
Free_Position	Varchar2 (2)		Freely chosen number (Positions 8 and 9)
Local_Code	Varchar2 (3)		Local Code helps to make the WGI-number clear (Positions 10 to 12)
Vaw_Add_Code	Number (3)	Not Null/PK	Additional Code to clear double WGI-numbers
Name	Varchar2 (16)		Name
Card_Point_Lat	Varchar2 (1)		Cardinal point (N or S)
Degrees_Lat	Number (2)		Latitude in degrees
Minutes_Lat	Number (4,2)		Latitude in minutes and seconds
Card_Point_Lon	Varchar2 (1)		Cardinal point (E or W)
Degrees_Lon	Number (3)		Longitude in degrees
Minutes_Lon	Number (4,2)		Longitude in minutes and seconds
Coordinates	Varchar2 (15)		Coordinates
Nr_States	Number (1)		Number of independent states
Official_Rem	Varchar2 (255)		Official remarks

Table A2.2 (suite)

Internal_Rem	Varchar2 (255)		Internal remarks
Fluctu_Key	Number (5)		Reference Key to the FoG

TABLE A2.3	WGI_Var		
Political_Unit	Varchar2 (2)	Not Null/PK	Political Unit part of the WGI-number (Position 1 and 2)
River_Basin	Varchar2 (5)	Not Null/PK	Code for the determination of the hydrological catchment areas (Position 3 designates the continent, positions 4 to 7 are for the drainage code)
Free_Position	Varchar2 (2)		Freely chosen number (Positions 8 and 9)
Local_Code	Varchar2 (3)		Local Code helps to make the WGI-number clear (Positions 10 to 12)
VAW_Add_Code	Number (3)	Not Null/PK	Additional Code to clear double WGI-numbers
Year	Number (4)	Not Null/PK	Year of the map or photo year
Nr_Drainage_B	Number (1)		Number of drainage basins
Map_Scale	Number (4)		Map scale
Photo_Type	Varchar2 (1)		Photo type
Photo_Year	Number (4)		Photo year
Total_Area	Number (8,3)		Total surface area
Area_Accuracy	Varchar2 (1)		Accuracy rating of the area
State_Area	Number (8,3)		Total area in the state concerned
Exposed_Area	Number (8,3)		Total exposed surface area
Ablation_Area	Number (8,3)		Ablation area
Mean_Width	Number (4,1)		Mean width
Mean_Length	Number (4,1)		Mean length
Max_Len_Tot	Number (4,1)		Maximum length total
Max_Len_Expo	Number (4,1)		Maximum length exposed
Max_Len_Abla	Number (4,1)		Maximum length ablation area
Expos_Acc_Area	Varchar2 (2)		Exosition accumulation area
Expos_Abl_Area	Varchar2 (2)		Exposition ablation area
Highest_Elev	Number (4)		Highest elevation
Median_Elev	Number (4)		Median elevation (mostly mean elevation)
Lowest_Elev_Tot	Number (4)		Lowest elevation total
Lowest_Elev_Exp	Number (4)		Lowest elevation exposed
Mean_Elev_Acc	Number (4)		Mean elevation accumulation area
Mean_Elev_Abl	Number (4)		Mean elevation ablation area
Prim_Classific	Varchar2 (1)		Primary classification
Form	Varchar2 (1)		Form
Frontal_Chars	Varchar2 (1)		Frontal characteristics
Longi_Prof	Varchar2 (1)		Longitudinal profile
Source_Nouri	Varchar2 (1)		Major source of nourishment
Acti_Tongue	Varchar2 (1)		Activity of the tongue
Acti_From	Number (4)		Beginning of the period for which the tongue activity was assessed
Acti_To	Number (4)		End of the period for which the tongue activity was assessed
Moracl1	Varchar2 (1)		Moraine classification
Moracl2	Varchar2 (2)		Moraine classification
Snow_Line_Elev	Number (4)		Snow-line elevation
Snow_Line_Accr	Varchar2 (1)		Snow-line elevation accuracy rating
Snow_Line_Date	Varchar2 (8)		Snow-line elevation date
Mean_Depth	Number (4)		Mean depth
Depth_Accuracy	Varchar2 (1)		Mean depth accuracy rating
Loading_Date	Date		Date of loading into database

A3 Appendix
(Example for retrieving data)

This appendix explains how data can be retrieved from the database system. The database is now public and can be reached via Internet. If a connection exists to Internet and one has access to ORACLE, a login to the database is possible via SQL*Net. The login to the database is done by means of the following statement:

sqlplus wgmsuser/wgmsuser@t:129.132.1.52:hil7:,,,I.

A 'wgmsuser' has all 'select' rights to the WGMS database and therefore to the tables described above in the form of SQL statements. The next three sections show some query examples:

WGI

This example shows a query on the World Glacier Inventory database. The following example selects the sum of total area of all glaciers, which are stored in the database at the moment for each country.

select political_unit, sum(total_area)
from wgms.wgi_var group by political_unit

Result:

Political_Unit	*Sum (Total_Area)*
A	542.23
CH	1,341.69
D	1.15
E	5.27
F	424.01
GL	1,070.20
I	754.29
N	35,865.51
NP	1,641.55
NZ	1,159.10
PK	301.26
S	313.07

FoG

The next example shows a more complex query on the FoG database. The query is simplified to enable the 'wgmsuser' to create different views. By combining these views, simpler queries are possible.

The first view to be created selects all information about the quantitative and qualitative length change of glaciers in Switzerland.

Query

create view variations as
select a.internal_key, a.name,b.year, b.variation_horiz, b.qualitative_var
from wgms.glacier a, wgms.glacier_state b
where a.political_unit like 'CH'
and a.internal_key = b.glacier
and (not b.variation_horiz is null or not b.qualitative_var is null)

The second view to be created selects all information on area- and volume-change of the glaciers in Switzerland.

Query

create view volume_changes as
select a.internal_key, a.name,b.year, b.area_change, volume_change
from wgms.glacier a, wgms.glacierpart_variation b
where a.political_unit like 'CH'
and a.internal_key = b.glacier
and b.lower_bound = 9999
and (not b.area_change is null or not b.volume_change is null)

After creating the two views, it is now possible to create a simple query which results in all glaciers of Switzerland having area- or volume-change data and qualitative or quantitative length change data.

Query

select a.name, a.year, a.variation_horiz, qualitative_var, b.volume_change, b.area_change
from variations a, volume_changes b
where a.internal_key = b.internal_key and a.year = b.year
order by a.name, a.year

Result:

Name	*Year*	*Variation_Horiz*	*QU*	*Volume_Change*	*Area_Change*
Gries (Aegina)	1979	–21.8		–11,251	–353
Gries (Aegina)	1986	–13.4		–2,350	– 88
Limmern	1977	– 0.7		–11,394	–194
Plattalva	1977		SN	–4,820	104

MBB

The last example shows a query on the FoG database too. The following query allows one to obtain mass balance-, equilibrium line altitude- and accumulation area ratio-data, which is published in the *Glacier Mass Balance Bulletin.*

Query

select a.name, b.year, b.equilibr_ln_alt, b.aar, c.net_balance from wgms.glacier a, wgms.mass_balance overview b, wgms.mass_balance c
where a.mbb like 'Y'
and a.name like 'STORGLACIAEREN'
and a.internal_key = b.glacier
and a.internal_key = c.glacier
and c.year = b.year
and c.lower_bound = 9999
order by a.name,b.year

(For the result, see the table below.)

Result:

Name	*Year*	*Equilibr_Ln_Alt*	*AAR*	*Net_Balance*
Storglaciaeren	1946	1,480	41	–1,130
Storglaciaeren	1947	1,600	18	–2,060
Storglaciaeren	1948	1,400	51	0
Storglaciaeren	1949	1,410	49	900
Storglaciaeren	1950	1,550	27	–1,290
Storglaciaeren	1951	1,500	37	–650
Storglaciaeren	1952	1,450	46	–160
Storglaciaeren	1953	1,530	29	–810
Storglaciaeren	1954	1,540	27	–970
Storglaciaeren	1955	1,470	42	–160
Storglaciaeren	1956	1,500	37	–480
Storglaciaeren	1957	1,480	40	–320
Storglaciaeren	1958	1,510	32	–650
Storglaciaeren	1959	1,540	27	–970
Storglaciaeren	1960	1,620	15	–1,610
Storglaciaeren	1961	1,575	21	–1,100
Storglaciaeren	1962	1,400	54	320
Storglaciaeren	1963	1,425	51	–190
Storglaciaeren	1964	1,400	54	490

Result:

Name	*Year*	*Equilibr_Ln_Alt*	*AAR*	*Net_Balance*
Storglaciaeren	1965	1,400	54	430
Storglaciaeren	1966	1,500	40	–530
Storglaciaeren	1967	1,500	40	–230
Storglaciaeren	1968	1,480	44	–100
Storglaciaeren	1969	1,570	24	–1,040
Storglaciaeren	1970	1,610	17	–1,520
Storglaciaeren	1971	1,490	41	–190
Storglaciaeren	1972	1,550	28	–1,050
Storglaciaeren	1973	1,490	41	50
Storglaciaeren	1974	1,480	43	–340
Storglaciaeren	1975	1,380	63	1,170
Storglaciaeren	1976	1,440	48	270
Storglaciaeren	1977	1,420	51	200
Storglaciaeren	1978	1,469	45	–80
Storglaciaeren	1979	1,497	39	–210
Storglaciaeren	1980	1,591	20	–1,270
Storglaciaeren	1981	1,500	39	–190
Storglaciaeren	1982	1,420	52	260
Storglaciaeren	1983	1,402	54	280
Storglaciaeren	1984	1,472	45	120
Storglaciaeren	1985	1,570	33	–720
Storglaciaeren	1986	1,430	52	–60
Storglaciaeren	1987	1,410	55	480
Storglaciaeren	1988	1,564	27	–840
Storglaciaeren	1989	1,374	66	1,240
Storglaciaeren	1990	1,495	60	590
Storglaciaeren	1991	1,460	47	170
Storglaciaeren	1992	1,393	58	880
Storglaciaeren	1993	1,397	58	1,000

A 'wgmsuser' has the ability to print the selected data on an ASCII-File. With the following commands at the beginning of a question, a file will be written to the host computer:

spool test.dat
set pagesize 2000
select ... from ... where ...
spool off

The newest information on the database will be available on the WGMS WWW-server in the near future.

4 Statistical analysis of glacier mass balance data

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4.1 INTRODUCTION

Mountain glaciers, like other parts of the cryosphere, atmosphere and hydrosphere, are affected by numerous regular and irregular processes and factors. That is why it is suitable to use statistical methods in the study of long-term temporal and spatial variabilities of glaciers. Two approaches to investigating glacier fluctuations, taking into account their irregular nature, must be discussed.

Statistical studies of spatio-temporal changes

In investigating glacier fluctuations, an attempt is made to use statistical models to describe spatial patterns unifying several glaciers within individual mountain areas, or several mountain areas within the whole hemisphere. The statistics of parameters of these models are then calculated and compared with each other. Investigations of this kind have produced important results with respect to relationships between dynamics of different glaciers, enabling conclusions to be drawn about the applicability of different glacier models, about methods of computing past balances and about the nature of the climate/glacier system. Most of the existing works dealing with mountain glacier statistics use similar approaches.

Stochastic studies

The aim of these investigations is to establish relationships between current and past states of glacier parameters using the theory of stochastic processes. Comparison of these models with stochastic models of various processes within the climatic system can help to understand fundamental features of the climate system and its atmosphere/glacier subsystem. Thus, the study of glacier dynamics moves in the opposite direction to the first approach: from random features to deterministic features.

It must be stressed that problems of glacier variabilities and climatic change can be a subject for discussion because of (a) the complexity of the climate system, (b) the incompleteness of observational data and (c) contradictions between different climate theories.

1 This work was performed in part while S.G. Dobrovolski held a National Research Council-NASA/GSFC Research Associateship.

4.2 STATISTICAL ANALYSIS OF GLACIER MASS BALANCE DATA

4.2.1 Mass balance distribution on a single glacier

Early available field data (Meier and Tangborn, 1965; Hoinkes, 1970), as well as more recent work (Funk, 1985), indicate that mass balance distribution over the glacier may be very complicated, depending mainly on the altitude but also on the distance from the edges (avalanches), on the surface albedo (morainic cover, crevasses), etc. Under such conditions, a simple single curve to represent the mass balance variation depending only on altitude should be considered as a rough approximation. Thus, in order to measure the actual mass balance of a glacier, a lot of surveys must be performed. However, glaciologists tend to utilize the specific annual mass balance of the glacier, which is the total mass balance equally distributed over the whole surface. In fact, for alpine glaciers, it seems that the yearly mass balance changes may affect the whole glacier more or less in the same way (Fig. 4.1 and 4.2) while, for other regions, the mass budget imbalances may be a criterion for a climatic classification of glaciers (Kuhn, 1984). The first effect is the physical basis for the application of the linear model of mass balance variations as proposed by Lliboutry (1974). In this statistical model, the specific mass balance b_{jt} as measured during the year t and at the site j over the time period T may be written as:

$$b_{jt} = \alpha_j + \beta_t + \varepsilon_{jt} \qquad (4\text{-}1)$$

Where α_j is the mean local specific mass balance,

$$\alpha_j = \frac{1}{T}\sum_{t=1}^{T} b_{jt} \qquad (4\text{-}2)$$

characteristic of the site j, β_t is yearly variation of specific mass balance, with $\sum_{t=0}^{T} \beta_t = 0$.

ε_{jt} is a centered random residual, the standard deviation of which gives the difference between the model and observations. Using this model, the β_t terms may be obtained with only a limited array of stakes, in the best selected place in the center of the glacier, avoiding all the troubles caused by surface irregularities, morainic cover, avalanches, etc. In this way, the β_t terms may indeed better represent the local climatic variations occurring over a limited range of altitude than does the mass balance integrated over the entire glacier.

4.2.2 Variations in equilibrium line altitudes

If this linear model of mass balance distribution works, the yearly variations of the equilibrium line altitude are linearly linked with the variations of the specific mass balance β_t (Fig. 4.3). Then $\beta_t = \langle db/dz \rangle_{EL} \cdot dz$ where $\langle db/dz \rangle_{EL}$ is the mean altitudinal variation of the mass balance near the equilibrium line and dz is the departure of the equilib-

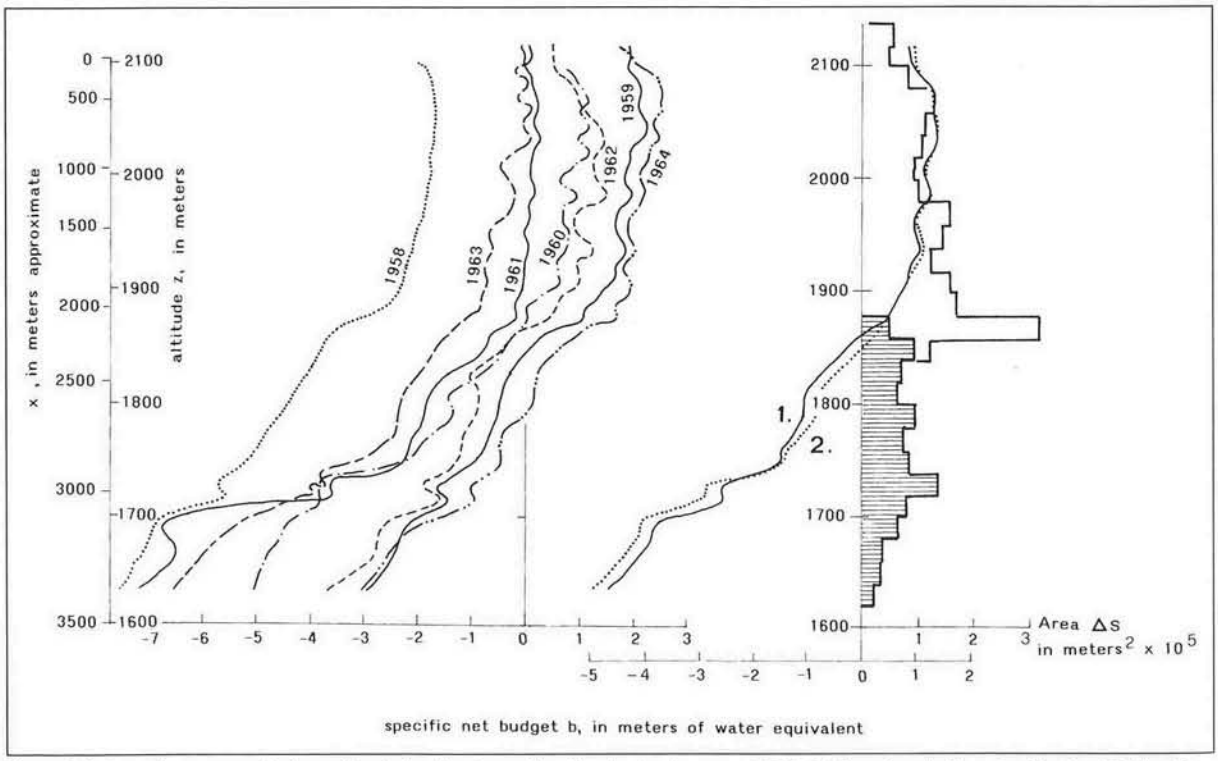

Figure 4.1 Specific net mass balance b(t,z) distribution with altitude for the years 1958–1964 on South Cascade Glacier, U.S.A. (after Meier and Tangborn, 1965).

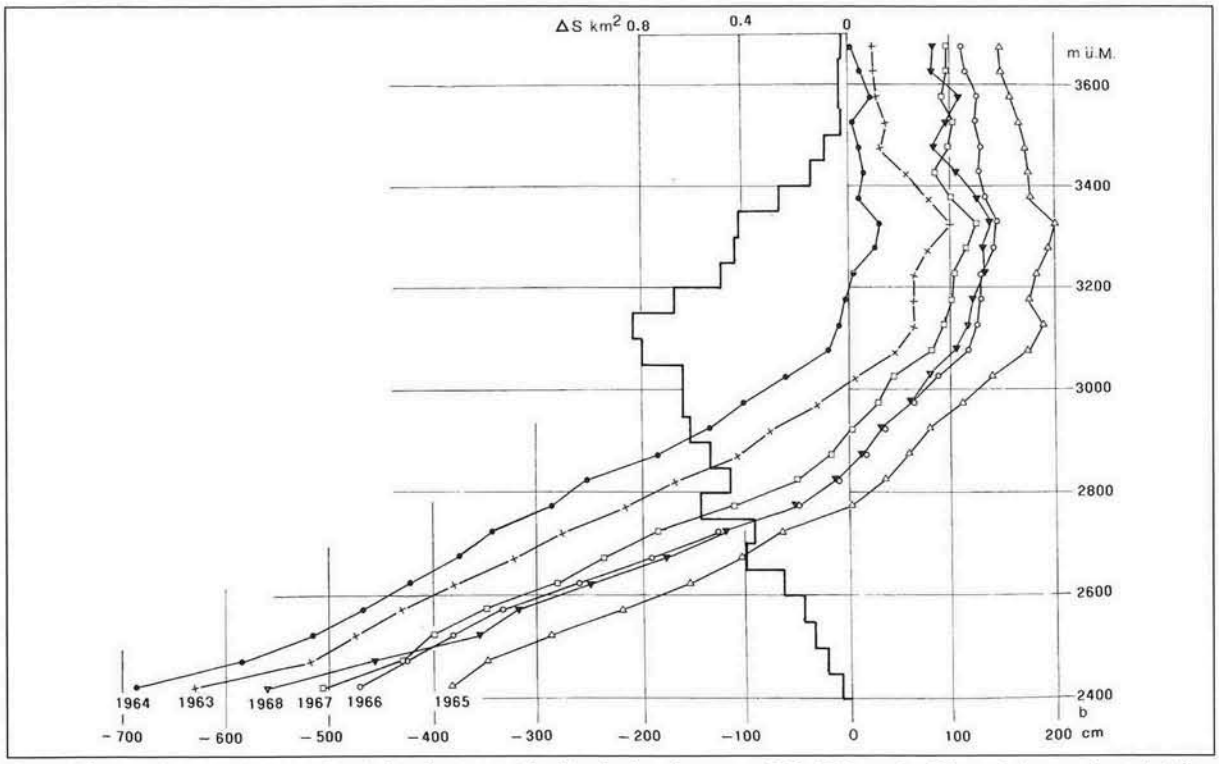

Figure 4.2 Specific mass balance b(t,z) distribution with altitude for the years 1963–1968 on the Hintereisferner, Austria (after Hoinkes, 1970).

rium line from its mean altitude for the considered span of time. This simple relation allows us to derive mass balance variations near the equilibrium line for large areas where <db/dz> is known and, if aerial photographs or satellite images of glaciers are available, at the right date of the year (Dedieu and Reynaud, 1992; Reynaud, 1993).

4.2.3 Mass balance reconstruction

An attractive application of the linear mass balance model is the statistical equation describing relationships with temperature and precipitation changes, such as

$$\beta t = a\ dT_t + b\ dP_t + e_t \qquad (4\text{-}3)$$

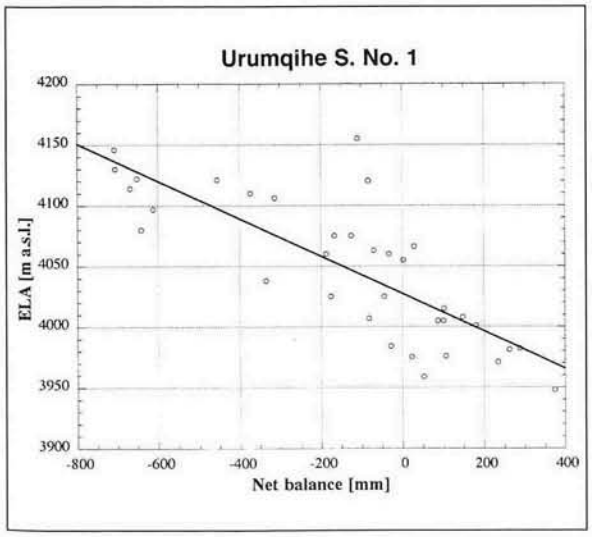

Figure 4.3 Equilibrium line altitude as a function of the specific annual mass balance for the Urumqihe S. No. 1, China.

where dT_t and dP_t are the departure from the mean of the selected time span for temperature and precipitation, respectively, and e_t is a random centered residue. As T_t and P_t are generally not independent variables, the best multivariate correlation to derive a and b works with data sets of P and T, which relate to different seasons within individual years. The significance of this statistical approach is not only found in the description of the relationships between the mass balance and climatic parameters but also in the reconstruction of a mass balance variations series using the P and T data from a neighbouring meteorological station for a period of time extending back beyond the time of field measurements on the glacier, in order to have a better knowledge of the climatic signal affecting the glacier. Around the 1980s, when direct glaciological mass balance measurements were only 20 to 30 years old, several attempts were made to reconstruct mass balance series for a century or even a bit more (Martin, 1978; Tangborn, 1980; Diurgerov and Popovnin, 1980; Tvede, 1982). In order to obtain consistent results over such long time periods, however, a linear relationship derived from only 20 or 30 (or even fewer) years of measured data must be assumed to have remained constant during the whole reconstruction period. In reality, such relationships may change a lot, especially with changing climatic conditions. Vallon *et al.* (1995), going back to the work of Martin (1978) which had been based on the 1949–1975 time interval, examined the whole 1949–1991 set of data in search of the variance percentage explained for shorter spans of time. With the 100 km distant meteorological station of Lyon-Bron, they again found the same general results that Martin had come up with for the same

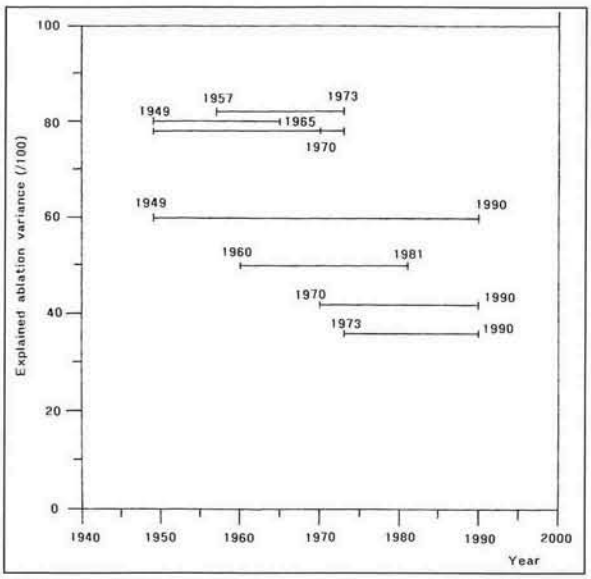

Figure 4.4 Variance percentage explanation of the Sarennes Glacier ablation by the sum of summer temperatures exceeding 17.5°C at Lyon Bron for different time periods (after Vallon *et al.*, 1995).

period. However, it appeared that the fairly good score previously explaining 74% of the variance had fallen to less than 40% during the last two decades (Fig. 4.4). Among the different explanations suggested as a cause of this lower performance are:

a) the less reliable representation of the insolation at high altitudes by air temperature data collected at low-altitude meteorological stations, and
b) the heavy change (reduction) in glacier surface albedo after several years of negative mass balance actually surveyed and underestimated by the model computations.

The latter type of changes is likely to have occurred in the past and may have introduced a major inaccuracy in the reconstructions, despite the obvious strength of the climate/balance relationships. As past evidence of marked albedo changes, some old photographs show alpine glaciers at the beginning of the 20th century with dark surfaces and high equilibrium lines (higher than 3,200 m for the Sarennes Glacier in 1906, when the glacier was, in fact, totally stripped of its firn).

4.2.4 Spatio-temporal mass balance distribution

The possibility of mass balance reconstructions using data from distant meteorological stations suggests that the mass balance fluctuations over an individual mountainous range might be quite similar, as also shown by the cumulative mass balance curves for different glaciers of the same area (IAHS(ICSI)/UNEP/UNESCO, 1994). A first attempt at mass balance intercomparison was made by extending Lliboutry's linear model. In this case, each glacier plays the role of a stake and b_{jt} (the specific mass balance of the year t for the glacier j) may be expressed as:

$$b_{j,t} = \alpha_j + \beta_{jt} \qquad (4\text{-}4)$$

where α_j is the mean specific mass balance over the T period and β_{jt} the deviation from the mean for each glacier. Application of this model to the data set given by 4 glaciers and 27 years, the largest common period of direct glaciological mass balance measurements, shows that the different terms β_{jt} are not ran-

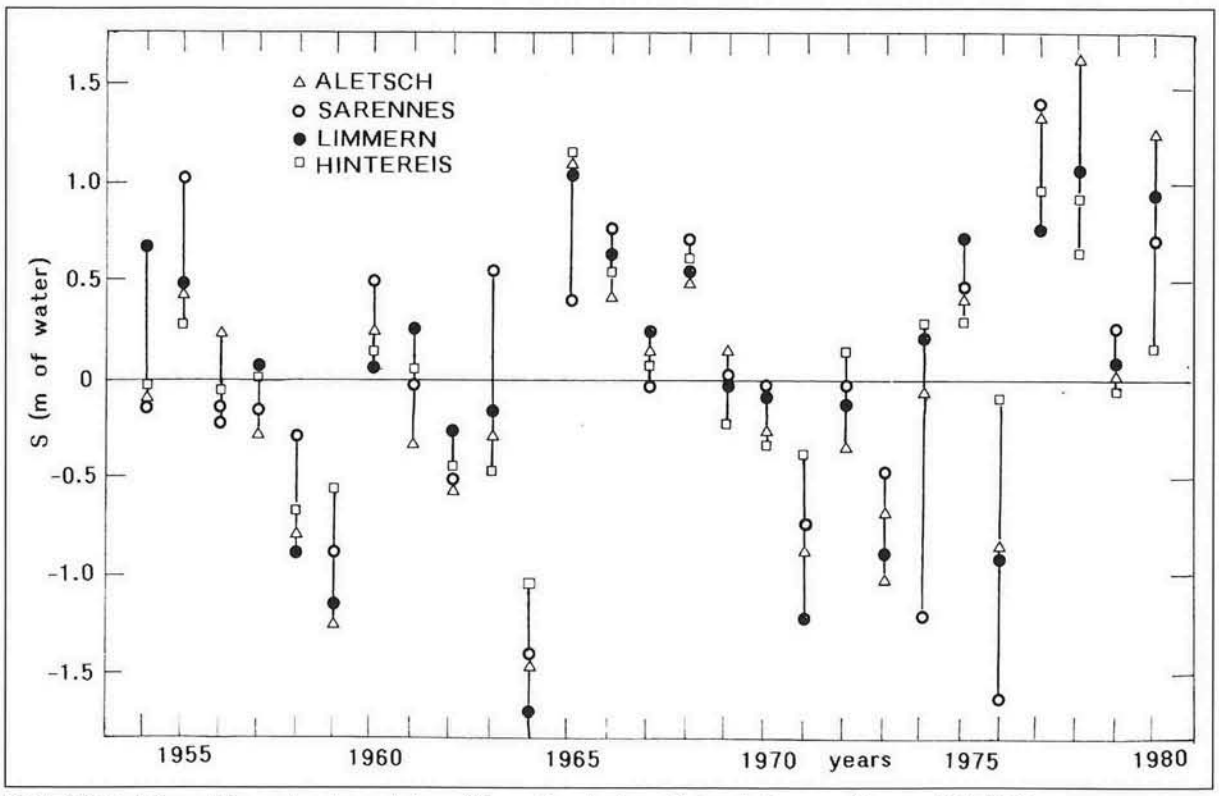

Figure 4.5 Variations of the centered mass balance βtj over time for four Alpine glaciers over 27 years: 1954–1980.

domly distributed, exhibiting a strong homogeneity of variation over time (Fig. 4.5). The factor analysis allows the score of this variation homogeneity, which reaches 85% of the total variance, to be globally quantified. It appears that the second factor accounts for 9%, this being the distribution of glaciers along the alpine range where the glaciers are sorted from France to Austria, which is clearly a geographical dependence. On the remaining 6%, no recognizable structure was found (Fig. 4.6). The same factor analysis applied to three other mountainous areas (Scandinavia, Tien Shan, North America) exhibits the same type of spatio-temporal distribution of mass balance, with comparable scores around 80% for the homogeneity and 10% for the geographical factor. Even in South America, where three Peruvian glaciers were surveyed, this structure could be found (Ames, 1985). It is noteworthy that the direct representation of b_{jt} fluctuations over time, such as those plotted on Fig. 4.5, is slightly different from the factor analysis that utilizes the standardized variation: β_{jt}/σ_j. As a consequence, the factor analysis results for the global homogeneity may be higher than any 2 by 2 correlation score. The directly extended linear model works well only when the standard deviations involved are similar. In the case of different σ_j like in Scandinavia, only β_{jt}/σ_j is comparable.

It must also be noted that the homogeneity of mass balance fluctuations fails in individual years when only part of the mountainous area experiences certain weather regimes, like in 1976 when summer precipitation did not occur in the southern Alps. Nevertheless, the yearly mass balance structure vanishes with the borders of each mountainous area, as already shown by various authors using the 20 to 30 year-long data set of field measurements. With the longest reconstructed mass balance series using precipitation and temperature data for the common 1896–1979 period, a comparison is possible for the following five glaciers:

Folgefonni Glacier	Norway	(Tvede, 1982),
Djankuat Glacier	Caucasus	(Diurgerov and Popovnin, 1980),
Igan Glacier	Polar Urals	(Khodakov, 1966),
Sarennes Glacier	French Alps	(Martin, 1978),
South Cascades	US Cascades	(Tangborn, 1980).

The results of this comparison, plotted on Fig. 4.7, show that four of these five reconstructions give a similar trend of evolution for the last century, while the South Cascade set of data differs notably. Vallon *et al.* (1986) examined the statistical validity of this comparison using the uncertainty of each series as a superimposed random noise. It was found that the four European glaciers were in agreement. Furthermore, Vallon *et al.* (1986) suggested a method to take into account the fact that some glaciers, such as South Cascade (USA) or Mer de Glace (France), must have been very far from their equilibrium position at the beginning of the comparison period but gradually returned to it. Once corrected for this influence, the American glacier behaved like the Norwegian glacier. More generally, it seems that, over the last century, only the most marked climatic changes have influenced two distant areas in the same way. Small global changes are not visible, disappearing within local variations or being hidden by the large natural variability. As a check of mass balance reconstructions, it can be noted that they are in good agreement with the simplest glacier fluctuation evidence: the cumulative length variations. In the European Alps, the causes of the intermittent glacier readvances in the 1890s, 1920s and 1970s can be found (cf. also the quantitative reconstruction of secular mass balances from cumulative length changes as reported by Haeberli, 1996).

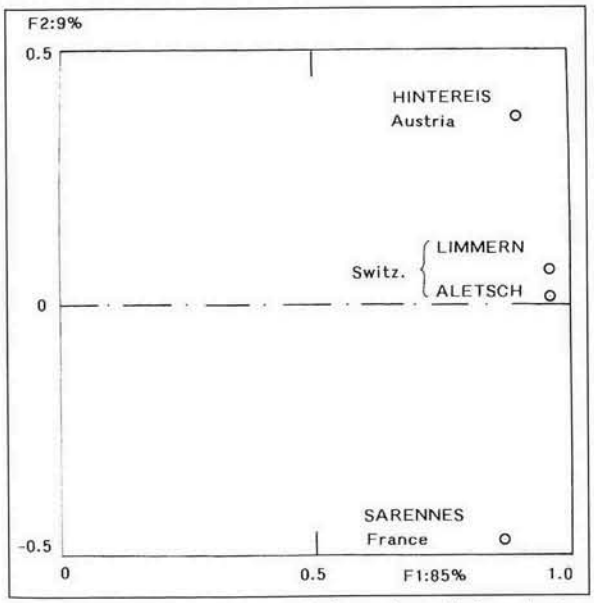

Figure 4.6 Factor analysis applied to four Alpine glaciers: 1954–1980. First and second axis graph (after Reynaud *et al.*, 1984).

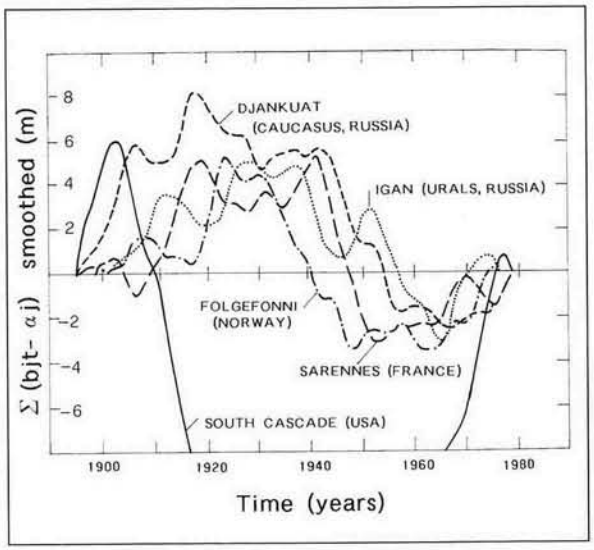

Figure 4.7 Accumulated deviation from the mean for five reconstructed mass balance series. For the clarity of the figure, the five curves have been smoothed using a five-point triangular filter (after Letréguilly and Reynaud, 1990).

4.2.5 Test of reconstruction models

Thanks to the regular measurements taken on a few glaciers since the beginning of the 20th century to follow the variations of a limited sector within the glacier's ablation area between 2 cross profiles, where the mean velocity and the mean altitude were surveyed each year, it has been possible, using modern geophysical tools, to determine the shape of the glacier valley and thereby derive the yearly specific mass balance of the monitored sectors. For this purpose, one can use the continuity equation written in finite differences in between two cross profiles, together with some reasonable assumptions about the internal flow conditions. For instance, the 43-year long series of a sector of Gebroulaz Glacier (France) is given in Fig. 4.8 for the 1907–1950 period (Reynaud *et al.*, 1986). It appears that the results given by the Sarennes Glacier reconstruction and those of Gebroulaz are in rather good agreement, even if they experienced very heavy changes in mass balance, such as in the 1910s or the 1940s. It should be borne in mind that the cumulative mass balance variation over time ($\sum_0^t \beta_{jt}$) allows us to enlarge the usual $\sum_0^t b_{jt}$ curves in order to facilitate the visual comparison of the relative variations. With this graphical plotting, the area between two individual curves is directly proportional to the correlation coefficient.

4.2.6 Conclusions

As for all statistical models, those established for the glacier mass balance distribution analysis on different spatial scales or those adopted to reconstruct the past mass balance history have to be checked against the field truth. Presently, owing to both the increasing length of field time series and the new mass balance measurements, some severe limitations are appearing for the usual simple statistical models. The future statistical mass balance analysis will probably refine our knowledge of the way in which the glacier mass balance tells us the spatio-temporal distribution of climatic changes. This knowledge may also serve as a test for various climatic models. At the same time, for this kind of research, field data play a decisive role, providing its fundamental base. Unfortunately, these field data and their analysis are quite recent, so much basic information can likely still be derived from them. As a consequence, particular attention must be focused on the ongoing measurements, as well as on those yet to be established

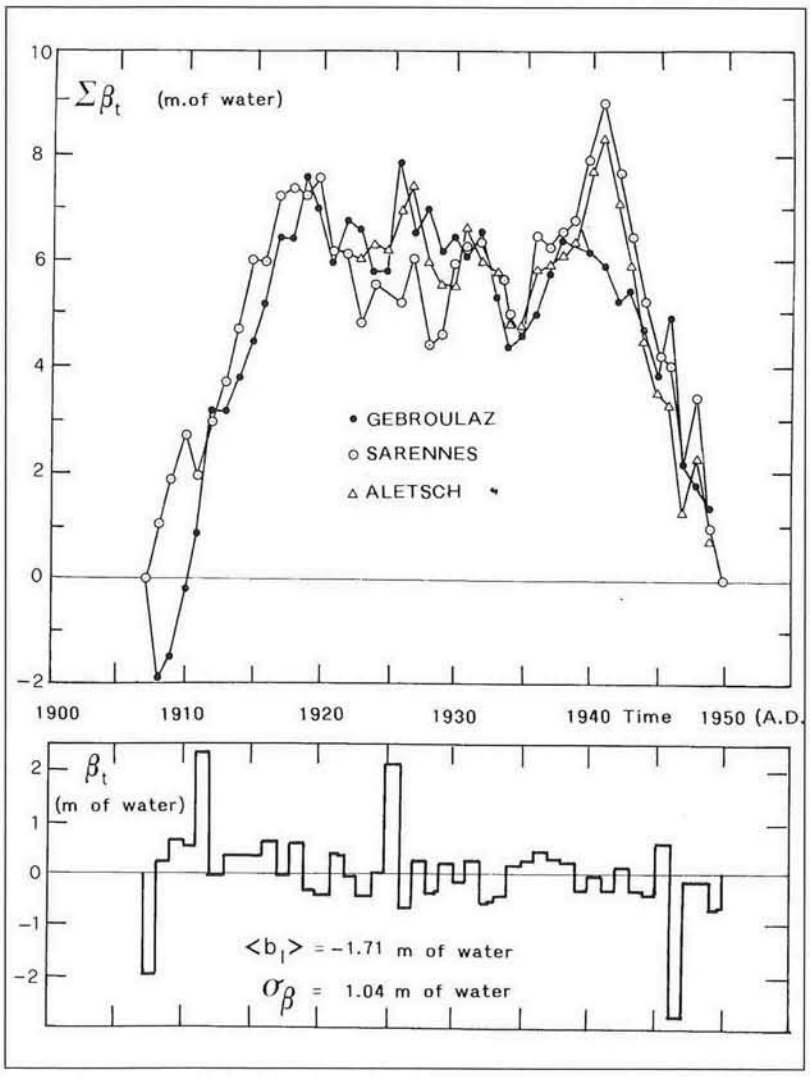

Figure 4.8 βt balance series from the continuity equation applied to the Gébroulaz Glacier (bottom) and a comparison of its cumulative variations with two long series available in the Alps: Sarennes Glacier and Aletschgletscher (top) (after Reynaud *et al.*, 1986).

in other parts of the world where the amount of glacial fluctuation data already available is minimal or even non-existent.

4.3 STOCHASTIC ANALYSIS OF GLACIER VARIABILITIES

Although statistical and stochastic methods are widely available nowadays, the results of their application by different researchers are often contradictory, even if relating to the same time series. Thus, changes in global temperature and mean sea level are still the object of heated discussion. The majority of researchers consider them the manifestation of man-made warming, whereas others suggest that there is still no evidence of global heating and its consequences (Elsaesser *et al.*, 1986) or describe global heating as a random walk (Gordon, 1991). From this point of view, it is expedient to implement detailed stochastic analysis of a data set as interesting as mass balances of mountain glaciers and to compare results with those obtained for other components of the climate system.

4.3.1 Data and methods of analysis

Time series of annual cumulative specific net balances of 26 mountain glaciers were analysed (Table 4.1). Data were taken from the *Glacier Mass Balance Bulletin* No. 2 issued by the World Glacier Monitoring Service (IAHS(ICSI)/UNEP/UNESCO, 1993). The shortest series (with a length of less than 20 years) were not taken into consideration. Thus, it was the balance variabilities documented for most of the glaciers, with long periods of appropriate observations (as described by Hoelzle and Trindler, 1996), that were studied. In the study, a new modification of the maximum entropy method (MEM) of spectral and stochastic analysis was used. Conventional MEM analysis in its variant described by Dobrovolski (1992) can be summarized as follows.

Each time series x(t) was considered a segment of one of the realizations of an autoregressive (AR) process X(t) of order m:

$$x(t) = k_1 X(t-1) + \mathrm{K} + k_m X(t-m) + a(t) \qquad (4\text{-}5)$$

Here, t = time in years; k = coefficient of AR (m) process; a(t) = sequence of uniformly distributed, non-correlated values (residual white noise). In equation (4-5), for the purpose of simplicity, the mean value (mathematical expectation) of the process X(t) is considered to be zero. Coefficients k were calculated using a modified Burg-Levinson's algorithm. Five criteria were used for choosing the order of the model, m: Akaike's, Akaike's Information Criterion, Parzen's, Schwarz-Rissanen's and Hannan-Quinn's (cf. Privalsky and Jensen, 1993). Confidence limits for estimations of spectral densities were calculated with the help of the asymptotic Cromer formula. Nevertheless, the characteristic equation of autoregression was solved for each m in order to check the stationarity of the model. The main modification to the applied method relates to the calculation of normalized spectra. A procedure of low frequency truncation (cf. Dobrovolski and Choudhury, 1996) was proposed in order to minimize confidence limits of normalized spectra estimations. As will be shown in the following paragraph, this procedure considerably reduces the mentioned confidence limits in the case of glacier balance variations.

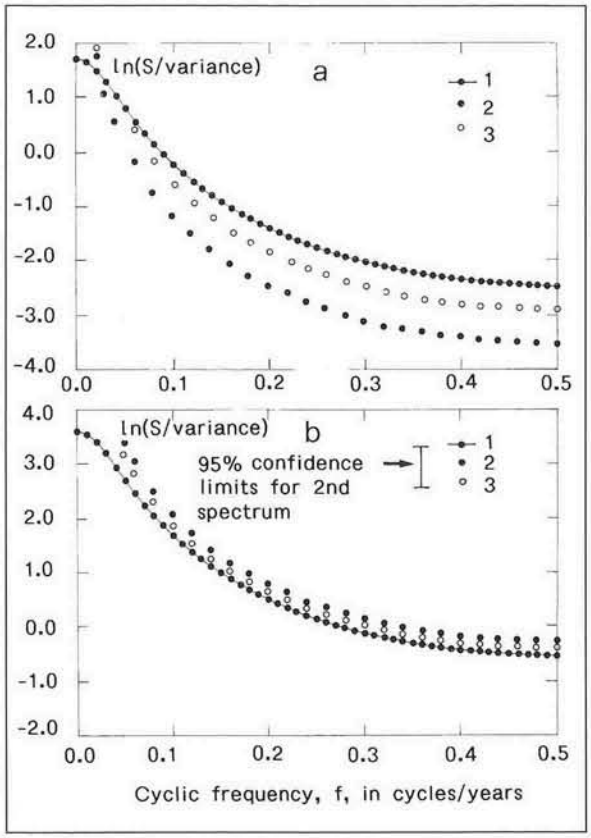

Figure 4.9 Traditional (a) and frequency–truncated (b) estimations of normalized spectra of glacier net mass balance series. 1: Djankuat Glacier (Russia), 2: Kesselwandferner (Austria). Curve 3 corresponds to the spectrum of the Wiener process realization.

4.3.2 Results of stochastic analysis

In Fig. 4.9, typical spectra of mountain glacier's specific balances (Djankuat Glacier in Russia and Kesselwandferner in Austria), together with spectra of a realization of the discrete-time Wiener process, are shown. It should be borne in mind that the discrete-time Wiener process is a nonstationary sequence of random numbers with stationary temporal increments, i.e., if m and k in Eq. (4-5) are both equal to 1, then

$$X(t) = X(t-1) + a(t) \qquad (4\text{-}6)$$

Fig. 4.9*a* presents traditional estimations of normalized spectra, whereas Fig. 4.9*b* shows frequency-truncated spectra. It can be seen that the difference

TABLE 4.1 Results of stochastic analysis of changes in mountain glaciers' net specific balances

Glacier	*Country*	*Mean incr.* $M\,\Delta x$ *(m/year)*	*St. dev.* $\sigma\,\Delta x$ *(m/year)*	*k'(1)*	*Order* *m*	*Result*
Abramov	Kirghizstan	–0.5	0.6	0.2	0	Trend –
Ålfotbreen	Norway	0.3	1.3	–0.1	0	Wiener
Au.Brøggerbreen	Svalbard (Nor.)	–0.4	0.3	0.0	0	Trend –
Caresèr	Italy	–0.6	0.7	0.4	0–1	Trend –
Djankuat	Russia	–0.1	0.6	–0.3	0	Wiener
Engabreen	Norway	0.7	1.2	–0.2	0	Trend +
Gråsubreen	Norway	–0.1	0.7	–0.2	0	Wiener
Gries	Switzerland	–0.4	0.9	0.2	0	Trend –
Gulkana	Alaska	–0.2	0.4	0.0	0	Trend –
Hintereisferner	Austria	–0.4	0.5	0.2	0	Trend –
Kara Batkak	Kirghizstan	–0.4	0.5	0.3	0–1	Trend –
Kesselwandferner	Austria	0.0	0.4	0.3	0	Wiener
Kozelskiy	Russia	–0.2	1.1	–0.1	0	Wiener
Mid.Lovénbreen	Svalbard (Nor.)	–0.3	0.3	–0.1	0	Trend –
Maliy Aktru	Russia	0.0	0.5	0.1	0	Wiener
Nigarsbreen	Norway	0.4	1.1	–0.1	0	Trend +
Place	Canada	–0.7	0.7	0.1	0	Trend –
Sarennes	France	–0.6	0.9	0.2	0	Trend –
South Cascade	USA	–0.5	0.9	0.0	0	Trend –
Silvretta	Switzerland	0.1	0.8	0.1	0	Wiener
Sonnblickkees	Austria	–0.1	0.7	0.2	0	Wiener
Storbreen	Norway	–0.2	0.6	0.1	0	Trend –
Storglaciären	Sweden	–0.3	0.5	0.2	0	Trend –
Saint Sorlin	France	–0.5	0.9	0.1	0	Trend –
Ts.Tuyuksuyskiy	Kazakhstan	–0.4	0.5	0.4	1	Trend –
Wolverine	Alaska	–0.1	1.2	0.1	0	Wiener

between spectra in the first case is large but, if more accurate estimations are used, spectra of glacier balances practically coincide with each other and with the spectrum of the Wiener process. This coincidence is nevertheless not enough for safe conclusions to be drawn about the Wiener character of glacier changes. Year-to-year increments of specific balances must also be analysed. If glacier variability is described by Eq. (4-6), these increments must be white noises a(t). Thus, investigation of stochastic properties of glacier dynamics is finally reduced to the analysis of year-to-year increments of net balances.

In Table 4.1, results of these calculations are shown. Here, $M\Delta x$ = the mean value of year-to-year increments of glacier specific net balances; $\sigma\Delta x$ = standard deviation of increments; k' = coefficient of the model (4-5) for the description of the series of increments; m = order of the model (4-5) for increments determined using the five criteria mentioned in paragraph 4.3.1 and taking into consideration confidence limits of spectra estimations. The result is the type of model for the description of initial time series of net balances (cf. below).

For almost all time series of year-to-year increments of net balance, save Tsentral'nyi Tuyuksuiskii and perhaps Kareser and Kara-Batkak, the zero-order model (noncorrelated in time process) was a good approximation. Yet the essential problem relates to the question of whether the series of increments contain a statistically significant, non-zero mean value. In other words: is there a linear trend within initial time series of net balances? In order to answer this question, the method of assessing the errors in the estimations of the mean value of the process described by Yaglom (1987) was applied to the above-mentioned three glaciers and very simple considerations were used for other series of increments described by non-correlated random time sequences. For the last group of glaciers, the mean error in the estimation of the mean value of balance increments, $\sigma(M\Delta x)$ can be expressed in terms of the standard deviation of increments, $\sigma(\Delta x)$, and the length of the series, N:

$$\sigma(M\Delta x) = \sigma(\Delta x)/N^{1/2} \qquad (4\text{-}7)$$

If the absolute value of mean increments is less than the error of estimation of the mathematical expectation of white noise with appropriate standard deviation (noncorrelated process with zero mean), the monotonous (linear) trend in the series of balances is not statistically significant and changes in balances are described by the Wiener process model. By contrast, if the absolute mean increment is much larger

than the mean error of estimation of the white-noise mean value, the trend (positive or negative) is significant and glacier balances are described by the following model:

$$X(t) = X(t-1) + a(t) + ct \qquad (4\text{-}8)$$

where c = constant (positive or negative). For the three glaciers with significant values of k', the model is:

$$X(t) = (k'+1) * X(t-1) - k'X(t-2) + a(t) + ct \qquad (4\text{-}9)$$

Thus, from 26 initial series of glacier mass balances, 15 are well described by the models (4-8) and (4-9) with a negative trend (denoted by 'Trend –' in Table 4.1), 9 by the Wiener process model (4-6) ('Wiener' in Table 4.1), and 2 by model (4-8) with a positive trend ('Trend +' in Table 4.1). It can be seen that this pattern of temporal changes in glacier mass balances differs from that suggested by simple consideration of virtually positive (4 glaciers) or negative (22 glaciers) balances during the period of observations. From the point of view of the calculations presented here, predominance of glaciers with negative trends is not so dramatic (though still evident).

The model of the Wiener process for the description of glacier variability, as well as the term 'trend', deserve some comment. One often uses the term 'trend' when there is some visible tendency towards increasing or lowering within the time series. From a mathematical point of view, it is more suitable to use the term 'trend' when dealing with a statistically significant monotonous, deterministic change in mathematical expectation (mean value) of the process under consideration. The Wiener process (4-6) possesses some paradoxical properties. Although its temporal increments are absolutely stationary and have a zero mean value, the process itself is not stationary (as for variances) and the probability of extremely high or extremely low values towards the end of the time series is considerable. At the same time, it is stationary in terms of its mean value (mathematical expectation) and does not contain any trend from a mathematical point of view. There is an interesting geographical feature which can be deduced from Table 4.1. Glaciers with the largest natural variability of year-to-year changes in balances are those with a pronounced Wiener or positive type of glacier variability (Alfotbreen, Engabreen and Nigardsbreen in Norway, Kozelskiy in Kamchatka, Wolverine in Alaska). For all these glaciers, $\sigma\Delta x$ is more than 1 m/yr. These glaciers are situated near major atmospheric and oceanic frontal zones with high interannual variabilities of atmospheric processes.

4.3.3 Discussion

It is important to compare the results of stochastic analysis of glacier variabilities with the results of stochastic modelling of other long-term processes on the surface of the Earth, in the atmosphere and in the ocean. In Ratkovich (1976), Privalsky (1985) and Dobrovolski (1991, 1992), several thousand time series of hydrological, climatological and oceanic parameters were studied stochastically and only a few processes containing significant monotonous trends were found (concentrations of several gases in the atmosphere, sea level changes in regions with movements of the earth crust and some others). Time series of annual values of atmospheric parameters affecting mountain glaciers - temperature, precipitation, humidity and wind speed - on local and regional scales, are usually well-described by stationary first-order Markov processes without (at least without well-pronounced) trends. There is, thus, a certain contradiction between the quasi-stationary character of atmospheric variability and the well-pronounced nonstationarity of mountain glacier's variability. The difference between atmospheric and glacier characteristic time scales could be the cause of this phenomenon. Further stochastic studies of glacier oscillations and experiments on stochastic forcing of deterministic glacier models (described in chapter 5, Modelling Glacier Fluctuations, of the present volume) are needed. Conception of two-scale weather-climate separation by Hasselmann (1976) can be used in order to create the atmospheric forcing of deterministic glacier models. On the whole, statistic and stochastic properties of glacier variations are far from being sufficiently well understood. The 'practical conclusion' by Forel (1895) is still true: 'Préparons-nous à de la patience, de la persévérance, de la prudence dans nos conclusions.'

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5 Modelling glacier fluctuations

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5.1 INTRODUCTION

Understanding glacier fluctuations is of great interest for at least three reasons. Firstly, individual glaciers may affect human activities (e.g., threat of ice avalanches, runoff for hydropower basins and irrigation, blocking of roads, ski areas). Secondly, world sea level is directly linked to ice volume. Thirdly, glaciers are very sensitive to changing meteorological conditions, so glacier records form an important source of proxy data on climatic change.

The level of sophistication required in modelling depends very much on the purpose. Forecasting of ice avalanches, in particular, requires detailed finite element calculations of the entire stress state in a glacier, with very accurate geometric input. Trying to interpret a four-century long record of snout positions, on the other hand, is basically a matter of mass continuity. Here, the difficulty lies in relating glacier mass balance to meteorological quantities, not in the treatment of ice mechanics.

In this contribution, the interpretation of long records (> 10 a) is the central theme. This means that only so-called 'continuity models' will be discussed. Ample attention is given to the formulation of the mass balance. Only valley glaciers will be discussed.

5.2 TIME SCALES AND EQUILIBRIA

Before describing models that can be used to analyse glacier fluctuations, a summary of some concepts and definitions is in order. As meteorological conditions change all the time and ice movement is slow, glaciers are never in full equilibrium with the forcing. The notion of time scales thus plays an important role. There has been some confusion in the past as to the definition of response time of a glacier. I propose the following definitions. Let V_i be the equilibrium volume of a glacier, with equilibrium length L_i, that is, in perfect balance with the prevailing (constant) climatic state C_i. Various time scales are then defined as follows:

Growth time (τ_g): The growth time is the time a glacier in a constant climatic state C_i needs to attain a volume of $(1-1/e)V_i$, starting from zero ice volume.

Relaxation time (τ_{rel}): Consider a glacier in a constant climatic state where a small part of the mass dV is removed. The relaxation time is the time needed to return to equilibrium, i.e., to attain a volume $V_i - dV/e$.

Response time (τ_{rV}): The climatic state is changed stepwise from C_1 to C_2. The corresponding changes in equilibrium glacier volume are V_1 and V_2. The response time now is the time a glacier needs to attain a volume $V_2 - (V_2 - V_1)/e$.

Length response time (τ_{rL}): As above for glacier length L_i.

It is obvious that the various time scales may differ significantly. The growth time is particularly large when bed slopes are very small because of height-

mass balance feedback (Oerlemans, 1981, 1989*a*). At the same time, the relaxation time can be small. For most glaciers, one expects relaxation and response time to have the same order of magnitude. Glacier length and volume can have different response times, depending on how mass balance conditions change. For a retreating glacier (flat snout), τ_{rL} may be larger than τ_{rV}. For an advancing glacier (steep snout), the opposite may be true. Note that, by definition, $\tau_{rV} = \tau_g$ when $V_1 = 0$.

The response times are of most practical interest. Can τ_{rV} and τ_{rL} be estimated for a particular glacier without detailed numerical modelling? Early attempts to infer values of τ_{rV} from theoretical considerations were based on linear kinematic wave theory (Nye, 1960, 1965*a*; Lliboutry, 1971; Hutter, 1983). Without going into the details of the theory here, the most important result was that τ_{rV} is to the order of L/u, where L is glacier length and u ice velocity at the snout. This yields characteristic response times of 100 to 1000 a for valley glaciers. Apart from the problem that u is poorly defined, many modellers have questioned the validity of this approach because there is substantial field evidence pointing to values of response time between 10 and 50 a. Numerical models, based on the continuity equation but lacking in sophisticated treatment of the glacier snout, yielded values of τ_{rV} or τ_{rL} more in line with the observations (e.g., Budd and Jenssen, 1975; Kruss, 1983; Oerlemans, 1986). Jóhannesson *et al.* (1989*a*, 1989*b*) have provided an elegant analysis of the problem. They render the statement plausible that a good estimate of the response time (which they referred to as volume time scale) is obtained from $\tau_{rV} = H/-b_{term}$. Here, H is a characteristic ice thickness and b_{term} the ablation rate at the terminus. They also show that, for regular valley glaciers, the details of the snout dynamics are not important for the global behaviour of the glacier and provide theoretical support for what many glaciologists have assumed intuitively.

The relation between climate and the equilibrium state of a glacier is not necessarily single-valued. Consider, for instance $L_{eq}(E)$, the equilibrium glacier length in dependence on the equilibrium-line altitude E. A range of values of E may exist for which two equilibrium states are possible, depending on the history. For an ice sheet on a flat bed with a sloping equilibrium line, this has been nicely illustrated by Weertman (1961). For valley glaciers, it has been demonstrated by computer experiments (Oerlemans, 1989*a*). In fact, the number of equilibrium states for given climatic conditions depends on the type of bed slope: every time the slope changes sign over some distance (i.e., overdeepening occurs), another stable steady state can be added to the system. In practice, such effects are only significant for large valley glaciers that flow on beds with very small slopes.

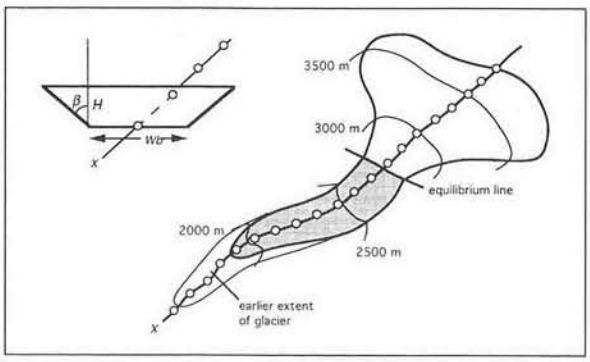

Figure 5.1 Characteristic geometry of a valley glacier. The equilibrium line separates ablation (dark) and accumulation (light) areas. A flowline can be identified to construct a grid for calculations. In the upper left-hand corner, a possible cross-section geometry is shown.

5.3 NUMERICAL MODELS OF VALLEY GLACIERS

In a model that is able to simulate the response of a valley glacier to climatic change, two components can naturally be distinguished: (i) an ice-flow model that calculates the movement of the ice in response to the action of gravity and (ii) a mass balance model that translates changing meteorological quantities into a changing mass balance field.

5.3.1 Modelling the ice flow

The basic assumption is that the geometry of the glacier under study permits the selection of one (or more) flowline(s), along which a grid for numerical calculations can be defined (Fig. 5.1). The grid points are normally spaced equally on the map projection of a glacier. This implies that every grid point represents the area of a glacier, or part of a glacier, in a specific elevation interval that generally varies for different grid points. The evolution of the glacier is calculated from the continuity equation (describing conservation of mass) integrated over depth and width. The basic equation reads:

$$\frac{\partial H}{\partial} = \frac{\partial (US)}{\partial x} + BW_s \qquad (5\text{-}1)$$

Here, S is the cross-sectional area perpendicular to the flow line, U the mean ice velocity in this cross-section, B the specific balance and Ws the width at the surface of the glacier. The horizontal coordinate (along the flow line) is x. Eq. (5-1) simply states that the cross-sectional area increases or decreases as a consequence of divergence of the mass flux along the flow line (the first term on the right-hand side) and the specific balance (second term). It is normally assumed that ice density is constant but it is not difficult to include a correction for lower density in the upper layers of the firn area. To arrive at a prognostic equation for ice thickness H, the cross-sectional shape has to be parametrized. Some modellers (e.g., Kruss, 1984) have used a power fit to describe the shape of the valley (i.e., $W_s = h^a$, where h is elevation above the bed at the center line and a is some

exponent chosen to fit observed valley cross sections (a may depend on x)). Others have used a trapezoidal shape (Fig. 5.1) which has one more degree of freedom (e.g., Oerlemans, 1986). In that case, Eq. (5-1) yields (W is the width of the bed)

$$\frac{\partial H}{\partial t} = \frac{-1}{W+2\mu H} \frac{\partial}{\partial x} \left[(W+\mu H)UH\right] + W; \mu = tan\beta \quad (5\text{-}2)$$

The calculation of the mean ice velocity U forms the heart of the matter. There are two contributions: one from internal deformation and one from sliding over the bed. For simulating the global behaviour of large valley glaciers, the following approach is quite acceptable (Budd *et al.* 1979):

$$U = U_d + U_s = f_d H S_d^3 + \frac{f_s^* S_d^3}{P} = f_d H S_d^3 + \frac{f_s^* S_d^3}{(H+\varepsilon)} \quad (5\text{-}3)$$

The subscripts d and s refer to deformation and sliding respectively. S_d is the driving stress (= sH; s surface slope). The parameter f_d is a generalized viscosity. Sliding is a highly controversial issue. A number of modellers have assumed that the sliding velocity is proportional to the third power of the driving stress as well, but divided by the water pressure at the bed (P). Some have criticized this approach but it is a fact that good results have been obtained in terms of simulated glacier profiles and velocities. Normally, information is not sufficient to justify the inclusion of a distinct routine for calculating P. One then assumes that P is a fraction of the overburden from ice pressure, leading to the right-hand side of Eq. (5-3).

Attempts have been made to obtain a theoretical expression for the sliding parameter f_s, but no consensus has been reached. Both f_d and f_s should be considered as semi-empirical parameters, varying from glacier to glacier. Shape factors also come into play here. They should compensate for the fact that the total drag on glacier movement depends on the shape of the valley. Shape factors have been determined for various valley shapes (Nye, 1965*b*). However, in the light of the uncertainty in f_d and f_s, in many applications a shape factor might also be absorbed in the flow parameters, as these have to be chosen in an optimal way for every single glacier.

As an example, Fig. 5.2 shows a simulation of the profile of Nigardsbreen, Norway. In this case, the model was driven by the observed specific balance (averaged over a great number of years). Profiles are shown for (a) the full calculation (b) a run where the glacier width was taken constant in time and along the flowline, and (c) a run in which the width was a function of x only. The differences are quite large and demonstrate how important it is to have the correct parameterization of the three-dimensional geometry (cf. also Furbish and Andrews (1984)). In fact, for most glaciers, the firn basins have a complicated geometry and cannot be represented well by one flowline. The general philosophy is to have the hypsometry right and focus the ice-flow calculation on the lower part of the glacier.

5.3.2 Modelling the mass balance

Profiles of specific balance vary significantly from glacier to glacier, as illustrated in Fig. 5.3. Ablation gradients on glaciers in the relatively humid middle latitudes, however, tend to be rather similar (approaching values of between 0.007 and 0.01 mwe/m on the glacier tongues). Glaciers in arid and subpolar regions have smaller balance gradients. The large differences are in the accumulation rates. Balance gradients are mainly determined by altitudinal gradients in air temperature, surface albedo and solid precipitation.

There are several ways of driving an ice flow model in climatic change experiments. The simplest of these is the so-called linear Lliboutry model, which assumes that annual mass balance perturbations are independent of elevation (Lliboutry, 1974). The change in balance, say ∂B, is then related to one or more meteorological quantities through regression analysis (assuming that data sets are available). This method has been used by Greuell (1992) and Huybrechts *et al.* (1989) in studies of the Hintereisferner (Austria) and Glacier d'Argentière (France) respectively. In another approach, climatic change is represented as a shift of the mass balance profile (so also of the equilibrium line) up or down the elevation axis. For a linear bal-

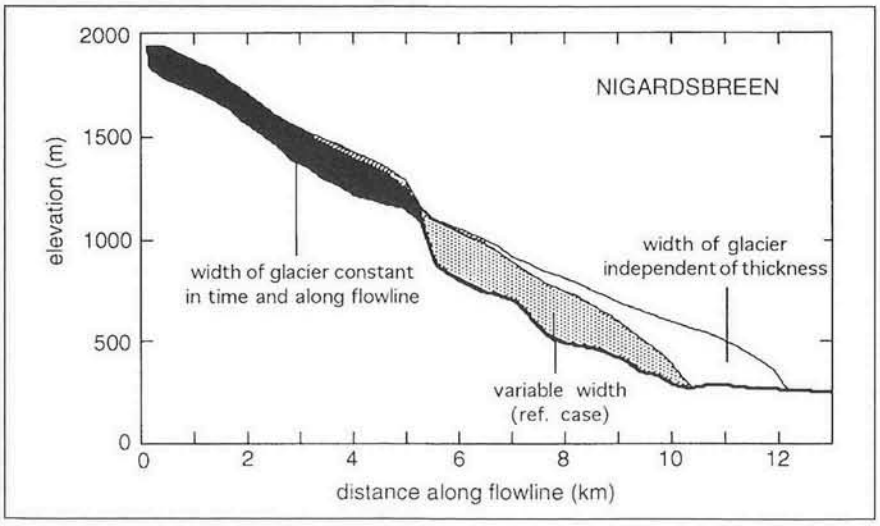

Figure 5.2 Calculated profile of Nigardsbreen. The reference case is compared with two other runs in which the geometry is not well represented.

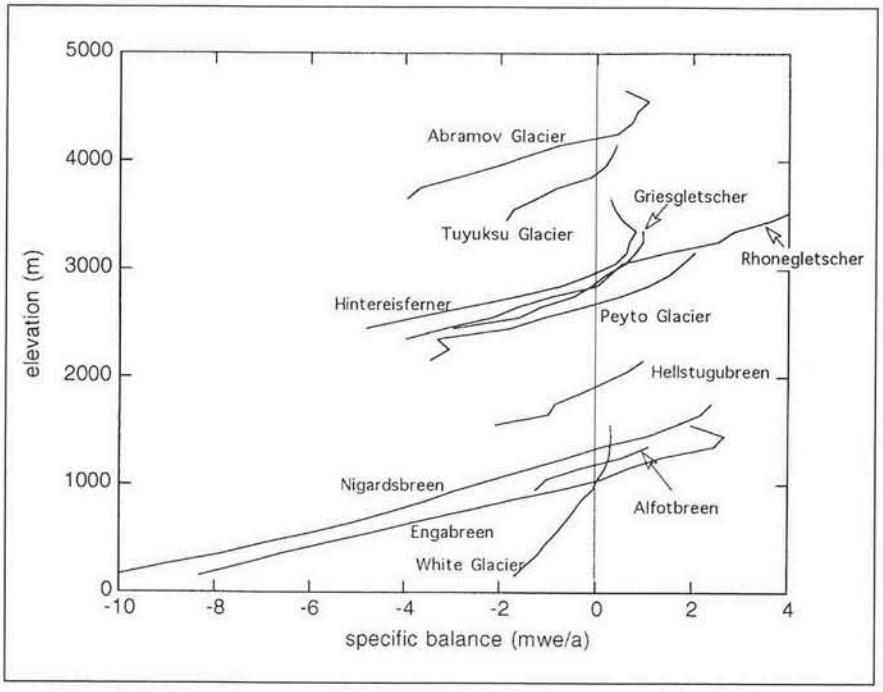

Figure 5.3 Profiles of measured specific balance for a number of valley glaciers in different climatic setting, showing the wide variety in balance gradients. Data from IAHS(ICSI)/UNESCO, 1967; IAHS(ICSI)/UNESCO,1977; IAHS(ICSI)/UNESCO,1985; IAHS(ICSI)/UNEP/UNESCO, 1988, with some additions.

ance profile, this would have the same effect as the linear model. For a profile in which the balance gradient is not constant, δB becomes a function of elevation. Shifting the mass balance profile with the equilibrium line has been done by a number of modellers: Oerlemans (1986), in the study of Nigardsbreen referred to above, and Stroeven *et al.* (1989), in simulating the historic front variations of the Rhonegletscher (Switzerland). In these studies, the basic assumption was that the effect of climatic change on the glacier could be described by $\delta E = c_1 + c_2\, G(t)$, where δE was the shift in the equilibrium-line altitude, G a climatic series (e.g., summer temperature, tree-ring width, etc.) and C_1 and C_2 constants that could be used to optimize the model results. Attempts to parametrize E in terms of climatological variables have been described by various authors, for example, Ahlmann (1924, 1933), Kerschner (1985), Ohmura *et al.* (1992) and Shumsky (1964).

Degree-day methods, introduced in glaciology by Finsterwalder and Schunk in 1887 (cited from Braithwaite, 1984), are based on the idea that the surface energy budget can be characterized solely by air temperature. The amount of energy available for melting over a given period of time is assumed to be proportional to the integral over max(0,T), where T is air temperature. The constant of proportionality, the degree-day factor, varies widely from glacier to glac-

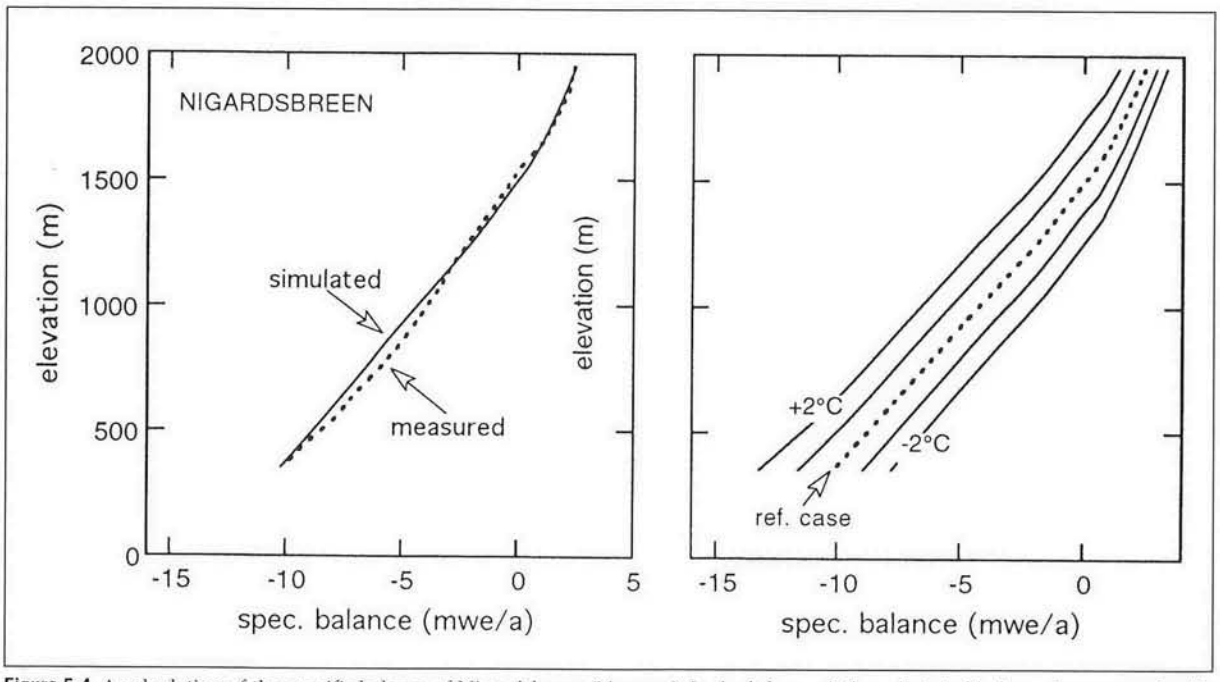

Figure 5.4 A calculation of the specific balance of Nigardsbreen (Norway). In the left panel, the calculated balance is compared with the mean over 25 years of observation. The right panel shows what the balance profile would look like for uniform changes in air temperature of +2, +1, −1, −2 °C.

ier because it has to account for all factors that cause variations in the radiation balance (notably albedo and cloudiness). It has been demonstrated that the degree-day method works well to explain year-to-year or month-to-month variations in melt on a specific location (e.g., Braithwaite and Olesen, 1989).

From a general point of view, full consideration of the energy budget of the ice/snow surface in space and time must be the best approach. Yet it is not always practical, nor worthwhile. For climatic change experiments, however, it is desirable to have all potentially important physical processes included in the calculation of a mass balance profile. Some modellers have studied shifts in the equilibrium-line altitude on the basis of changes in the energy balance at the equilibrium line (Kuhn, 1980; Ambach and Kuhn, 1985; Kuhn, 1989). This has provided valuable insight into how the individual energy balance components (turbulent exchange of sensible heat and latent heat, longwave radiation, shortwave radiation, effect of clouds, etc.) affect the mass balance. A natural extension of this work has been the development of numerical models that calculate the energy balance on a grid covering the entire elevation range of a particular glacier (Greuell and Oerlemans, 1986; Oerlemans and Hoogendoorn, 1989; Oerlemans, 1991, 1992; Oerlemans and Fortuin, 1992).

In its most complete form, the energy balance model calculates temperature and density profiles in the upper layer of the glacier. Here, we briefly discuss a simpler model that seems adequate for most valley glaciers. The basic equation reads:

$$B = \int_{year} \left\{ (1-f)\,min\left(0; \frac{-\psi}{L}\right) + P^{*} \right\} dt \qquad (5\text{-}4)$$

In Eq. (5-4), L denotes the latent heat of melt, Y the energy balance at the surface, and f the fraction of meltwater that does not run off but refreezes in the underlying snow pack when it is sufficiently cold. P* is the rate at which solid precipitation is added to the surface. The energy balance has the following components: solar radiation, atmospheric (infrared) radiation, turbulent fluxes of heat and moisture, and energy used for heating up the upper snow or ice layers. All these components are calculated in the model according to schemes that are used widely in boundary-layer meteorology.

The performance of an energy balance model in simulating glacier mass balance depends to a large extent on the way albedo is treated. Albedo varies strongly in space and time, depending on the melt and accumulation history itself. There is significant feedback involved, implying that a mass balance model designed to study the response to climatic change must generate the albedo internally. This is difficult, however, as so many factors are involved. The albedo depends in a complicated way on crystal structure, ice and snow morphology, dust and soot concentrations, morainic material, liquid water in veins, water running across the surface, solar elevation, cloudiness, etc. The best one can do is to construct a simple scheme in which the gross features broadly match available data from valley glaciers. The scheme will not be discussed here (cf. Oerlemans, 1991).

5.3.3 An example: Combined ice flow – mass balance model for Nigardsbreen

As an example, the application of the mass balance model to Nigardsbreen is discussed. With appropriate input from nearby climatic stations (annual mean temperature, seasonal temperature range, daily temperature range, annual mean cloudiness, annual mean humidity, constant precipitation rate through the year, altitudinal gradients in temperature and precipitation) the mass balance profile can be calculated. The simulation was done with a model version including the full daily cycle. The mass balance was generated on a one-dimensional grid, with grid points 100 m apart in terms of surface elevation. Several years of integration were needed to obtain an equilibrium mass balance profile. The results are summarized in Fig. 5.4. The simulation is quite satisfactory. It should be noted, however, that some model parameters, like altitudinal precipitation gradient or mean albedo of snow, can easily be varied within their range of uncertainty to give a good fit. This applies to many glaciers, as will be discussed later.

One of the advantages of a mass balance model of this type is that the effect of many processes can be studied explicitly. One can vary albedo, cloudiness, humidity, etc., to get a feeling for what is important and what not. Air temperature is certainly very important, as demonstrated in Fig. 5.4. An interesting result is the increase in balance gradients with increasing temperature. This is understandable from a physical point of view (and a degree-day model would produce this as well) but contradicts the Lliboutry model. However, this finding is hard to verify from observational data. First of all, one has to be careful in using interannual variability to quantify the effect of long-term changes in meteorological quantities. The albedo pattern on a glacier, for instance, is related to the ice velocity at the surface (whether it is up- or downwards) and will not follow the year-to-year changes in the balance regime. In existing data, one can find very different pictures of interannual variability, as illustrated in Fig. 5.5. There is ample room for further investigation here.

'Coupling' of ice flow and mass balance models is a straightforward matter, at least for valley glaciers. The calculated balance profile for a given set of meteorological input quantities can be expressed as a function of elevation then imposed on the ice flow model. As an example, consider the response of Nigardsbreen to stepwise climatic change. Fig. 5.6 shows what happens to the glacier for a uniform temperature change of +1°C (left) and –1°C (right), starting from a perfect, steady state. In the case of warm-

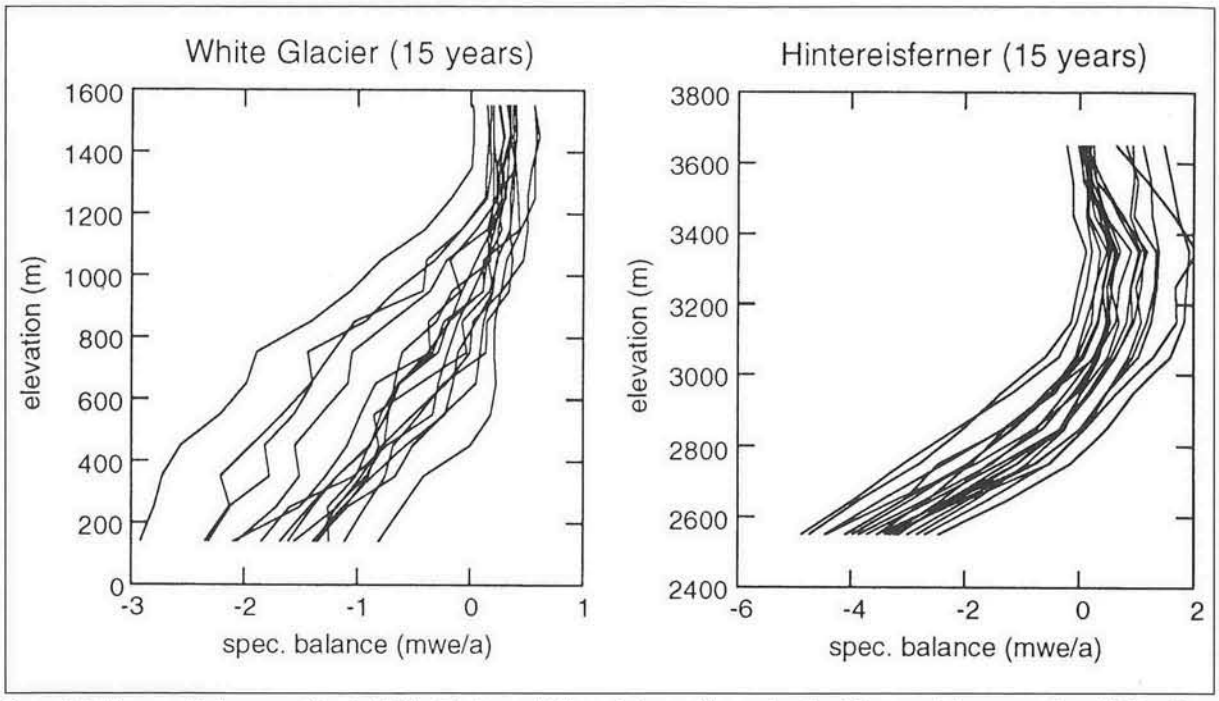

Figure 5.5 Measured balance profiles for White Glacier and Hintereisferner, illustrating the difference in interannual variability. Note the difference in horizontal scale.

ing, the retreat is dramatic (7 km in the final steady state). During the first 50 years, the average rate of retreat of the terminus would be 75 m/a. For a 1°C cooling, the glacier would advance more than 2 km.

The glacier's great sensitivity is due mainly to the specific hypsometry of Nigardsbreen. In the present configuration of the glacier, the area of the ablation zone increases very rapidly when the equilibrium line goes up. In addition, the fact that the mass balance gradient is larger/smaller for a warmer/cooler climate tends to enhance retreat/advance of the glacier terminus.

5.3.4 A note on calibration of models

From a scientific point of view, the modelling macroscopic natural systems (like a glacier) should be based on physical laws. At the same time, it has to be accepted that the degree of detail is limited and that, in many cases, a pragmatic approach to the use of such laws is indispensable for obtaining a workable model. It is not generally possible to make an absolute quantitative assessment of the state of balance of a natural system, unless it is very insensitive to input parameters (in that case, however, the absolute state of balance varies very little and is of no interest). The implication is that models need to be calibrated before they are used for sensitivity studies.

The mass balance model discussed earlier provides a good example. Owing to the considerable sensitivity of glacier mass balance to temperature, the model cannot be used to assess the absolute mean specific balance of a glacier. Even if the model were almost perfect, the uncertainty in the climatological input data would produce a large error bar on the mean specific balance. This does not mean, however, that tuned models would be less reliable with regard to sensitivity experiments. In fact, careful calibration, together with an explicit description of the dominant physical processes, is the best guarantee of a sensitivity study providing reliable results.

Flow-line models also have to be calibrated, either by adjusting the flow parameters, the bed profile and/or small adjustments in the mass balance field. In practice, it is very difficult to decide on the basis of field data to what extent a glacier is in equilibrium. This implies that the best way of calibrating a model is a dynamic approach in which simulated time-dependent evolution of the glacier is compared with observations. Time series for this purpose must come from long-term data collection at an international level.

Lastly, in many cases, calibration – or tuning – is not a process that determines the control parameters in a unique way. Different combinations of control parameters may give similarly good results. In such cases, the extent to which the sensitivity of the model depends on the specific choice of control parameters should always be checked. Fortunately, this dependence generally seems weak for (non-surging) glacier systems.

5.4 HISTORIC LENGTH VARIATIONS

Several valley glaciers have been studied with numerical models to gain an understanding of their basic response to climatic change (Variegated Glacier: Bindschadler, 1982; Griesgletscher: Bindschadler, 1980; Vernagtferner: Kruss and Smith, 1982; Lewis Glacier: Kruss, 1983, 1984; Nigardsbreen: Oerlemans, 1986, 1992; Rhonegletscher: Stroeven *et al.*, 1989; Glacier d'Argentière: Huybrechts *et al.*, 1989; Hintereisferner: Greuell, 1992). In some cases, integrations have been carried out with time-dependent forcing.

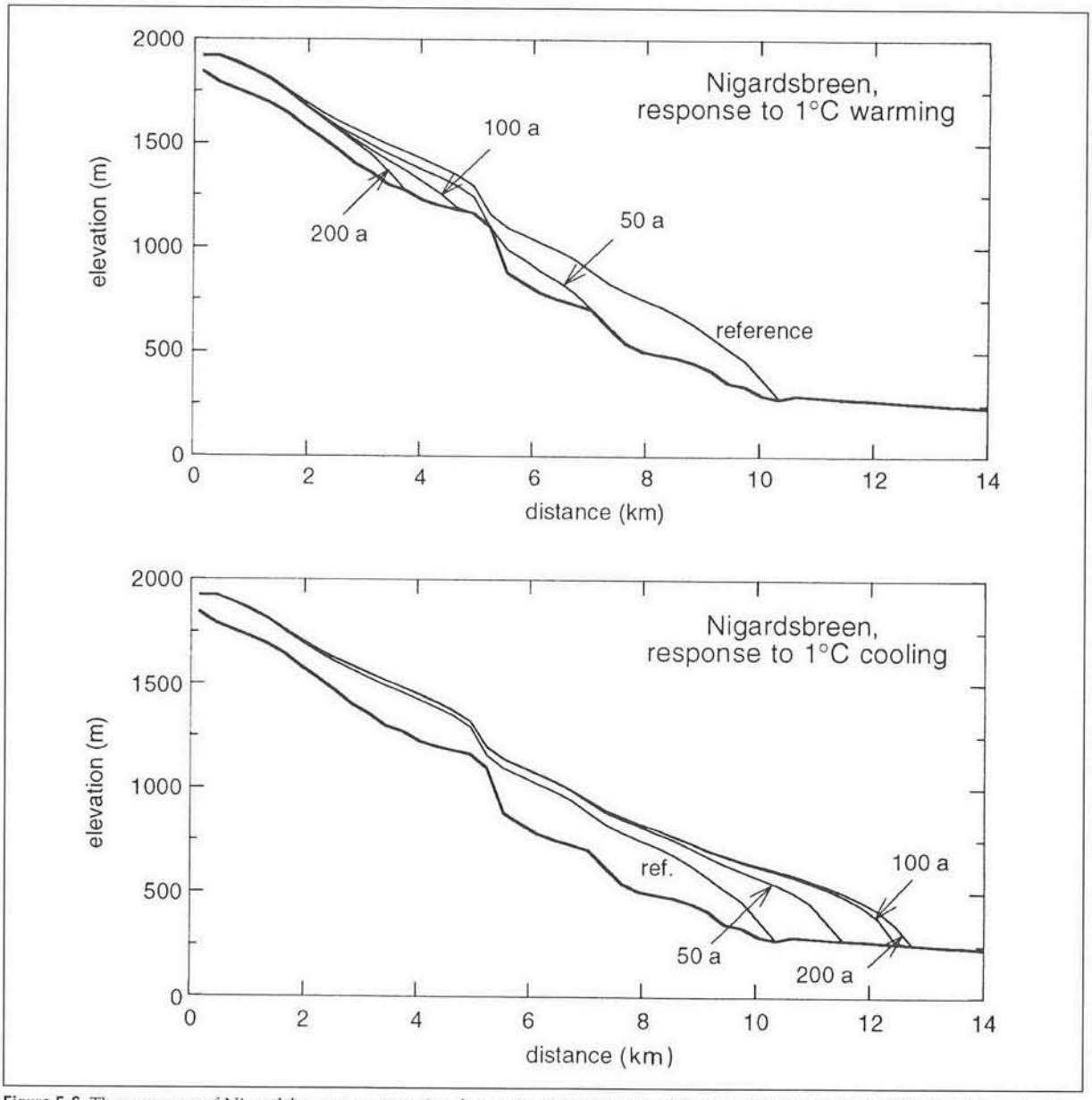

Figure 5.6 The response of Nigardsbreen to a stepwise change in air temperature. These results were obtained by forcing an ice flow model with the outcome of a mass balance model (after Oerlemans, 1992).

Unfortunately, results have not been published in a uniform format and it is hard to make a comparison between the different studies. For four glaciers, it is possible to derive what can be considered as a basic property of a valley glacier: the length in dependence of mean air temperature. Fig. 5.7 shows how, according to numerical models, the length of Nigardsbreen, Glacier d'Argentière, Hintereisferner and Rhonegletscher would vary with changes in temperature if the glaciers were in equilibrium. $\delta T = 0$ reflects the present-day climatic situation. Numerical experiments for changes in equilibrium-line altitude (Rhonegletscher) or for uniform changes in specific balance (Hintereisferner, Glacier d'Argentière) were 'translated' into changes in mean air temperature. In spite of this, the comparison is useful and illustrates nicely how the geometry of glaciers affects climate sensitivity.

First of all, for an individual glacier, $dL/\delta T$ varies greatly. Consider, for instance, the curve for

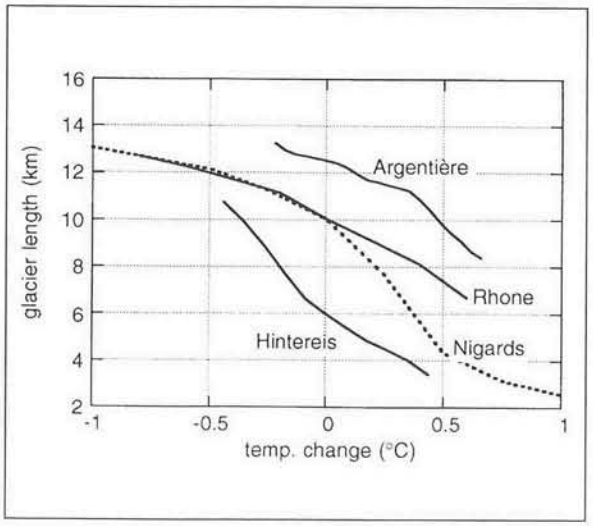

Figure 5.7 Equilibrium length of four valley glaciers as a function of a change in annual mean air temperature. Based on Oerlemans, 1986; Stroeven *et al.*, 1989; Huybrechts, 1989; Greuell, 1992.

Nigardsbreen: in the range $0 < \delta T < 0.5°C$, a value of 12 km per degree is found, in the range $-0.5 < \delta T < 0$, this is 4 km per degree. Hintereisferner shows extreme sensitivity to slightly colder conditions than in the 'present' climate. The figure also makes clear that values of $dL/\delta T$ for the present climate differ substantially from glacier to glacier (by a factor of two for the glaciers shown). It is clear that one cannot model all valley glaciers in the world for which the length has been measured. Nevertheless, Fig. 5.7 suggests that it is worthwhile to model a selection of individual glaciers for which long records of glacier length are available.

Many modellers have expressed the opinion that the worldwide retreat of glaciers since the 19th century is a response to the warming over that period as registered in the instrumental records. In the very global sense, this is probably true. It is a fact, however, that attempts to simulate the length records of individual glaciers by forcing numerical flow models with (proxy) climate records have not been very successful. For instance, using a tree-ring index from northern Scandinavia to force Nigardsbreen yields a minimum glacier stand around 1900 AD and slight advance afterwards. In reality, Nigardsbreen has been retreating strongly during most of the 20th century. Using the Central England temperature series as forcing makes it worse (Oerlemans, 1986). A similar story applies to the Rhonegletscher. Stroeven *et al.*, (1989) imposed 7 forcing functions (including the extended 'Basler Temperaturreihe' and Pfister's summer temperature and winter precipitation series). None yields a maximum glacier stand in the mid-19th century and the substantial retreat afterwards. Also, the retreat of the Hintereisferner (Greuell, 1992) and the Glacier d'Argentière (Huybrechts *et al.* 1989) has been of much greater magnitude than can be simulated with existing climatological series. Interestingly, when using a global temperature forcing derived from a volcanic index and, since 1850 AD, enhanced greenhouse forcing, the calculated glacier response is somewhat more realistic (Oerlemans, 1988).

It is unlikely that the shortcomings of the ice flow model can explain the discrepancies between the observed and simulated glacier length discussed above. Most modellers agree that the reconstruction of the mass balance history is the weak point. Possible explanations are: (i) The long-term signal in (proxy) climate records is unreliable. (ii) Trends in precipitation have been more pronounced and of a larger spatial character than normally assumed by climatologists. (iii) There have been significant global changes in the surface radiation budget that have been registered by glaciers but which are only weakly reflected in surface temperature series. (iv) Existing mass balance models are inadequate.

5.5 GLOBAL MODELLING

The previous discussion dealt with the modelling of individual glaciers. However, it is impossible to model all glaciers in the world in such detail, so other methods have to be used when the global behaviour of glaciers (e.g., with regard to the problem of sea-level change) is the subject of interest. Two basic questions can be formulated. Firstly, how does the mean specific balance over a sample of glaciers change, with respect to some kind of reference state, for a given change in one or more meteorological quantities? Secondly, to what ice volume changes does this lead in the course of time?

5.5.1 The mean specific balance

To answer the first question, it is not necessary to consider glacier mechanics. The problem is of a purely meteorological nature. To calculate the initial change in ice volume, all one has to know is the change in mean specific balance and the total glacier area in the region considered. For a small region in which climatic characteristics are relatively constant, the climate sensitivity of the specific balance can be considered constant as well. However, such regions are usually microscopic. It is more practical to try to express climate sensitivity in terms of basic climatic quantities such as annual mean air temperature or precipitation, which then allows an extrapolation to larger regions. Oerlemans and Fortuin (1992) have tried this approach in a model study of 12 glaciers for which good mass balance observations exist. A summary of their results is given below.

All 12 balance profiles (for Abramov Glacier, Tuyuksu Glacier, Griesgletscher, Rhonegletscher, Hintereisferner, Peyto Glacier, Hellstugubreen, Nigardsbreen, Engabreen, Ålfotbreen, White Glacier, Devon Ice Cap) could be simulated well with the energy balance approach described earlier. A study of climate sensitivity, in which uniform changes in temperature and precipitation are imposed on the mass balance model, then revealed that climate sensitivity is determined mainly by the amount of annual precipitation, i.e., by the continentality of the glacier. This has been noted earlier of course. The results of energy-balance modelling merely support this view. Fig. 5.8 shows the change in equilibrium-line altitude and mean specific balance (∂B_m) for the 12 glaciers, plotted in dependence on the annual precipitation. The difference between the subpolar glaciers (White Glacier and Devon Ice Cap) and the maritime glaciers of Norway is impressive: a factor of 6 in the change of mean specific balance for a uniform 1°C warming.

Glaciers with large mass turnover tend to exist in relatively low elevations with a high air temperature. The ablation season thus lasts longer, which is the main reason for greater sensitivity. This is illustrated further by the fact that when the summer temperature only is increased in the model calculations, cli-

mate sensitivity does not further increase for annual precipitation greater than about 1 m/a.

The response to a 10% change in precipitation shows two regimes. For P < 4 m/a, the albedo feedback is important and makes the increase in specific balance larger than the added mass. For very large precipitation regimes, the fact that part of the additional precipitation falls as rain dominates the albedo feedback, so the increase in specific balance is less than the added mass. In a scenario where temperature is increased by 1°C and precipitation simultaneously by 10%, the change in mean specific balance is still strongly negative.

The differences in climate sensitivity have important consequences for the global sensitivity of glaciers and small ice caps. To make an estimate of this, it is practical to express the relation between P and ∂B_m in a logarithmic fit (see Fig. 5.8):

$$\partial B_m = -0.512 - 0.662 \log(P) \qquad (5\text{-}5)$$

Now, suppose that each glacierized region can be characterized by glacier area A_k and annual precipitation P_k, where the index k refers to the specific region. The global mean change in specific balance for a 1°C can then be defined as:

$$\delta G_m = \frac{1}{A_{tot}} \sum_{k=1}^{k} A_k (-0.512 - 0.662 \log P_k) \qquad (5\text{-}6)$$

Working this out by dividing all glaciers and small ice caps (outside Greenland and Antarctica) into 100 glacierized regions and assigning characteristic values of P to those regions then yields
$\partial B_m = -0.40$ mwe.
This is significantly less than values normally quoted, which are in the −0.5 to −1.0 mwe range (Kuhn, 1993). The reason for this is the large share of the relatively dry subpolar ice caps with their lower sensitivity. The exercise can be repeated with a change in precipitation included. For instance, assuming that, for 1°C warming, precipitation increases in proportion to the saturation vapour pressure, this leads to $\partial B_m = -0.31$ mwe.

5.5.2 Ice volume

The question of how the volume of all glaciers and small ice caps has varied over recent centuries and how it will vary in the near future is a difficult one. Pioneering work on past changes has been done by Meier (1984). He has estimated mass loss from gla-

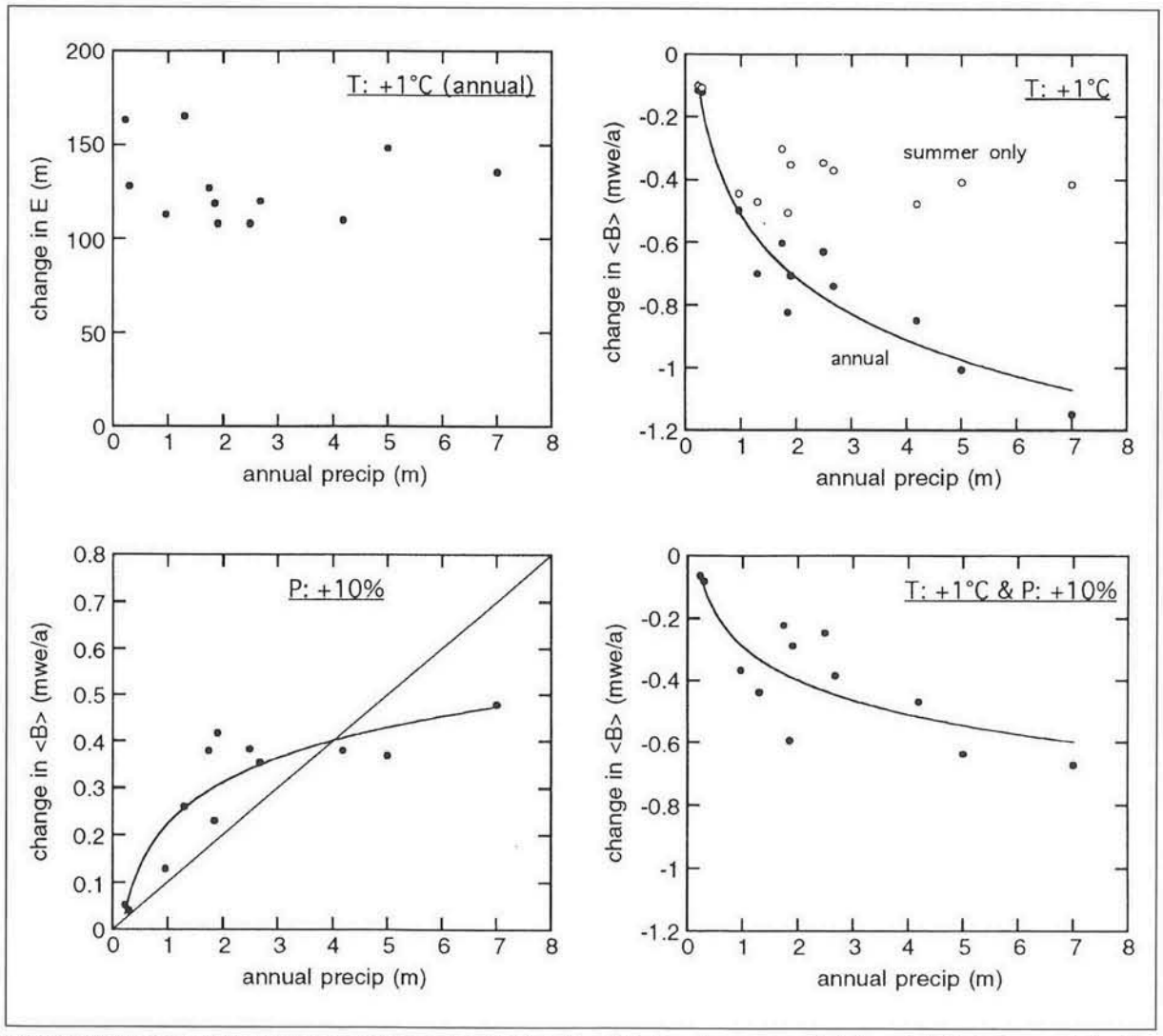

Figure 5.8 Climate sensitivity of the mass balance (equilibrium line altitude and mean specific balance) of 12 glaciers as derived by energy-balance modelling (Oerlemans and Fortuin, 1992). Curves are logarithmic fits.

ciers by extrapolating various kinds of measurements for the period 1900–1961. The extrapolation to all glaciers is the most critical step, of course, and this was done by assuming that the change in mean specific balance was proportional to the 'activity' of glaciers, defined as the difference between summer and winter balance. In this way, glaciers in wet regions (e.g., in Alaska) have a larger weight than the subpolar ice caps. In fact, this assumption is supported by the results of mass balance modelling described in the previous section. Meier considered 13 glacier regions and arrived at a mean loss of 19.15 mwe of ice, corresponding to a mean change of specific balance of about 0.31 mwe/a. The equivalent amount of sea-level rise is 28 mm for the period 1900–1961.

In attempts to predict sea-level change in the coming centuries, very schematic methods have been used to calculate the contribution of glaciers. When the period of interest exceeds a few decades, the glacierized area will change and affect estimates of mean specific balance. Oerlemans (1989*b*) and Wigley and Raper (1993) have used very simple schemes to relate equilibrium glacier volume to global mean temperature. There is no sound basis for the approach followed in these papers, however.

Recently, Haeberli and Hoelzle (1995) have attempted to design a scheme with very simple glacier dynamics. Without modelling individual glaciers in detail, simple dynamics based on estimates of glacier length, mean slope, mean (or maximum) thickness are included in the scheme. It has the advantage of enabling calculations to be done for all glaciers for which basic quantities as documented in the *World Glacier Inventory* are measured, while taking into account that the glaciers have different response times and climate sensitivity.

5.6 OUTLOOK

The climatic interpretation of field data (proxy or modern) through process-based modelling of glaciers has only just begun. One of the central problems for the near future will be the extrapolation of knowledge obtained from a small and not necessarily representative sample of glaciers to a regional or world scale. One can envisage two ways of tackling this problem:

(i) Detailed modelling of a set of glaciers. The results should then be used to quantify sensitivity parameters in terms of geometric and climatological characteristics. With appropriate data from the World Glacier Inventory, this then allows extrapolation to larger samples of glaciers (e.g., Meier, 1984, Oerlemans and Fortuin, 1992; Oerlemans, 1994).

(ii) Simple modelling of basically all glaciers. Here, the dynamics describing glaciers are reduced to a minimum but are still based on physical considerations (e.g., Haeberli and Hoelzle, 1995). Such simple schemes for dynamics contain many empirical parameters. These have to come from analysis of field data, as well as from results of more detailed modelling studies.

Apart from this, interest will remain in modelling glaciers in as much detail as possible for local applications (runoff for hydropower, calving in lakes, ice avalanching, etc.). Also, all glaciers for which long historic records on a terminus position exist should be modelled in some detail. This will significantly improve the climatological interpretation of such records.

Finally, the interplay between remote sensing and physical modelling deserves more attention. Mass balance and ice flow models could be used as assimilation tools for data of different kinds (e.g., visible range data on albedo, microwave data on glacier faces, both feeding an energy balance model to calculate summer melt).

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6 Use of remote-sensing techniques

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6.1 INTRODUCTION

The geosphere and the biosphere are the two principal components of the dynamic Earth system. The geosphere consists of the lithosphere, atmosphere, hydrosphere and cryosphere. The cryosphere includes frozen water in four principal forms: snow, floating ice (sea, river and lake), glaciers and frozen-ground (permafrost) features.

Measurement of the world's glaciers, especially direct or indirect measurement of decadal and long-term changes in the mass balance of the Earth's glaciers, can play a quantitatively significant role in the accurate assessment of global environmental change. Glaciers are present on all continents except Australia; they provide an integrated response to long-term variations in regional temperature and precipitation by increasing or decreasing their total mass. Changes in mass balance eventually result in the advance or retreat of the terminus of a glacier or the margin of an ice cap or ice sheet. Generally speaking, smaller glaciers respond more rapidly to changes in climate, although even large ice masses, such as ice sheets, ice caps and ice fields,[1] because they are actually aggregates of many independent glaciers, can also show annual or decadal changes in their margins and in the termini of their outlet glaciers. Glacier movement and changes in mass balance often occur quickly enough to be observed by humans; annual- and decadal-scale changes in the glacier's terminus or margin and decadal-scale changes in volume may be measured in the field (ground-based observations) and by imaging and non-imaging remote sensors mounted in an aircraft or spacecraft.

At present, the estimated volume and surface area of glacier ice on Earth are 32,950,000 km^3 and 15,888,209 km^2 respectively (covering about 10.7% of the land area and 3.1% of the Earth's surface) (Allison *et al.*, 1989; Haeberli *et al.*, 1989; Kurter *et al.*, 1991; British Antarctic Survey *et al.*, 1993; Østrem *et al.*, 1993; Williams and Hall, 1993). Antarctica contains about 30,110,000 km^3 and 13,612,690 km^2

1 Armstrong *et al.* (1973) considered an ice field (icefield, ice-field) to be a variety of ice cap and thus not warranting distinction as a separate term. The authors prefer to differentiate between an ice cap (large domal ice mass covering the terrain) and an ice field (large ice mass incompletely covering the terrain) from which one or more outlet glaciers emanate. They follow, with some modification, the definitions used by IAHS(ICSI)/ UNEP/UNESCO (1977) and Bates and Jackson (1987).

(Drewry *et al.*, 1982; Drewry, 1983; British Antarctic Survey *et al.*, 1993); the Greenland ice sheet contains about 2,600,000 km^3 and covers 1,736,095 ± 100 km^2; other glaciers in Greenland contain about 20,000 km^3 and cover 48,599 ± 100 km^2 (Weidick, 1995). Altogether, the Greenland ice sheet represents 7.9% of the total global volume of glacier ice, the Antarctic ice sheet and associated ice shelves representing 91.4% of the volume. The Greenland and Antarctic ice sheets account for 99.3% of the volume and 96.8% of the area covered by glaciers on our planet. All of the other glaciers on Earth account for only 0.7% of the total volume of glacier ice.

Glaciers are the largest reservoir of freshwater on Earth (77% of the total) and represent an important source of surface water for irrigation and hydroelectric power generation in many countries. Change in global volume of glacier ice is directly linked to changes in sea level (Allison, 1981; Warrick *et al.*, 1993), with 400 km^3 of glacier ice equivalent to a sea-level change of 1 mm (Williams and Hall, 1993). During the past 17,000 years, sea level has been 125 m lower than at present (Fairbanks, 1989); the remaining volume of glacier ice is sufficient, if melted, to raise sea level about another 75 m (Williams and Hall, 1993).

Owing to the immense area and volume of the Earth's glaciers, large-scale mapping and formal monitoring of changes in mass balance, area and termini, (UNESCO, 1969; IAHS(ICSI)/UNEP/UNESCO, 1993) and formal inventorying (UNESCO, 1970; IAHS(ICSI)/UNEP/UNESCO, 1977; 1978; 1983; 1989) of glaciers have been carried out on only a comparatively few of the world's glaciers, virtually all of which fall within 3.2% of the total area (0.7% of the volume) of glaciers on our planet. Precision mapping of glaciers and monitoring and the fluctuation of glacier termini were started in the 19th century in Europe; the longest continuous mass balance measurements of a glacier using the direct glaciological method, Storglaciären in Sweden, were initiated in 1945–1946.

6.2 ACHIEVEMENTS MADE WITH LONG-TERM DATA

To facilitate an increase in the number of mapped, monitored and inventoried glaciers, various types of remote-sensing instruments have been used to measure salient parameters. These parameters include surface area (Williams, 1983*a*), volume (Drewry *et al.*, 1982; Drewry, 1983; Björnsson, 1988), termini fluctuations (Hall *et al.*, 1992) (Fig. 6.1), position of the transient snow-line (Østrem 1975; Østrem and Haakensen, 1993), surface velocity (Lucchitta and Ferguson, 1986; Bindschadler and Scambos 1991), glacier facies (Williams *et al.*, 1991), inventories (Fig. 6.2a) and changes in area (Fig. 6.2b) (Williams, 1986*a*). Measuring glacier parameters remotely has

Figure 6.1 Comparison of Landsat MSS (1426-12070;22 September 1973) and TM (51242-11530; 26 July 1976) images of the Sídujökull outlet glacier, Vatnajökull, Iceland, showing change in its termini between 1973 and 1976. Supraglacier morainic debris on the terminus of the receding outlet glacier has essentially the same reflectivity of proglacial outwash, making precise delineation of a glacier margin sometimes difficult (after Hall *et al.*, 1992).

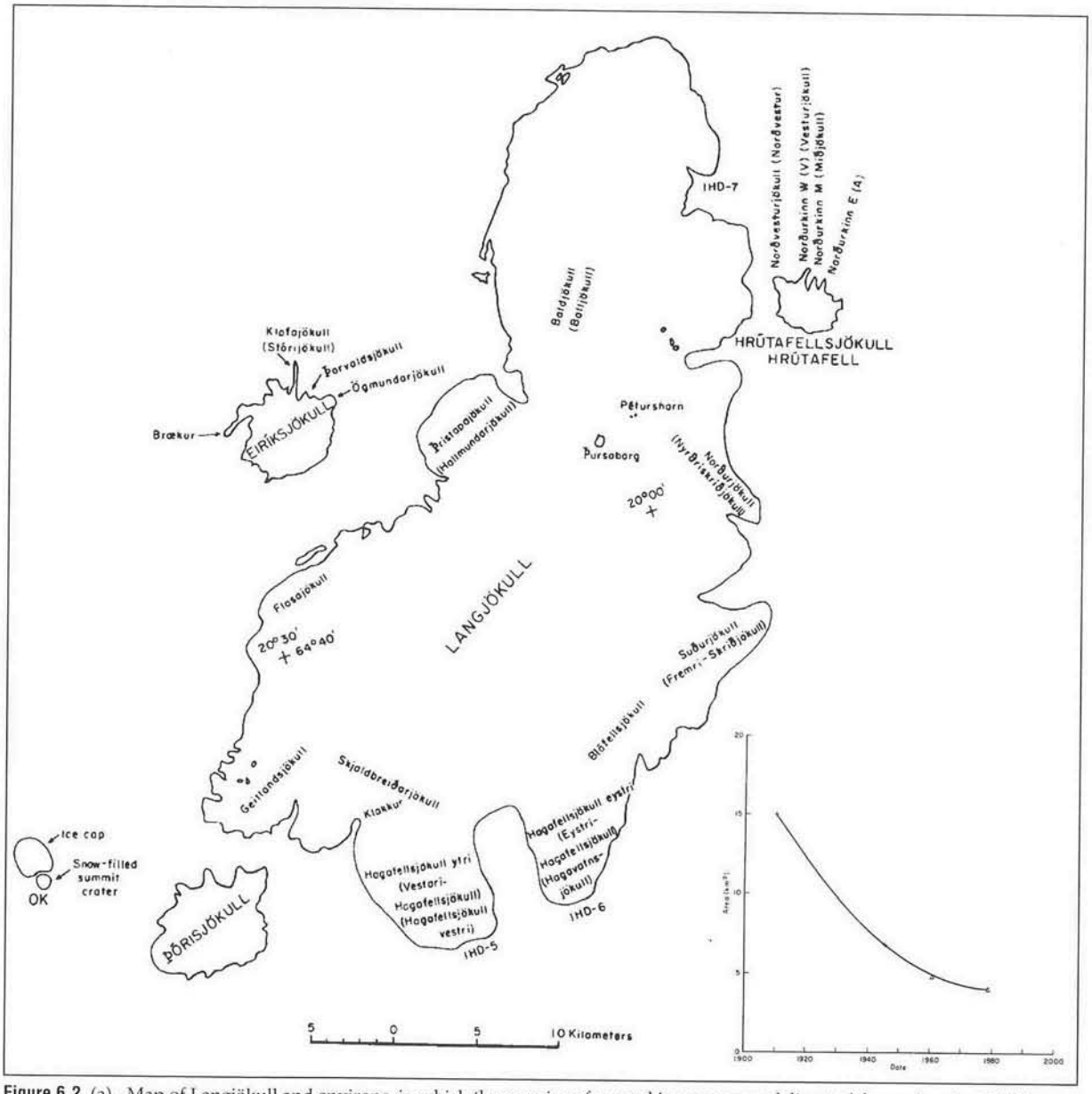

Figure 6.2 (a) - Map of Langjökull and environs, in which the margins of several ice caps were delineated from a Landsat MSS image (22045-12131; 28 August 1980); a preliminary inventory of ice caps and outlet glaciers are also presented. Three International Hydrological Decade (IHD) glaciers are shown (IHD 5-7).
(b) - Graph of Ok Glacier, Iceland, showing the decrease in area based on 1910 and 1945 topographic maps, a 1960 vertical aerial photograph and a 1980 Landsat image (after Williams, 1986*a*).

produced mixed results but glaciologists will continue to test and, when appropriate, use data from newly developed remote sensors carried by aircraft and satellites to further increase the global capability to monitor changes in glaciers (Fig. 6.3).

Shortly after the invention of the camera in the mid-19th century, Shaler and Davis (1881) used one in the first systematic, ground-based documentation of glaciers (and glacial land forms). Nigardsbreen, in Norway, was first photographed in 1864 (Østrem and Haakenen, 1993). In the late 20th century, Williams (1986*b*) used satellite imagery to review the variety of glaciers and glacial landforms on Earth that can be mapped and monitored from imaging sensors on satellite platforms. Terrestrial (ground-based) photography and stereophotogrammetry have been used by various scientists since the late 19th century to document the spatial position of glacier termini. Glaciers are still being documented in this way in many places.

The camera achieved its greatest success in the mapping of glaciers through aerial stereophotogrammetry. Aerial photogrammetry was eventually adopted by virtually all of the national mapping agencies after World War II for basic topographic mapping, replacing ground-based, planetable-surveying methods. Planetable-surveying techniques certainly produced excellent maps but the method was too slow and labour-intensive. To make accurate maps, aerial photogrammetry requires the placement of geodetic-control stations within the region being mapped so that they are visible on overlapping, stereoscopic, vertical (metric) aerial photographs.

In addition to vertical aerial photography, radio-

echo sounding, often used in ground and airborne surveys, has been the other remote-sensing method used very successfully for glaciological studies (Drewry *et al.*, 1982; Drewry, 1983; Björnsson 1988). Radio-echo sounding is also being used increasingly for studies of the internal structure of glaciers, their thermal conditions and conditions at the glacier-bedrock interface. Before radio-echo sounding was developed, three laborious methods were used to determine the thickness of glaciers: drilling, seismic surveys and local gravity surveys. The combination of metric aerial photography and geodetic ground-control to construct accurate topographic maps of small glaciers and radio-echo sounding to produce ice-thickness maps provides the baseline data for monitoring changes in surface area (including changes in termini) and volume and to complete glacier inventories.

Except for parts of Antarctica, stereoscopic, vertical aerial photographs have been acquired at least once for most of the world's glacierized regions. These data are generally available to glaciologists, although some countries place military security restrictions on their accessibility; in addition, accurate geodetic control is lacking for many regions.

National mapping agencies use aerial photogrammetry to produce maps of glacierized regions within their national domains or areas of interest (such as Antarctica). Aerial photographs for topographic-mapping purposes are not usually acquired at the end of the melt season (end of the glacier budget year); thus, the areal extent of glaciers shown on such maps may be inaccurate. This inaccuracy is caused by snow cover that can mask the margins of glaciers, as well as snow patches that may also be mapped as glaciers by untrained photogrammetrists who are unable to tell the difference. Stereoscopic, vertical aerial photographs are, however, used by field glaciologists as the primary remote-sensing data source for mapping, monitoring and inventorying glaciers. In Norway, accurate topographic maps have been made of many ice caps and other glaciers; Østrem and Tvede (1986) analyzed maps of glaciers as a source of climatological information; Haakensen (1986) used accurate maps of ice caps compiled at different times to confirm conventional, ground-based mass balance measurements.

Although vertical aerial photographs and published maps have been the basic sources of material used by glaciologists to map, monitor and inventory glaciers until recently, data from other remote-sensing instruments, on both airborne and satellite platforms, are now being used more frequently to supplement or complement photographs and maps. In many of the Earth's glacierized regions, remotely-sensed data from satellites are the only data available to glaciologists for mapping, monitoring and inventorying glaciers. This is especially true in the polar regions, where satellite remote sensing has been and will continue to be used to map (U.S. Geological Survey, 1991) and (or) monitor changes in the cryosphere (Williams, 1983*a*; 1983*b*; Zwally *et al.*, 1983*b*;

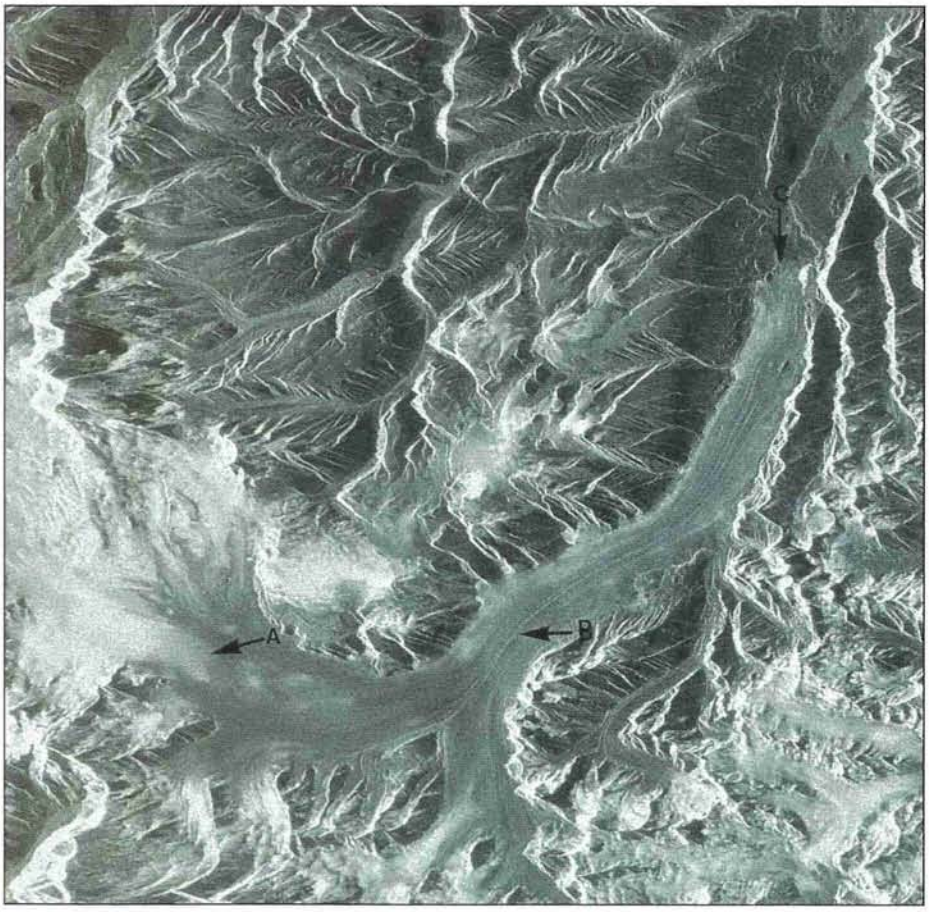

Figure 6.3 ERS-1 SAR image of the Nabesna Glacier, central Wrangell Mountains, Alaska. The arrow at 'A' points to the snow-line; 'B' points to the medial moraine; and 'C' points to the glacier terminus. SAR image courtesy of European Space Agency; image processed by the Alaska SAR Facility at the University of Alaska, Fairbanks, Alaska (from Hall, in press).

Hall and Martinec, 1985; Ferrigno and Gould, 1987; Parkinson *et al.*, 1987; Massom, 1991; Thomas, 1991; and Williams *et al.*, 1995).

Østrem (1975) and Krimmel and Meier (1975) were among the first to recognize the potential of satellite images for glaciological studies. In this paper, it is impossible to cite all of the scientific papers that have been published on studies of glaciers with satellite images; the following citations will provide a few examples; a review paper by Williams and Hall (1993) provides additional examples. Thorarinsson *et al.*, (1974) and Williams (1987) used Landsat images in studies of the Vatnajökull Ice Cap, Iceland. Desinov *et al.*, (1987) used satellite photographs acquired by Salyut-6 to study the Southern Patagonian ice field, South America. Dowdeswell (1984) and Dowdeswell and McIntyre (1987) used Landsat images in their investigations of glaciers in Svalbard, Norway. Dwyer (1993) and Weidick (1995) used Landsat images to study glaciers in Greenland. Williams and Ferrigno (1994), in association with more than 60 glaciologists, are using Landsat images from the 1970s to produce an 11-volume Satellite Image Atlas of Glaciers of the World in the U.S. Geological Survey Professional Paper series of books; five have been published (Swithinbank, 1988; Allison *et al.*, 1989; Kurter *et al.*, 1991; Østrem *et al.*, 1993; Weidick, 1995).

6.3 PRESENT UNDERSTANDING OF TECHNIQUES

Remote-sensing instruments can be divided into two broad classes: imaging and non-imaging. Imaging sensors include the camera (essentially an analogue, broad-band spectrophotometer), optical-mechanical scanners, passive microwave and active microwave (radar). Non-imaging sensors include radar and laser altimetry, radio-echo sounders and gravimeters. Each remote-sensing instrument operates at one or more bands or frequencies of the electromagnetic spectrum. Imaging sensors produce images in relatively broad regions of the visible, near-infrared, short-wave infrared and thermal-infrared bands of the electromagnetic spectrum. Most remote-sensing devices are carried in aircraft or spacecraft but some may be used on the ground (e.g., camera, radio-echo sounders).

6.3.1 Imaging sensors

Polar-orbiting satellites, such as NOAA satellites, Landsat and SPOT (discussed below) pass over the same spot on the Earth's surface at periodic intervals. In the case of the U.S. National Oceanic and Atmospheric Administration's NOAA meteorological satellites, which overfly the same area of the Equator every 6 hours (more frequently at higher latitudes because of convergence of orbits), advanced very high resolution radiometer (AVHRR) images have a picture-element (pixel) size (pixel resolution) of about 1 km at nadir. The (U.S.) Defense Meteorological Satellite Program (DMSP) acquires images with a pixel resolution of about 500 m at nadir. Images from meteorological satellites, because of the high frequency of coverage and thus greater probability of encountering cloud-free conditions, have been successfully used to locate the position of the transient snow-line on glaciers (Østrem *et al.*, 1979). When such information is combined with topographic maps or a digital elevation model (DEM), the transient snow-line altitude can be ascertained. If the images are acquired at the end of the melt season and a time series of mass balance measurements has been made on the ground of a specific glacier, then the altitude of the snow-line at the end of the melt year can be used to determine whether the glacier has had a positive or negative mass balance for the past budget year (Østrem, 1975; Østrem and Haakensen, 1993).

In 1972, the first in a series of U.S. Landsat satellites was launched with the multispectral scanner (MSS) on board; it digitally imaged features on the Earth's surface between about latitudes 82° N and 82° S in the visible and near-infrared parts of the electromagnetic spectrum with a pixel resolution of about 80 m and an 18-day repeat cycle. The Landsat thematic mapper (TM) was first launched in 1982; it has visible, near-infrared, short-wave infrared and thermal-infrared sensing capabilities at a pixel resolution of about 30 m for six of its seven bands, 120 m for data from the seventh (thermal-infrared) band and a 16-day repeat cycle. The 16-day repeat cycle permitted mapping of most (but not all) glacierized areas of the Earth over a period of many years. However, because of persistent cloud cover in certain regions, some of the Earth's glaciers have not yet been mapped by Landsat sensors. Part of the Landsat programme was transferred from the U.S. Government to the EOSAT Corp. in 1985 (Williams, 1991); image acquisition then became less systematic and repetitive, and image costs increased. Sequential images of all glacierized areas of the Earth and single-time coverage of large regions were not always available and their cost became prohibitive to the glaciological community. As a result, Landsat data and other than pre-EOSAT data are presently under-utilized by the glaciological community.

Both the U.S. and Russian governments have, for the past 30 years, operated an extensive series of Earth-orbiting satellites that acquire remotely-sensed data of the Earth's surface in various parts of the electromagnetic spectrum to support strategic-intelligence requirements. Except for satellite photographs collected by the U.S. Intelligence Community during the 1960s and early 1970s (U.S. Geological Survey, 1995; McDonald, 1995*a*; *b*), all of the U.S. data remain classified and are unavailable to the scientific community unless appropriate high-level security clearances are obtained; the classified data must not be discussed or published; thus, they have limited scientific value.

The Russian government, through its Soyuzkarta marketing arm, has been selling previously classified imagery, both multispectral imagery and photography; the satellite photographs have lines-per-millimeter (lpm) ground resolution of a few meters in some cases. Kostka and Sharov (1995) cited a photographic resolution of 200 lpm for stereophotographs acquired by the Russian Priroda MK-4 multispectral film camera. Soyuzkarta data are also expensive, so they are not widely used by glaciologists. The (U.S.) Government Applications Task Force (GATF) continues to review classified data sets, including satellite images and photographs, to determine what additional data, if any, can be released to the scientific community.

The French Satellite Pour l'Observation de la Terre (SPOT) series of satellites, first launched in 1986, provides panchromatic (10-m pixel resolution) and multispectral (20-m pixel resolution) images of the Earth's surface, which are useful for glacier-monitoring studies. Data at the 10-m resolution are especially useful for detailed studies of a particular glacier. SPOT also offers stereoscopic coverage of selected areas and, upon request, can schedule the acquisition of stereoscopic coverage of any area on Earth between about latitudes 83.5° N and 83.5° S (Jonathan F. Williams, SPOT Image Corp., 1994, pers. commun.). SPOT images are costly when many scenes are needed (e.g., global glacier monitoring), so the data are not generally used by glaciologists in the study of large ice caps, ice fields and ice sheets.

In 1978, the Seasat satellite was launched by the National Aeronautics and Space Administration (NASA) with an L-band (1.28 GHz) synthetic aperture radar (SAR) on board to carry out studies of the sea surface. The L-band sensor operated for only about 100 days during the summer and autumn of 1978. During its operation, it provided much data on the Earth's glaciers through cloud cover and darkness, although Antarctica was not imaged. In 1991, the first European remote-sensing satellite (ERS-1) was launched with a C-band (5.25 GHz) SAR on board. This sensor, currently in operation, has provided much useful data on the Earth's glaciers (Fig. 6.4) (Hall, in press), including data permitting delineation of faces boundaries on the Greenland ice sheet (Fahnestock *et al.*, 1993) and smaller glaciers. The first Japanese Earth Resources Satellite (JERS-1) was placed in orbit in 1992 and carries an L-band (1.275 GHz) SAR, also useful for glacier studies. Though more difficult to interpret than data from the solar-reflective-energy part of the spectrum, SAR data offer an opportunity to monitor and measure glaciers repetitively, independently of clouds and solar illumination. Subsurface measurements of some glaciological features are detectable on SAR images when the glacier surface is dry (Hall *et al.*, 1995); in some studies, Landsat and SAR data of the same area are used together to improve the analysis (Hall and Ormsby, 1983).

Passive-microwave satellite data have been used to study the Greenland and Antarctic Ice Sheets.

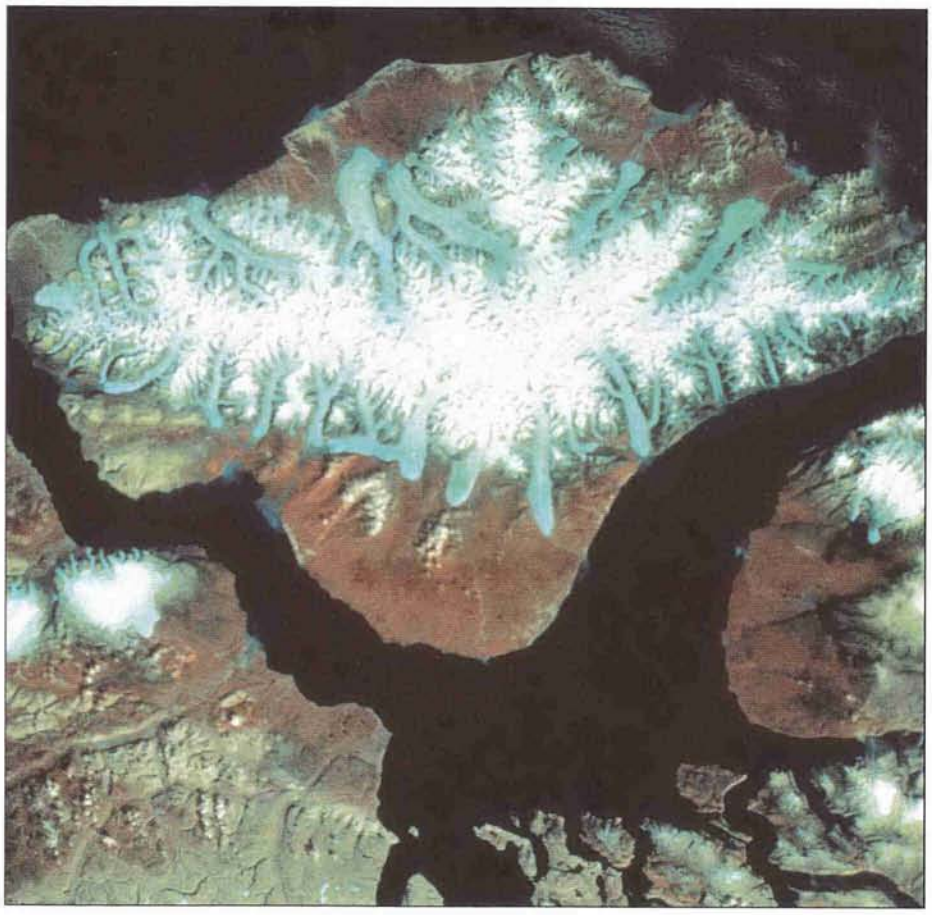

Figure 6.4 Landsat MSS image (20204-16513; 14 August 1975) of Bylot Island, Northwest Territories, Canada, showing cirque and valley glaciers. Two small, unnamed ice caps and outlet glaciers are situated to the west on the Borden Peninsula. Small ice caps and outlet glaciers can also be seen on Baffin Island (after Williams, 1986*b*).

Owing to the poorer resolution of passive-microwave sensors, compared to sensors operating in the reflective and even thermal-infrared parts of the electromagnetic spectrum and compared to radars, the passive-microwave data are most useful for synoptic ice-sheet studies; they are not very useful for studies of small glaciers. Data from the Nimbus-5 electrically scanning microwave radiometer (ESMR), launched in 1972, have been used to estimate the snow-grain size (model dependent) and to relate that to the accumulation rate in Antarctica and Greenland (Zwally, 1977). A spatial resolution of 25 km characterized the ESMR data. Continued work using data from the scanning multichannel microwave radiometer (SMMR) and special sensor microwave/ imager (SSM/I), the latter having additional channels relative to the ESMR and a 15-km pixel resolution channel, has revealed other interesting brightness-temperature patterns that may relate to snow-accumulation patterns (Thomas, 1993).

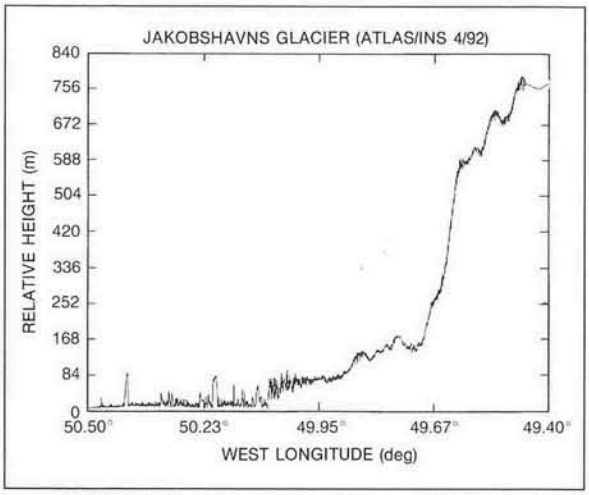

Figure 6.5 Geodetic airborne laser altimeter (GALA) profile of the lower 17 km of the Jakobshavn Isbræ outlet glacier, Western Greenland, on 22 April 1992. Iceberg-filled fjord to the left of the terminus. Note the concentrations of crevasses and surface undulations up-glacier from the terminus (after Garvin and Williams, 1993).

6.3.2 Non-imaging sensors

The surface topography of ice sheets can be measured by radar altimeters (Brooks *et al.*, 1983; Zwally *et al.*, 1983*a*) and laser altimeters. Such systems also offer a means of determining ice thickening or thinning rates over the ice sheets (Thomas *et al.*, 1995). When used in conjunction with data from sensors that can delineate ice-sheet areal extent, net changes in surface-ice volume may be determined on the ice sheets in the future (Williams, 1985). Satellite radar altimeters have operated on-board NASA's GEOS and Seasat satellites and the U.S. Navy and NOAA 'Geosat', as well as NASA's TOPEX spacecraft. Bindschadler *et al.*, (1989) used Seasat radar altimetry to carry out a comprehensive survey of the surface topography of the Greenland Ice Sheet. Radar altimeters transmit short microwave pulses and measure the time delay to receipt of the reflected pulse. Over the polar ice sheets, the altimeter measurements from the GEOS and Seasat altimeters were not vertically precise enough to track small changes in ice-sheet elevation. The Geosat altimeter represented an improvement and acquired more continual records over the ice sheets (Thomas, 1993).

A laser altimeter potentially measures ranges more accurately than does a radar altimeter. Satellite laser altimeters will operate with a very narrow beam, so that the footprint will be only a few tens of meters in diameter (e.g., 30–100 m). The result is an improved capability to measure thickening or thinning of the ice sheets from space, relative to the use of radar altimetry. The smaller footprint size also offers a potential for measuring elevation changes on smaller glaciers using laser altimetry (Garvin, 1993; Garvin and Williams, 1993; James B. Garvin, 1994, oral comm.). Unlike radar altimeters, however, laser altimeters must operate in cloud-free conditions.

Several airborne laser altimeter surveys of the Greenland Ice Sheet have been conducted with Global Positioning System (GPS) receivers on board that provide the accurate navigation required to measure the height of the aircraft above the ellipsoid, identify the ground track and to direct the aircraft as nearly as possible along repeat tracks. Repeated measurements of targets placed at the surface of an ice sheet will provide accurate estimates of ice velocity and strain rate over distances of tens to hundreds of kilometers over short periods (Thomas, 1993).

Geodetic airborne laser altimeter surveys of ice caps and outlet glaciers in Iceland and outlet glaciers in Greenland (Fig. 6.5) have also been carried out (Garvin and Williams, 1993). Orbital laser altimeter surveys of the polar ice caps of Mars are now scheduled for the 1998–2000 time frame using an experimental instrument called the Mars Orbiting Laser Altimeter (MOLA) on NASA's newly approved (to replace the failed Mars Observer mission) Mars Global Surveyor mission (1996 launch).

Radio-echo sounding was discussed in Section 6.2 on 'Achievements made with long-term data'. Radio-echo sounding will continue to be used to obtain information on the thickness of glaciers.

6.4 GAPS AND NEEDS

There is a continuing need for the publication of accurate maps of glaciers of the world. Except for countries like Switzerland, Norway and a few others, where very accurate maps of glaciers have been produced to support a well-developed hydroelectric power-generating system, glacier mapping will continue to be carried out by national mapping agencies as part of their topographic-mapping programme. Over time, larger-scale topographic maps of glaciers will become available; the long-term goal of glaciologists is to have high-quality 1:25,000 scale maps of smaller glaciers and 1:250,000 scale maps of the Greenland and Antarctic Ice Sheets (e.g., geodetically correct map,

accurate delineation of areal extent and topography of all glaciers within the map, correct geographic place-names for all glaciers, etc.). This goal represents a map scale that is ten times better than is generally available for glacierized regions today, with the exception of Antarctica where even accurate 1:1,000,000 scale maps are still lacking for the entire continent. As aerial photographs or satellite images used to produce topographic maps are generally acquired in mid-summer (high solar-elevation angle) and photogrammetrists are not trained to differentiate glaciers from snowpack, acquiring accurate topographic base maps of glaciers in most glacierized regions will continue to be a problem. It would be desirable for a glaciologist to serve as a consultant to national mapping agencies that map glacierized areas, so that more accurate maps of glaciers could result.

Although most topographic mapping is carried out using aerial photogrammetric methods, technology currently exists to produce topographic maps using satellite photogrammetric methods. For the Greenland and Antarctic ice sheets, an interim solution to the lack of adequate maps is to prepare satellite image and image mosaic maps of these areas from NOAA AVHRR, Landsat (Williams *et al.*, 1982), or other satellite image data (Ferrigno and Molnia, 1989). Such maps can be planimetric (U.S. Geological Survey, 1991), they can superimpose topographic contours over the image base to produce a topographic image map.

Accurate topographic maps are needed to support long-term monitoring of glaciers and for producing glacier inventories. Except for Antarctica, 1:250,000 scale maps (DMA Series 1501) have been produced by the (U.S.) Defense Mapping Agency directly or in collaboration with other nations' mapping agencies of many parts of the world, including most of the glacierized regions of Earth. The DMA Series 1501 maps are produced as paper maps and as Level 1 digital terrain elevation data (Level 1 DTED). Except for the United States, distribution of the DMA Series 1501 maps, either in paper or DTED format, is extremely limited. The DMA Series 1501 maps are needed by the scientific community to support a variety of global environmental change programmes, including a world glacier inventory. Release of DMA Series 1501 maps and DTED is on the agenda of the U.S. Government Applications Task Force (GATF).

There is a continuing need to monitor changes in glacier area, position of termini or margin, position of the transient snow-line, absolute or relative volume and other glacier parameters. Medium-resolution sensors (e.g., from Landsat, SPOT, some Soyuzkarta, SAR, etc.) can, in many cases, provide a source of image data to delineate changes in glacier area, in the position of termini and in the position of the transient snow-line. Satellite image data are quite costly for regional and global studies, and their use for glacier monitoring is still severely limited. Cloud cover, lack of systematic, repetitive coverage of the Earth's land areas and lack of stereoscopic data (except for some SPOT) also limit the use of satellite data for glacier studies.

Another gap in monitoring changes in glaciers is that of an accurate measurement of changes in absolute or relative volume. Improved surface and airborne radio-echo sounding techniques can likely achieve better precision in determining changes in absolute glacier thickness, especially with the use of the Global Positioning System (GPS) of satellites for precise navigation. These measurements should be taken every 10 years if they are to give meaningful information. Geodetic airborne laser altimetry (GALA) has proven capable of measuring surface profiles of glaciers in Iceland and Greenland (Garvin and Williams, 1993; Thomas *et al.*, 1995) to sub-metre accuracies. Plans call for geodetic satellite laser altimeter surveys of the entire world, including the glacierized regions, to the highest latitude reached by a polar orbiting satellite (about 82°). Other planned satellite-instrument designs include advanced spaceborne thermal emission and reflection radiometer (ASTER) on an Earth Observing System (EOS) platform (15-m pixel); fore-, aft-, and nadir-pointing sensors on Mapsat (10-m pixel); and future in-line stereoscopic sensors on SPOT 5 (5-m pixel) (Jonathan F. Williams, SPOT Image Corp., 1994, written comm.); these instruments will provide capabilities to acquire higher resolution stereoscopic satellite imagery on a consistent basis. These data will provide national mapping agencies with the source material to produce topographic maps of unmapped or poorly mapped regions, if precise satellite ephemeris can be achieved, and will also provide glaciologists involved in monitoring changes in glaciers and compiling glacier inventories with important new source material.

The World Glacier Inventory (WGI) project has as its long-term goal the compilation of a complete inventory of all of the Earth's glaciers, small and large. The availability of adequate maps, aerial photographs and regional glaciologists has meant that most of the emphasis of the WGI project since its inception in 1976 has been on inventorying relatively small glaciers in Europe, Asia and North America. Eventually, the WGI project will have to tackle an inventory of larger glaciers, including the two remaining ice sheets in Greenland and Antarctica (Swithinbank, 1983), and ice caps, ice fields and associated outlet glaciers, ice streams and ice shelves.

Long-term, systematic monitoring of changes in the area, termini and mass balance of small glaciers and the preparation of glacier inventories, the fundamental mission of the World Glacier Monitoring Service (WGMS), provides an excellent way of measuring regional changes in temperature and precipitation. Mass balance measurements of glaciers must still be taken in the field at present (Østrem and Brugman, 1991). Mass balance measurements of glaciers are limited to small glaciers because they require costly logistics and are very labour-intensive; except for a few glaciers of academic interest, mass balance measurements are done to support hydro-

logical information needs of hydroelectric-power-generating facilities. As a result, the larger glaciers (ice sheets, ice caps and ice fields) are the least known. It is not even known whether the mass balance of the Greenland and Antarctic Ice Sheets is positive or negative (National Research Council, 1985)! Changes in the large ice masses can have a profound effect globally because of the effect of changes in sea level. Meier (1984, 1985) suggested that small glaciers have contributed at least one-third of the 25-cm rise in sea level measured during the past century; if 25 cm is accurate, then at least 33,300 km^3 of glacier ice has melted from either non-ice-sheet glaciers (13.9 %), the Greenland Ice Sheet (1.3 %), the Antarctic Ice Sheet (0.1 %), or some combination of all three.

6.5 SUGGESTED FUTURE DEVELOPMENT OF MONITORING ACTIVITY

Mapping, monitoring and inventorying the Earth's glaciers in the 21st century will continue to depend on data from remote sensors, both imaging and non-imaging, carried in aircraft and spacecraft, with increasing emphasis on the latter platform because of the global perspective and because of improvements in sensor capabilities (Swithinbank, 1983). In small glacierized regions, vertical aerial photography and, in the future, high-resolution satellite imagery, will continue to be the primary data used in preparing maps, monitoring changes in glacier area and position of termini and preparing comprehensive glacier inventories. For larger glaciers, imaging and non-imaging sensors will be used for mapping, monitoring changes and preparing preliminary glacier inventories (IAHS(ICSI)/UNEP/UNESCO, 1983). Richard S. Williams Jr. and Jane G. Ferrigno of the U.S. Geological Survey are leading an international project team to prepare 24 1:1,000,000 scale maps of the coastal regions of Antarctica that will show the changing position of the dynamic coast and define it glaciologically; the team will plot the velocity vectors of outlet glaciers, ice streams, and ice shelves and compile a preliminary inventory of all glaciers and ice streams previously mapped or mappable on Landsat multispectral scanner (MSS), return beam vidicon (RBV) and thematic mapper (TM) images (Fig. 6.6a/b) (Williams *et al.*, 1995).

Non-imaging sensors, especially laser altimeters, will provide, with Global Positioning System (GPS) control, accurate (sub-metre) profiles of glacier surfaces on a global basis. Not only will the topography of the Greenland and Antarctic Ice Sheets be measured consistently and accurately for the first time but changes in surface elevation over time will be determined. When combined with information from imaging sensors about areal changes, changes in surface elevation will permit relative changes in volume to be determined. Seasonal and inter-annual measurements will allow snow cover depth variations to be calculated. Measurements of this kind taken over a period of several decades will permit mass balance changes to be calculated for large glaciers for the first time.

Some of the satellites and remote-sensing instruments planned for the near future that have applications to glacier monitoring are described in the next few paragraphs. The Landsat 7 spacecraft, scheduled to be launched in 1998 as part of the EOS, will carry an enhanced thematic mapper (ETM) sensor that will operate in the visible, near-infrared, short-wave infrared and thermal-infrared bands of the electromagnetic spectrum; it will have a repeat cycle of 16 days and a pixel resolution of 15 m for the panchromatic band, 30 m for the four visible and near-infrared bands and for the two short-wave infrared bands, and 60 m for the thermal-infrared band. The ETM will enable researchers to acquire data about glaciers and ice sheets so that a multidecade database will be established, a key to deriving useful global environmental change information. As at October 1995, the ERS-2 satellite was expected to be launched by the end of the year. It would have essentially the same payload as was currently available on ERS-1, described previously. A JERS-2 satellite was also planned. Radarsat, a Canadian satellite, was to be launched in the autumn of 1995; it would provide C-band (5.3 GHz) SAR data on most of the Earth, including most of the glacierized regions, at pixel resolutions between 10 and 100 m.

The moderate-resolution imaging spectroradiometer (MODIS) will be launched as part of the Earth Observing System (EOS) payload in 1998. MODIS will acquire data in 36 spectral bands from 0.45 to 14.4 μm. Spatial resolution will range from 250 m to 1 km and will be suitable for studying some of the larger glaciers, including ice fields, ice caps and ice sheets. Additionally, the MODIS sensor will not saturate in the visible bands, allowing reflectance variability on the ice caps and ice sheets, which may relate to facies boundaries, to be discernible. The multifrequency imaging microwave radiometer (MIMR) is a European Space Agency (ESA) instrument to be flown on-board the Earth Observing System (EOS)-PM (afternoon equatorial crossing) satellite to be launched in the year 2000. Over the ice sheets, passive-microwave data will allow studies of glacier facies and the snow-accumulation rate, extending the studies begun using ESMR, SMMR and SSM/I data but at an improved (up to 10-km) pixel resolution. Global coverage of the Earth will be achieved every 3 days, with daily coverage at high (>45°) latitudes (Asrar and Dokken, 1993).

The geodynamics laser altimeter system (GLAS), with a planned launch on an EOS polar-orbiting platform in about the year 2002, will provide high-resolution altimetric profiles of the ice sheets with a vertical accuracy of 5–10 cm. The altimeter mode will provide measurements of average surface elevation within a footprint about 70 m in diameter. The primary application of these data will be to measure

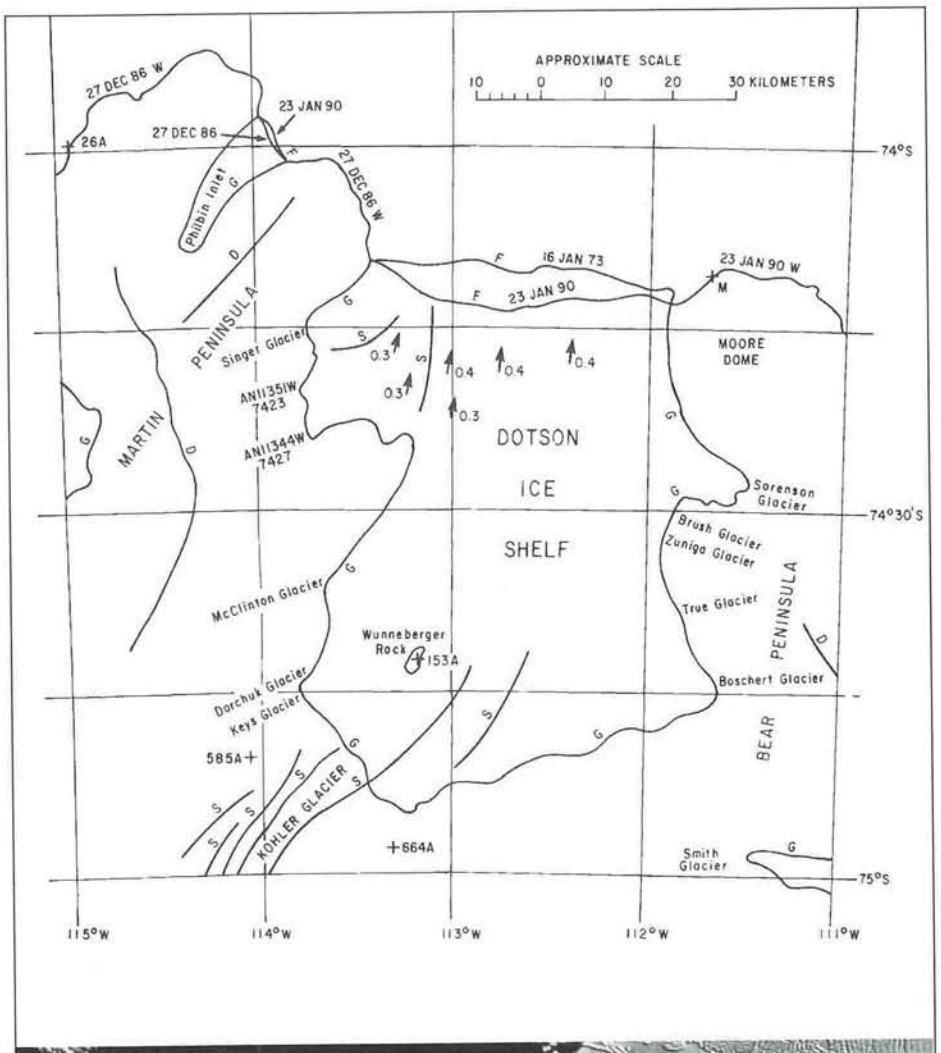

Figure 6.6 (top) - Map showing the floating- and grounded- ice margin, preliminary glacier inventory, velocity vectors, delineation of the grounding line, and glacier flow lines of part of the Bakutis Coast, West Antarctica, derived from the analysis of three Landsat images, 16 January 1973 MSS image (1177-14500), a 27 December 1986 TM image (51031-15133), and 2 January 1988 TM image (41996-14580). On this map the letters indicate: W, ice wall; D, ridge line on ice; S, flow lines; G, grounding line; and F, ice front (after Williams *et al.*, 1995).
(bottom) - A Landsat image (1160-14554; 30 December 1972) of the Dotson Ice Shelf and environs used in the glaciological analysis of coastline changes in Bakutis Coast, West Antarctica.

rates of ice sheet thickening or thinning and, ultimately, to calculate changes in mass balance.

As the Landsat 4 and 5 MSS and TM data become more reasonable in cost and, in the future, when Landsat 7 data become available, it will be possible to do comparative studies to show areal and termini changes in the Earth's glaciers with reference to the global baseline study of glaciers carried out by Williams and Ferrigno (1994) on the 'Satellite Image Atlas of Glaciers of the World' project, using Landsat 1, 2, and 3 MSS and return-beam vidicon (RBV) data from the 1970s and '80s. It will be possible to monitor decadal-scale changes in glacier-terminus position and firnline elevation on a regional and even global scale. Other data sets, for example the ERS-1 and -2 and JERS-1 and -2 data, will augment and complement the data available from the Landsat series of sensors. Using the ASTER sensor on EOS, it will also be possible to monitor glaciers on a regular basis, with a 15 m resolution, and to acquire stereoscopic coverage.

If existing and future remotely-sensed data sets are to be useful for mapping, monitoring and inventorying glaciers, the World Glacier Monitoring Service (WGMS) must provide leadership about which data sets are important for reaching WGMS goals and objectives, how the data can be made more accessible to the glacier-monitoring community, what standards should be established to achieve consistency in analyses of remotely-sensed data and format for reporting to WGMS, how information derived from the data sets can best be incorporated into WGMS archives and in what form the information should be summarized and distributed to the glaciological community and decision-makers.

Monitoring global change in the cryosphere is a growing concern to the international global environmental change research programme (International Geosphere-Biosphere Programme and Intergovernmental Panel on Climate Change (IPCC) (Fitzharris, in press)) and monitoring variations in glaciers is especially important for accurately determining long-term changes in regional and global climate and in determining which glacierized areas are contributing to changes in sea level (Warrick and Oerlemans, 1990). What began, more than 100 years ago, as a basic scientific programme, is now of growing international scientific importance.

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7 Glaciers in North America

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7.1 INTRODUCTION

The North American land mass, comprising Mexico, the United States, and Canada, spans latitudes from 15° to 83° north, includes mountains exceeding 6,000 m above sea level, encompasses climatic environments ranging from desert to temperate-wet maritime to dry continental to polar desert and includes glaciers of all kinds other than ice sheets (Meier, 1990). The presently glacierized area in North America is estimated at some 275,000 km^2 with about 200,000 km^2 in Canada, the country with the largest concentration of perennial surface ice outside the polar ice sheets, roughly 75,000 km^2 in the USA (mainly in Alaska) and a few small glaciers on the Mexican volcanoes. The following text concentrates on glaciers in Canada, with a brief discussion of glaciers in the United States.

7.2 DISTRIBUTION AND CHARACTERISTICS

The Canadian land mass extends from the 53rd to the 141st meridians west and from latitude 42°N, an area of almost 10 million km^2, with mountains rising up to 6,000 m above sea level. It contains a variety of environments suitable for sustaining glaciers. In his study of mountain glaciers, Field (1975) and his collaborators described their distribution and reported on much of the work done on them. Small glaciers exist on both continental margins, in the Torngats of Labrador and in the central mountains of Vancouver Island. Larger glaciers are found in the Rockies, the Interior Ranges and the Coast Mountains. They get progressively larger towards the Panhandle, the boundary between British Columbia and Alaska, through the Juneau Ice Field region and up to the Ice Field Ranges of the Saint Elias Mountains, where large glacier systems, such as the Seward Glacier, are found. Some smaller outliers are located in the eastern Yukon and western District of Mackenzie.

The mean height of the snow-line in the west rises steadily across the Cordillera, from 1,700 m in the Coast Mountains to over 2,700 m in the Rockies, reflecting a continentality effect. Northwards, the

glaciers increase in size and reach to lower elevations, demonstrating a latitudinal effect due to the reduction of mean annual temperature. This is best seen in the eastern Arctic, where the mean height of the snow-line drops from 700 m on Baffin Island to sea level on the Ward Hunt Ice Shelf. In the eastern Arctic, glaciers are distributed along the mountain and fiord coast of Baffin Island, with two larger concentrations in the Barnes and Penny Ice Caps; the concentrations become larger and more common further north. Axel Heiberg, Ellesmere and Devon Islands are all covered by ice sheets several thousand km^2 in area. Generally, the ELA in the Queen Elizabeth Islands lies at just over 1,000 m a.s.l.. Andrews and Miller (1972), Miller *et al.* (1975), and Østrem (1966*a*; 1972; 1973), have reported on the regional characteristics of glaciation levels, snowlines and equilibrium lines throughout Canada.

In the United States, glaciers occur in Alaska, Washington, Wyoming, Oregon, Montana, California, Colorado, Idaho and Nevada. In Alaska and adjacent Yukon/British Columbia, vast intermountain ice fields occur in the ranges bordering the Gulf of Alaska (Coast, Saint Elias, Wrangell, Chugach, and Kenai Mountains), where precipitation is heavy (locally exceeding 8,000 mm per year), equilibrium line altitudes are low (to less than 600 m above sea level), and individual glaciers are large (Bering Glacier is about 200 km in length and 5,800 km^2 in area). Large valley glaciers also occur in the Alaska Range; smaller glaciers occur on the volcanic peaks of the Aleutian Range as well as in the Kilbuck/Ahklun and Talkeetna Mountains, and the Brooks Range and Seward Penninsula in the Arctic. Valley glaciers also occur on the volcanoes of the Cascade Range in Washington, Oregon, and California, and in the Olympic Mountains of Washington and the Wind River Mountains of Wyoming. Small cirque glaciers occur in the Sierra Nevada, the Basin-and-Range province and on the many ranges that make up the Rocky Mountains of the United States. These glacier settings are described in Field (1975), Meier *et al.* (1971) and Meier (1990).

7.3 EXISTING INVENTORIES

Originally proposed for the International Geophysical Year (IGY) in 1958, the suggestion that an inventory be made of the basic characteristics of all the world's glaciers was adopted as part of the International Hydrological Decade (1965–1974) and the inventory was carried out by a number of countries. In Canada, the IGY glacier inventory was initiated at the University of Toronto and focused on the Arctic Islands. Glacier tongues were identified and numbered. When the principal investigator moved to the Geographical Branch of the Department of Mines and Technical Surveys, the emphasis changed to snout behaviour and variations that could be identified through historical records (Falconer, 1962) but the work was wound up when the Branch was disbanded. However, Canada subsequently played a leading role in developing the initial recommendations for the IHD inventory, by undertaking pilot studies and completing the first partial inventory (Ommanney, 1969) at McGill University under the direction of Fritz Müller, the Chairman of the ICSI working group.

In 1968, the Glaciology Subdivision of the Department of Energy, Mines and Resources adopted the programme and established a group to implement it (Ommanney, 1980). Limited resources and the vast territory to be covered meant progress was slow. Work in the early years built on that done for the IGY, concentrating on the Arctic Islands. Glaciers were identified and mapped but not measured because of limitations in the 1:250,000 scale base maps revealed through analysis of the Axel Heiberg Island inventory (Ommanney, 1969). Maps showing inventory numbers were published in the Glacier Atlas of Canada series (Ommanney, 1989). Pilot studies that included the full inventory data set were completed for Coburg Island (Whytock, 1974), d'Iberville Fiord (Wilkins, 1973) and the areas contributing to the North Water (Kraus, 1983), all on or around Ellesmere Island.

A major effort was made to complete an inventory of glaciers in the Yukon Territory. Again, limitations in the base maps available meant that a full data set was not compiled. The data were sufficient to be used in a study of likely glaciological impacts on a proposed pipeline between Kluane Lake and the St. Elias Mountains (Canada, 1977) and in an analysis of surging glaciers (Clarke, 1991; Clarke *et al.*, 1986). The Yukon inventory was picked up again in the 1990s as a contribution to a Canadian programme looking at global change through studies of the cryosphere. Efforts were well under way within the National Hydrology Research Institute (NHRI) to transfer much of this data into a geographical information system when the principal investigator left and was not replaced. However, the basic data have been compiled and summaries presented (Ommanney, 1993).

In an earlier effort to look at the limits of glacierization in Canada and their possible sensitivity to change, an inventory was completed of the glaciers on Vancouver Island (Ommanney, 1972*a*).

Interest in glaciers as a water resource in western Canada and a proposal to develop the hydroelectric resources of the Stikine-Iskut area led to the compilation of data on the thousands of glaciers in that area. These conform to the international inventory standard but also include information on the area of ice within 100-m elevation bands. Unfortunately, declining resources led to the suspension of the Glacier Atlas mapping programme and the data were never published.

Inventory maps were also compiled for substantial areas of the Rockies, though no data were tabulated, and for Glacier National Park in British Columbia. For this latter area, inventories were com-

pleted for three different time periods to assess changes that had taken place in the ice cover over the last century (Champoux and Ommanney, 1986).

In the United States, detailed inventories have been produced for the North Cascades (Post *et al.*, 1971) and Olympic Mountains in Washington, the Sierra Nevada in California (Raub *et al.*, 1980) and the Brooks Range in Alaska. An estimate of the glacier-covered area, with statistics on many of the larger glaciers, has been compiled for Alaska (Post and Meier, 1980).

Although much of the data were deposited with the World Glacier Monitoring Service, more could be made available if resources were devoted to finishing off some of the more advanced work.

7.4 LONG-TERM OBSERVATIONS (LENGTH CHANGES, MASS BALANCES, MAPS)

It was the opening up of the Canadian west by the Canadian Pacific Railway (CPR) in the late 1800s and the linking of British Columbia with the rest of Canada that gave glaciologists access to the Rockies and Interior Mountains and created an opportunity for the first systematic glacier observations. The Illecillewaet Glacier and neighbouring Asulkan Glacier became the object of the first glaciological investigations. The early activities have been reviewed by Cavell (1983). Other studies of note were conducted on the Victoria and Yoho Glaciers. A. O. Wheeler, the founder of the Alpine Club of Canada, made sure that scientific observation of glaciers was included in its mandate.

The First World War created a hiatus in the recording of glacier variations that continued until after the Second World War.

Compared to more highly-populated mountain areas, such as the European Alps, the record of Canadian glacier variations is quite sparse. However, at least for the four glaciers discussed above, there are fairly complete records of recession during a period of major retreat.

The immediate postwar period saw a significant increase in the number of glaciological studies, mostly due to the initiation of an annual survey of specific glaciers in the Cordillera by the Dominion Water and Power Bureau, the forerunner of the Water Survey of Canada. This was part of a study of the water resources of mountainous rivers. In 1945, seven glaciers in Alberta – Angel, Athabasca, Freshfield, Peyto, Saskatchewan, Southeast Lyell and Victoria – were investigated by the Calgary office and eight glaciers in British Columbia by the Vancouver office – Bugaboo, Franklin, Helm, Illecillewaet, Kokanee, Nadahini, Sentinel and Sphinx. The position of the snout and changes in its areal extent were measured and a set of plaques were placed on the ice surface to measure velocity. Summaries of the reports for Peyto Glacier and Victoria Glacier have been published by Ommanney (1971; 1972*b*).

Canadian participation in the IGY led to a University of Toronto Expedition to study the Salmon Glacier. On Ellesmere Island, the Defense Research Board (DRB) started a programme on the Gilman Glacier and continued studies on the Ward Hunt Ice Shelf (Hattersley-Smith, 1954).

In the period immediately following the IGY, McGill University launched the Jacobsen-McGill Arctic Research Expedition to Axel Heiberg Island under the direction of F. Müller, initiating studies of the White and Baby Glaciers that have continued intermittently to the present (Cogley *et al.*, 1995). The Arctic Institute of North America (AINA) mounted an expedition to Devon Island, and DRB continued and expanded its studies on Ellesmere Island. These activities combined to raise the level of glaciological research in Canada. W. S. B. Paterson joined the Polar Continental Shelf Project and started working on the Meighen Ice Cap (Paterson, 1969); building on the programme started earlier by K. C. Arnold (Arnold, 1965). By the mid-1960s, his programme on that ice cap had been expanded to include the Melville Island Ice Caps and the Devon Ice Cap (Paterson, 1976), taking over from AINA whose programme involved mass balance and meteorological studies on the Devon Ice Cap and some detailed investigations of the Sverdrup Glacier (Koerner, 1970). DRB's Operation Hazen was a large multi-disciplinary investigation, similar to one under way on the neighbouring Axel Heiberg Island. The glaciological part of the programme was concentrated on Gilman Glacier, the Ward Hunt Ice Shelf and Ward Hunt Ice Rise. It resulted in reports on glacier surveying, mass balance, temperatures and radio-echo sounding (Ommanney, 1982).

The Geographical Branch was continuing the work begun by the Baird expedition on the Barnes Ice Cap (Løken and Sagar, 1968). Included were studies of the Barnes Ice Cap itself and the small Lewis Glacier at its northern margin. Some observations were also made on the Penny Ice Cap.

In 1961, the American Geographical Society (AGS), in conjunction with AINA, established the Icefield Ranges Research Project. This was similar in scope and intent to the McGill and DRB expeditions. It included detailed glaciological and climatological studies, particularly on the Kaskawulsh Glacier and around Mount Logan. It was partly an outgrowth of the earlier Operation Snow Cornice. The results of the scientific investigations were published in four volumes by the AGS (Bushnell and Marcus, 1974, Bushnell and Ragle, 1969; 1970; 1972).

The most important stimulus was provided by the IHD (1965–1974, also referred to as 'the Decade'), which led to a major expansion of glaciological investigations. These developments have been reviewed by Ommanney (1975). In the Cordillera, five glaciers – Place, Sentinel, Woolsey, Peyto and Ram River – were selected for an east/west transect of the Cordillera and Berendon Glacier was added to provide a link in the north-south chain. The programme was run by

the Glacier Section of the Geographical Branch, Department of Mines and Technical Surveys, the forerunner of the Snow and Ice Division of the Department of the Environment and NHRI. It followed a set of standardized measurements for mass balance outlined by Østrem and Stanley (1966).

Decade Glacier on Baffin Island was selected for the north-south chain in the eastern Arctic, which included the DRB studies on Per Ardua Glacier and the McGill studies on the White and Baby Glaciers (Young, 1972). The effective network was much larger than the official 'representative glacier basins', as existing research investigations continued or were expanded to include a larger hydrological component.

On the mainland, new mass balance studies began on Rusty, Cathedral and Drummond Glaciers. Many did not continue throughout the Decade and of the representative studies, those on Woolsey, Ram River, Berendon, Decade and Per Ardua were terminated during or at the end of the Decade. However, the availability of semi-permanent facilities at most of these glaciers and core staff to maintain a measurement programme throughout the summer melt season led to the development of many other complementary glaciological investigations.

In the Coast Mountains, continuous records have been maintained on Sentinel and Place Glaciers, which are now being used as benchmarks for comparison with shorter-term mass balance investigations in other parts of the range. Studies relate to the operational needs of the water management agencies and often only run for a few years, such as those on the Bridge River glaciers, in the Stikine and Iskut river basins, and in the Homathko basin.

In the mid-1960s, following the Glacier Mapping Symposium held in Ottawa and recommendations from the National Research Council of Canada's Subcommittee on Glaciers, the Water Survey of Canada (WSC) switched to a programme of terrestrial photogrammetry that involved mapping the ablation areas of the glaciers every two years (Reid and Charbonneau, 1979*a*; 1979*b*). Snout and plaque surveys of the Athabasca and Saskatchewan Glaciers were continued in the intermediate years by the Calgary office of the WSC (Canada, 1982).

An initiative by K. E. Ricker, a private consultant, in conjunction with W. A. Tupper of the B.C. Institute of Technology, has added to our knowledge of recent glacier variations in the Coast Mountains. Their studies extend from the St. Elias Range through the Hazelton Mountains, the Pacific, Chilcotin, Elaho, Clendenning and Lillooet Ranges to Garibaldi Provincial Park (Ricker, 1976; 1980; Ricker and Tupper, 1988; Ricker *et al.*, 1983).

Today, there are no groups in Canada charged with routine monitoring of glacier variations. Some measurements of snout variations are being carried out in conjunction with the mass balance programmes described below.

Length- and thickness-change observation programmes have a long history in the United States but only a small amount of systematic observation is being done at present. Perhaps the most complete record is from Nisqually Glacier, Mount Rainier, Washington, where observations began in 1840. Starting in 1931, the front of the glacier and thickness changes at several profiles across the tongue have been measured annually and a topographic map of the tongue made every five years. These measurements were made because of the impact of glacier recession on hydropower production. Although thinning seemed to be pervasive for many years, thickening was detected in 1946 (Johnson, 1960) as a result of a period of climatic cooling/higher precipitation which was followed by major advances by glaciers in the Pacific Northwest in the late 1950s and 1960s. Many other programmes of terminus retreat or thinning observations were made, and continue to be made, on glaciers in the conterminus United States but no centralized authority is involved in either data collection or dissemination. In Alaska, recession/ advance measurements on several major tidewater glaciers (e.g., Taku, Muir, Hubbard, Columbia) have showed changes, both plus and minus, amounting to tens of kilometres.

7.4.1 Glacier Mass Balance

The small mass balance programme now being run in western Canada is largely a remnant of the network established in the mid-1960s for the International Hydrological Decade (Østrem, 1966*b*). Annual measurements of summer, winter and net balance are taken on Peyto Glacier in the Rockies (Fig. 7.1a/b), and on Place (Fig. 7.2a/b), Sentinel and Helm Glaciers in the Coast Mountains. The techniques used for assessing mass balance have been reported by Østrem and Brugman (1991) and the overall Canadian programme reviewed by Ommanney (1992). Results are compiled and submitted to the World Glacier Monitoring Service (e.g., Ommanney, 1991) for publication in the IAHS(ICSI)/UNEP/ UNESCO *Fluctuations of Glaciers* series and the biennial *Glacier Mass Balance Bulletin*. The interest of BC Hydro, a major provincial energy company, led to support for an additional set of surveys in the Coast Mountains (Mokievsky-Zubok, 1991*a*; 1991*b*). The mass balance studies are frequently supplemented by other research, such as that into the modelling of surface meltwater discharge (Munro 1991) and volumetric change (Glenday, 1991).

The Geological Survey of Canada's glacier programme in the Canadian Arctic is also a long-standing one, previously the responsibility of the Polar Continental Shelf Project. Surveys have been carried out on ice caps on Ellesmere, Devon, Meighen and Melville Islands (Koerner, 1979; 1985; 1986; Koerner and Paterson, 1974). Recent work has concentrated on the Agassiz Ice Cap (Fisher and Koerner, 1994; Koerner and Fisher, 1990).

After a slight hiatus, the glaciological programme on Axel Heiberg Island has been revived, under the

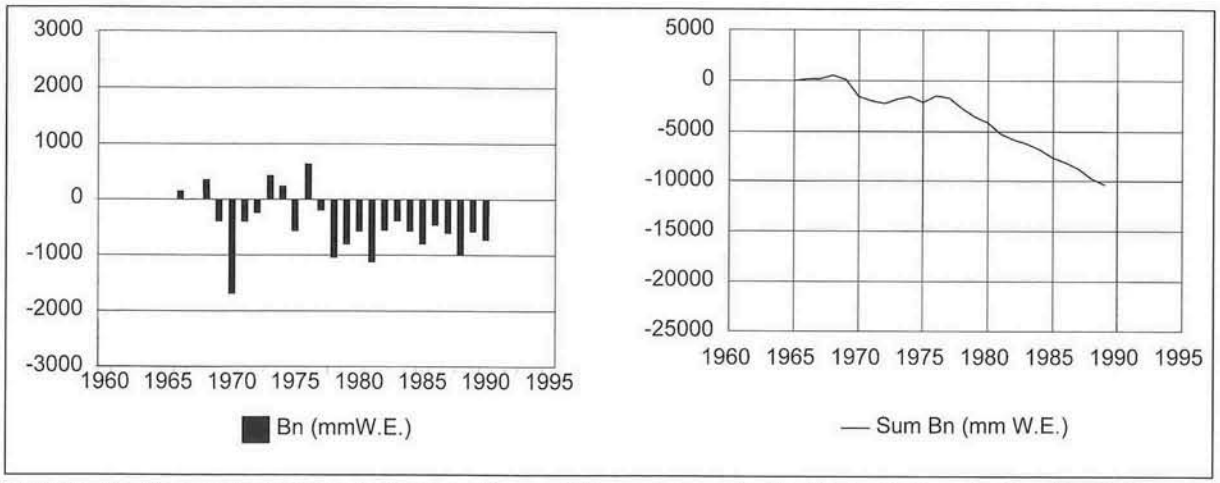

Figure 7.1 Specific annual net balance (left) and specific cumulative net balance (right) of Peyto Glacier in the Canadian Rockies.

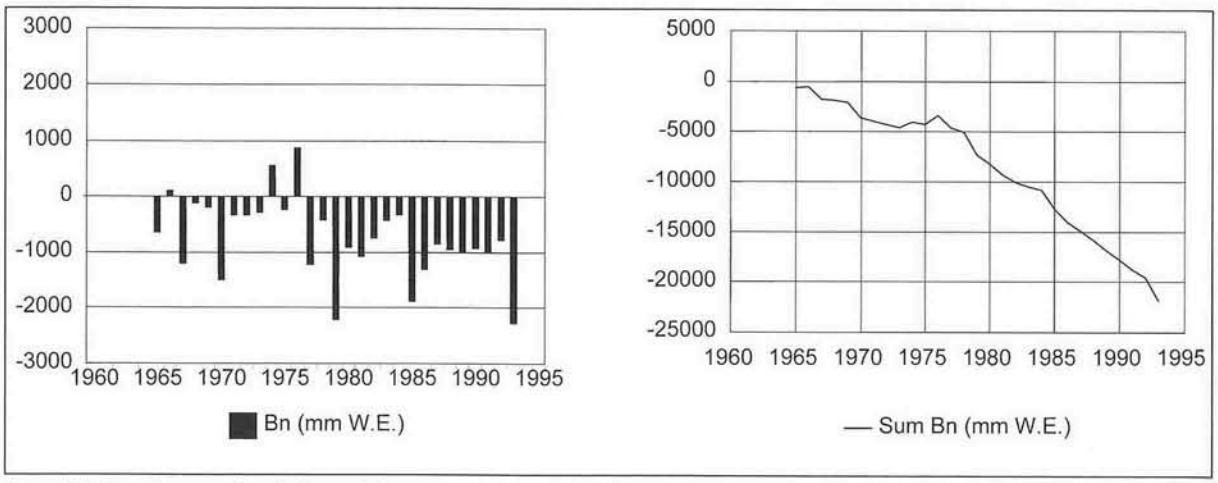

Figure 7.2 Specific annual net balance (left) and specific cumulative net balance (right) of Place Glacier in the Coast Mountain Range of Western Canada.

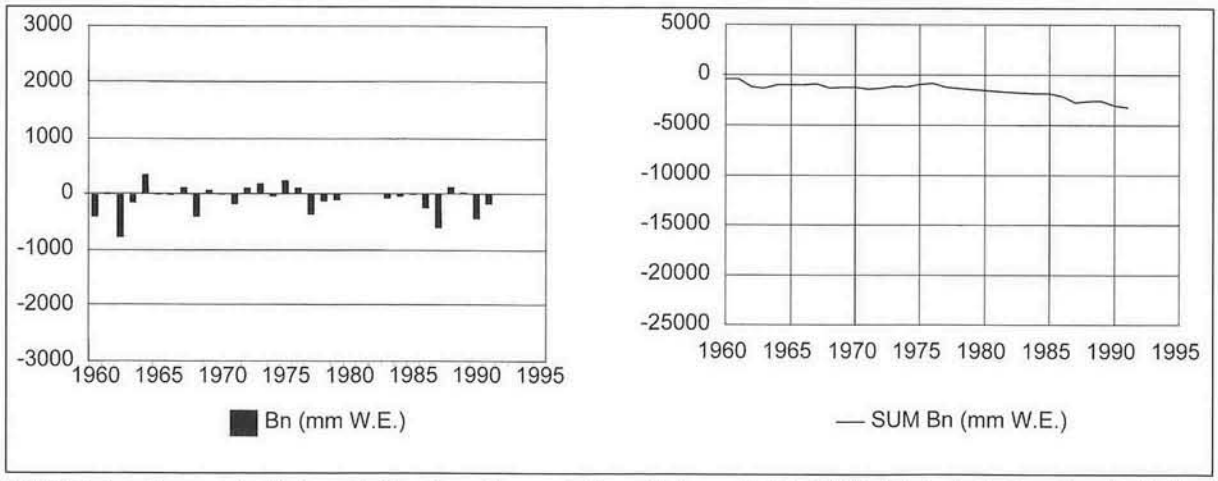

Figure 7.3 Specific annual net balance (left) and specific cumulative net balance (right) of White Glacier in the Canadian Arctic (Axel Heiberg Island). A gap exists between 1980 and 1982. The missing period was filled with the long-term average of –93 mm W.E.

auspices of McGill and Trent Universities. Surveys are once again being made on the White (Fig. 7.3) and Baby Glaciers and results have been published (Adams and Ecclestone, 1991; Cogley *et al.*, 1995; Tolland *et al.*, 1991). There is even renewed interest in the Barnes Ice Cap that has resulted in new surveys of its margin and the installation of weather stations (Jacobs and Heron, 1990).

Intensive mass balance programmes began in the United States in the late 1950s, with International Geophysical Year programmes on the Blue and South Cascade Glaciers in Washington, followed shortly thereafter by a programme on McCall Glacier in northern Alaska. The McCall Glacier programme faltered after the IGY but has been revived several times and is now active. With the beginning of the

International Hydrological Decade, intensive programmes were instituted on the MacClure Glacier, California, and the Gulkana and Wolverine Glaciers in Alaska. The mass balance programmes on the South Cascade, Gulkana, and Wolverine Glacier continue to this day, with a full suite of winter balance, summer balance, hydrologic balance and climate observations, although persistent problems with icing caused the Gulkana streamflow gaging station to be discontinued. The Blue Glacier programme, which reported a net balance only, has been discontinued, as has the MacClure Glacier study. Occasional mass balance determinations have been made on a number of other glaciers, as described in Meier *et al.* (1971) and in reports of the World Glacier Monitoring Service.

7.4.2 Glacier mapping

Small-scale maps on a scale of 1:1,000,000, suitable for planning glaciological research and with accurate information on glacier distribution, were compiled by the Geographical Branch. Based on the best available maps in 1965 (Falconer *et al.*, 1966), they have been used for plotting the distribution of glaciation levels, equilibrium lines, ice-cored moraines, transient snow-lines and glacier mass balance (Østrem, 1973), and as base maps for the National Atlas.

A comprehensive review of the programmes having produced a variety of Canadian glacier maps was made by Ommanney (1986). There have been three principal agencies involved: the Water Survey of Canada, with biennial terrestrial photogrammetric maps of the Athabasca and Saskatchewan Glaciers in Alberta and the Bugaboo, Helm, Kokanee, Nadahini, Sentinel and Sphinx Glaciers in British Columbia; a programme terminated in the late 1970s; the Photogrammetric Section of the National Research Council of Canada, responsible for mapping Salmon Glacier in British Columbia and the White, Baby, Thompson and Crusoe Glaciers on Axel Heiberg Island; and the National Hydrology Research Institute and its predecessors, whose large-scale maps, produced in support of mass balance studies, included ones of the Peyto, Woolsey, Ram River, Sentinel, Decade and Berendon Glaciers, as well as a 12-colour map of the Columbia Ice Field. The Topographic Survey of Canada, in conjunction with the Polar Continental Shelf Project, published an excellent map of the Meighen Ice Cap in the early 1960s.

In the conterminous United States, the Blue, South Cascade, Nisqually, Eliot, Collier, Palisade, Grinnell, Sperry and Dinwoody Glaciers have had at least two special topographic mappings to detect volume changes. Lemon Creek, Worthington, Columbia, West Gulkana, McCall and several other glaciers in Alaska have been subjected to remapping programmes to document volume change.

Remote-sensing techniques being developed at the National Hydrology Research Institute and Canadian Center for Remote Sensing will improve the all-weather mapping of ice extent, transient snow-lines and surface facies (Adam *et al.*, 1995; Adam, 1996). This will assist ongoing efforts to bring the state of glacier mapping into the realm of the Geographic Information System and enable more efficient coupling to other data bases, such as mass balance and glacier fluctuations, for both site-specific and regional studies.

7.5 SPECIAL EVENTS

Glacier surges are noteworthy in the United States and Canada. Work by Austin Post on glaciers in the Alaska Range (Post, 1960) led to the first complete description of this interesting phenomenon, work which was later extended to other areas (Post, 1969, 1972; Meier and Post, 1969). Major research programmes have been conducted on the Variegated and Black Rapids Glaciers in Alaska. A fortuitous observation of the surge of the Steele Glacier led to studies of its cause and an influential symposium (Ambrose, 1969). Since then, Clarke has focused much of his research effort on the elucidation of this problem, particularly through studies of Trapridge Glacier (Clarke, 1976; Clarke and Blake, 1991; Clarke *et al.*, 1984). Other surging glaciers observed in Canada are the Backe, Rusty, Donjek, Tweedsmuir, Lowell, Walsh, Otto and Good Friday Bay Glaciers and the Barnes Ice Cap.

Catastrophic outburst floods are quite common in Canada but fortunately occur mainly in unpopulated areas. Notably, however, the Cathedral Glacier drainage (Jackson, 1979) continues to be a concern to highway and railway officials. Most of the principal sites in the northwest and the Yukon have been identified (Post and Mayo, 1971; Canada, 1977). Detailed reports have been written on several events in the Yukon and British Columbia (Clague and Mathews, 1973; Clarke; 1980; 1982; 1986; Clarke and Mathews, 1981; Desloges and Church, 1992; Desloges *et al.*, 1989; Evans, 1986; Jackson, 1979; Mathews, 1965; 1973; Mokievsky-Zubok, 1980; Young 1977), as well as on Ellesmere Island (McCann and Cogley, 1977). Glacier outburst floods are also common in Alaska (Post and Mayo, 1971) and have caused problems with highway and railway corridors, as well as other human works. Some of these floods followed the damming of a river valley by a glacier, as in the famous Lake George and Blockade Lake situations. Glacier-volcano interactions are important in Washington and Alaska. The eruption of Mount Saint Helens, Washington, caused the beheading of its glaciers; the resulting ice melt contributed to the disastrous debris flows that coursed down into populated areas. Other volcanos in the Cascades show a history of similar debris flows and are being watched carefully. Similar problems have recently occurred on Mounts Spurr and Redoubt in Alaska, and many other Alaskan volcanoes have the potential to produce floods and debris floods triggered at least in part by ice melt.

Tidewater glaciers that calve icebergs into the sea are a special feature of the southern Alaskan coast. These glaciers show the largest historical retreats observed on Earth. Nevertheless, a typical fluctuation cycle also includes a longer period of slow advance (Meier and Post, 1987), which has been well documented in Alaska. As is the case with glacier surges, these fluctuations have little, if any, relation to climatic change. A particularly thorough study of a tidewater glacier, which switched from a stable mode to rapid irreversible retreat in 1982, is described in the series of reports on Columbia Glacier published by the U. S. Geological Survey as Professional Papers 1258 A-H.

7.6 GAPS AND NEEDS

At a workshop held in 1990, the North American Committee on Climate and Glaciers (1991) concluded that the Canadian network was inadequate from the point of view of: regional representation as part of global glacier mass balance; representation of end members; assessment of continental water resources; and addressing global-change issues, such as rising sea level. At present, there is no agency charged with routine observations of either glacier mass balance or glacier variations. Observations being made are in support of programmes for which data collection itself is not a sufficient justification.

The east/west transect in southern British Columbia, with the possible periodic inclusion of the Ram River and Woolsey Glaciers, is probably adequate for addressing regional water resource concerns in that area but large gaps exist in north and central British Columbia, the Panhandle of Alaska and the Rockies. Specific areas needing attention are the Yukon, Nahanni, Juneau Ice Field, central British Columbia, Labrador and Baffin Island.

7.7 SUGGESTED FUTURE DEVELOPMENT OF MONITORING ACTIVITY

It is unrealistic to suggest that those principal agencies involved in monitoring glaciers will expand their networks to cover the known gaps, although this is what is needed. In Canada, the original proposal for the IHD network included a plan to establish a network of benchmark glaciers that would be linked to the principal representative basins. Those proposals were well designed and have stood the test of time. Implementation of them at any time in the future would go a long way towards providing the coverage needed.

In the meantime, broad, regional inventory studies directly linked to the present sites would place existing measurements in a broader geographical context and give a better indication of how spatial variability might be applied to make better use of the data being collected.

In most of North America, it is clear that snow and ice play an important role in society and the landscape. Glaciers are an important and rapidly changing resource and element of the environmental system. However, this is not recognized by most North Americans, who generally live a significant distance from regions where glaciers are extensively found. For the northern hemisphere, remote Canadian and Alaskan glaciers and glacier systems, unlike those situated near or in the industrial heartlands of Europe, for example, are unique sources of information on natural variability. Therefore, while water resources and environmental systems will be affected by the consequences of a change in glacial extent regionally, Canadian and U.S. monitoring and research programmes are in a unique position to assess the consequences of climatic change, both regionally and in the global context.

The recognition of uniqueness and the opportunity to contribute in these contexts are paramount. Environmental monitoring and research, however, are often at odds with the planning framework of today's society in that not all activities meet an immediate operational need. In Canada, for example, the motivation to develop and maintain glacier inventories and monitoring programmes is often linked to industrial developments, such as hydropower schemes. Support and funding can thus be cyclical. Clearly, monitoring and related research is adequately justified by the need to understand the phenomenon of global change and, for example, the factors which cause glacier extent to expand and contract over time. Unfortunately, in Canada and other developed countries, monitoring is still viewed as a relatively low status scientific activity. As outlined in a recent review of glacier research in Canada (Munro, 1995, Demuth and Munro, 1995), it will be crucial to:

i) establish both short- and long-term societal and scientific issues as a basis for monitoring/research programmes,
ii) recognize that, in research into glacier systems undergoing unknown future change, monitoring, because it must remain flexible, is neither trivial nor routine, and
iii) expand the relevance of monitoring/research activities to other disciplines. It will then be possible to create an integrated, multi-disciplined monitoring/research environment where the conduct and management of monitoring and research is more relevant and productive and, by example, promoted by enduring levels of support and funding.

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8 Glaciers in South America

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8.1 INTRODUCTION

The presently glacierized area in South America is estimated at some 26,000 km^2 (IAHS(ICSI)/UNEP/UNESCO, 1989), with the bulk of this ice mass being found in the Patagonian Ice Fields and Tierra del Fuego. A great number of mountain glaciers exist in the Andes of Argentina, Bolivia, Chile, Colombia, Ecuador, Peru and Venezuela. The following text illustrates mainly the situation in Argentina, Bolivia, Chile and Peru.

8.2 DISTRIBUTION AND CHARACTERISTICS

Glaciers in South America occur along the high Andes. More than half of the Andean range is located in Chile and Argentina (Fig. 8.1). In northern Chile and Argentina (17°–27°S), the highest summits rise above 6,000 m a.s.l., with a high plateau (Altiplano) extending several tens of kilometres to the east. To the south, the Andes are concentrated along a narrower belt only a few tens of kilometres wide. Aconcagua, located in Argentina at 33°S, is the highest summit not only of South America but of the entire western hemisphere. South of 34°, the elevation of the highest summits decreases rapidly (Table 8.1).

In their northern part, the Chilean Andes start at latitude 17°S and straddle Bolivia to 23°S. The climate here is dominated by the subtropical high pressure of the Pacific with associated arid conditions. The main precipitation source comes from the Atlantic and occurs during the summer months,

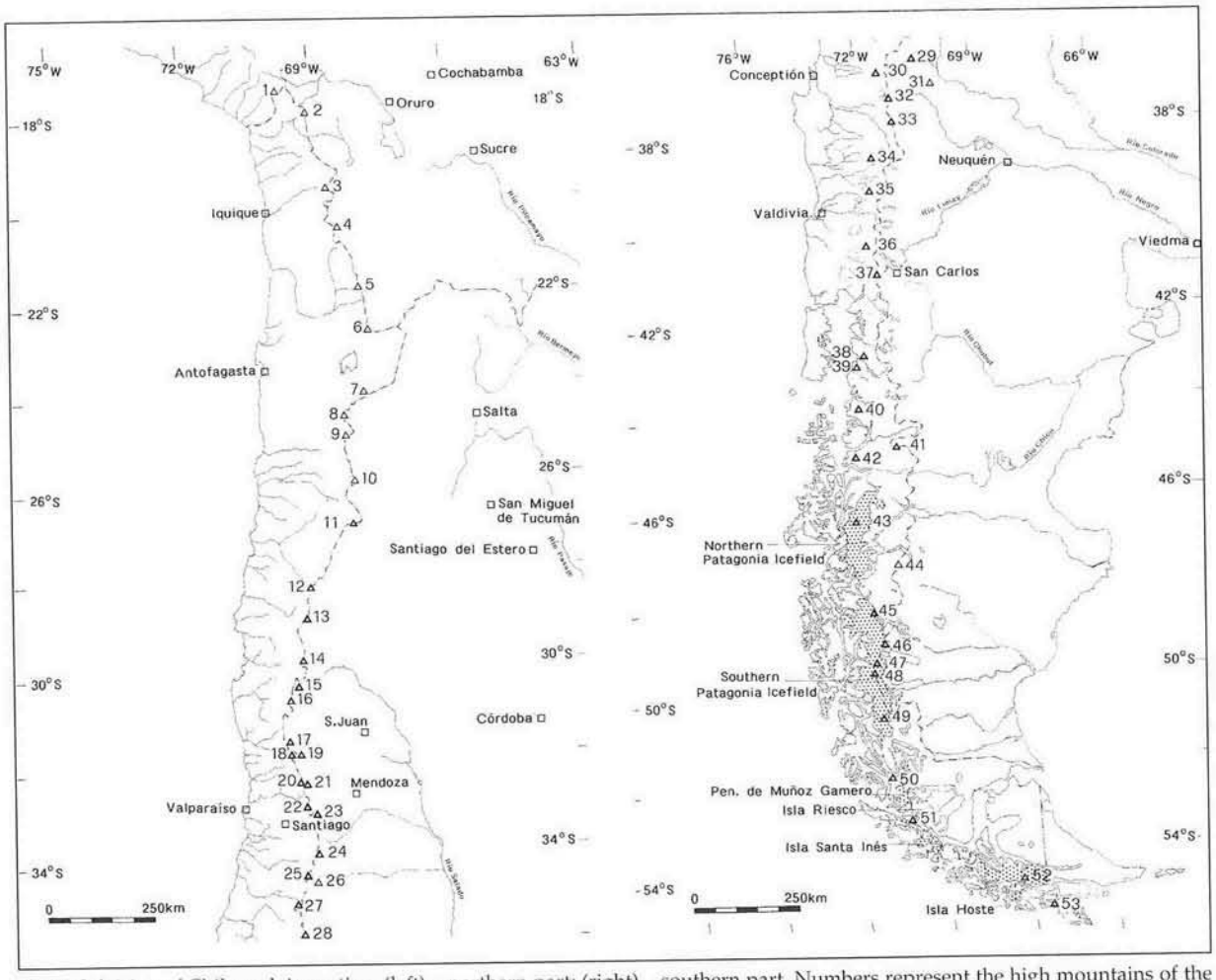


Figure 8.1 Map of Chile and Argentina. (left) – northern part; (right) – southern part. Numbers represent the high mountains of the Andes as shown in Table 8.1.

amounting to only a few tens of mm/ year in the mountain areas. The orographic effect on precipitation results here in an east-west gradient of the equilibrium line altitude (ELA, Fig. 8.2), the precipitation being higher in the east. The precipitation decreases southward as the distance from the humidity source in the Atlantic increases. Only very few mountain glaciers with areas generally smaller than 15 km^2 occur on high volcanoes in this area (e.g., Parinacota and Pomerape).

Of all the tropical regions (i.e., areas between the tropic of Capricorn and the tropic of Cancer), the mountains of Peru have the largest number of glaciers. At the same time, the Andes of Peru are densely populated. On the one hand, the local population benefits from continuous and reliable meltwa-

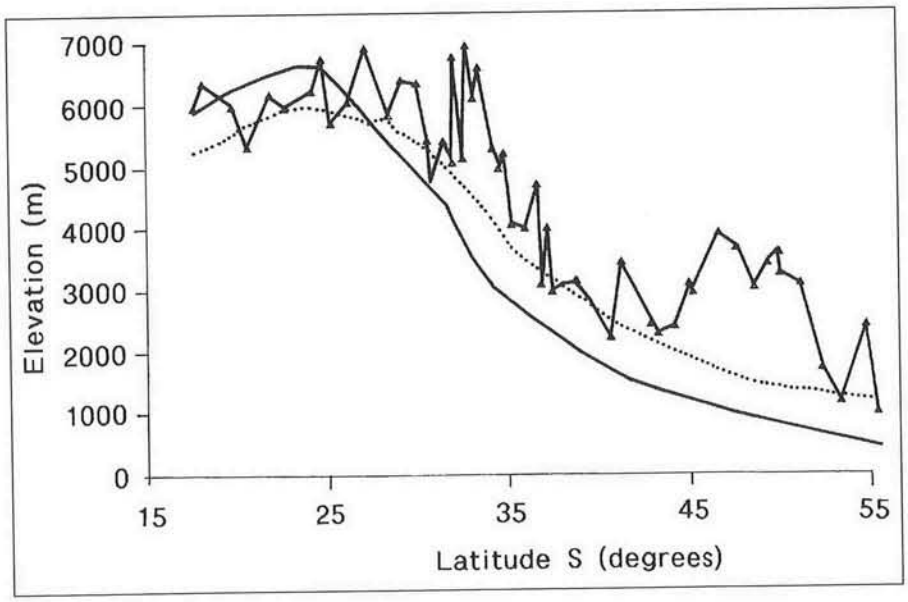


Figure 8.2 North-south variation of the regional snow-line on the western margin of the Andes (continuous line) and on the eastern side (dotted line), and highest elevation of peaks (line joining triangles). The snow-lines are adapted from Nogami (1972).

ter runoff but, on the other hand, is frequently affected by glacier catastrophes of various kinds. Catastrophic events caused by ice avalanches and proglacial lake outflows happen from time to time in all Cordilleras (high mountain ranges) of the Peruvian Andes. They have caused serious damage in the lower valleys, destroying towns, villages, roads, bridges, paths, etc. In many cases, people were killed. Since the catastrophes usually occurred unexpectedly and suddenly, there was little chance for people to escape. For example, during a particularly tragic event in 1941, at least 4,000 people drowned in the town of Huaraz as a result of the outflow from two small proglacial lakes in the Cordillera Blanca. This event led to the realization that it was necessary to (a) make an inventory of dangerous proglacial lakes and (b) lower or to strengthen the outlet of the most dangerous lakes if settlements were endangered and if this was technically at all possible. For this purpose, the Comisión de Control de Lagunas de la Cordillera Blanca was set up. The development of hydroelectric power further increased interest in glaciers, particularly in the Cordillera Blanca. At first, four glaciers were chosen for a pilot study on mass balance measurements. Two more glaciers were added to this network later on. Understanding glacier fluctuations in the Peruvian Andes is also of great importance in connection with the formation of new lakes as a result of rapid tongue retreat. In parallel with this work, cooperation with the Temporary Technical Secretariat for the World Glacier Inventory was initiated. The inventory was made possible by the availability of various sets of aerial photographs, some of which were taken in the aftermath of the biggest glacier catastrophe to occur in Peru in recent centuries, the gigantic rock and ice avalanche from Nevado Huascarán in 1970.

From 23°S to 51°S, the Andes are shared by Argentina and Chile. Climatic conditions from 23° to 30°S range from arid to semi-arid, with precipitation increasing slightly south of 25°S as the region affected by the westerlies is approached. The precipitation increase results in an ELA decrease to the west from 6,600 m at 24°S to 4,700 m at 30°S (Fig. 8.2). The transition from an eastern precipitation source to a western source appears in Fig. 8.2 as the intersection of the ELA on the western side of the Andes and the ELA on the eastern side at 27°S. Glacierization is limited in this area to very few small mountain glaciers (in general less than 1 km^2) occurring on the highest summits (e.g., Cerros Colorados, Tres Cruces, Los Tronquitos). At 30°S, a clear influence of the westerly circulation occurs, with pronounced seasonality in precipitation during the winter months associated with low pressure systems from the Pacific. By contrast, summer conditions are dry owing to the presence of the subtropical high-pressure system. A marked precipitation increase explains the existence of important mountain and valley glaciers from 32°S to 35°S, where many summits exceed 5,000 m a.s.l. and a few even 6,000 m

TABLE 8.1 Highest Andean summits along a north-south transect from the Bolivian-Chilian border to the very south of the American continent.

	Mountain	*Latitude S* (°)	(′)	*Elevation* (*m*)
1	Volcán Tacora	17	43	5,988
2	Volcán Parinacota	18	10	6,330
3	Cerro Sillajhuay	19	45	5,995
4	Cerro Copa	20	37	5,320
5	Volcán San Pedro	21	54	6,154
6	Sairecábur	22	42	5,970
7	Pular	23	11	6,225
8	Llullaillaco	24	42	6,723
9	Volcán Lastarria	25	11	5,700
10	Cerros Colorados	26	11	6,049
11	Ojos del Salado	27	07	6,880
12	Cerro del Potro	28	23	5,830
13	Cerro del Toro	29	07	6,380
14	Cerro de la Tórtolas	29	57	6,332
15	Cerro Alto	30	34	5,430
16	Cerro Maruez	30	42	4,750
17	Chanchones	31	56	5,370
18	Cerro Ojotas	31	57	5,070
19	Mercedario	31	28	6,770
20	Cerro Volcán	32	32	6,130
21	Cerro Aconcagua	32	39	6,959
22	Nevado Juncal	33	04	6,110
23	Volcán Tupungato	33	22	6,550
24	Maipo	34	11	5,290
25	Alto los Arrieros	34	37	4,986
26	Sosneado	34	45	5,159
27	Volcán Peteroa	35	15	4,090
28	Campanarío	35	56	4,002
29	Volcán Domuyo	36	37	4,709
30	Chillán	36	50	3,122
31	Volcán Tromen	37	08	3,979
32	Volcán Antuco	37	25	2,985
33	Volcán Callaqui	37	55	3,080
34	Volcán Llaima	38	42	3,124
35	Volcán Villarrica	39	25	2,840
36	Volcán Puyehue	40	35	2,240
37	Monte Tronador	41	10	3,460
38	Nevado Minchinmávida	42	52	2,470
39	Volcán Corcovado	43	12	2,300
40	Cerro Melimoyu	44	05	2,400
41	Cerro Catedral	44	56	3,060
42	Cerro Macá	45	06	2,960
43	San Valentín	46	37	3,910
44	Cerro San Lorenzo	47	37	3,700
45	Cerro Mellizo Sur	48	33	3,050
46	Cerro Fitz Roy	49	17	3,441
47	Murallón	49	48	3,600
48	Cerro Bertrand	49	57	3,270
49	Cerro Paine Grande	51	02	3,100
50	Monte Burney	52	20	1,750
51	Monte Córdova	53	19	1,219
52	Monte Darwin	54	44	2,438
53	Monte Hardy	55	24	1,036

a.s.l. (e.g., Mercedario, Aconcagua, Juncal Tupungato, Marmolejo). South of 35°S, the highest summits rarely exceed 4,000 m, which restricts the development of glaciers to mainly volcanic cones (e.g., Peteroa, Domuyo, Antuco, Callaqui, Llaima, Lanín, Osorno) to 42°S. Precipitation increases from about 200 mm/year at 30°S to 2,000 mm at 40°S. There is a strong orographic effect on circulation, with an important reduction in precipitation on the eastern side of the Andes. South of 42°S, in the region known as Patagonia, the climate is wet temperate, being completely dominated by the westerly circulation, with high precipitation and reduced seasonality. This results in a larger glacierized area. Three small ice caps from 40 to 100 km² exist on the Chilean side from 42°S to 44°S: Minchinmávida, Yanteles and Melimoyu. The Patagonian Andes have been carved extensively by Pleistocene glaciers, resulting in a complex pattern of fjords both in the Pacific and in large pedemontane lakes to the east. The orographic effect on circulation in Patagonia is highly pronounced, with precipitation amounts in excess of 5,000 mm/year in certain areas to the west and as little as 200 mm/year or less on the Pampa to the east. South of 46°S, the two largest ice bodies of the southern hemisphere outside Antarctica are to be found: the Northern Patagonia Ice Field, with an area of about 4,200 km² (Aniya, 1988), and the larger Southern Patagonia Ice Field covering about 13,000 km² (Aniya *et al.*, 1992). Glacier Upsala, a 60 km-long outlet on the eastern margin of the Southern Patagonia Icefield, is the largest glacier in South America.

From 51°S to 55°S, the Andes lie completely in Chile, changing from a north-south to an east-west orientation. In this area, the third-largest glacier body in South America is found: Cordillera Darwin in Tierra del Fuego, with an estimated area of 2,000 km². Smaller ice caps exist in Peninsula Muñoz Gamero, Isla Santa Inés and Isla Hoste. At 55°S, near Ushuaia in Tierra del Fuego, a subdued portion of the Andes with reduced glacierization is again shared with Argentina.

Glaciers in Bolivia, Colombia, Ecuador, Peru and Venezuela are mostly mountain glaciers existing under tropical climatic conditions. Recent studies carried out in the Central Andes (Thompson, 1992; cf. also Hastenrath and Kruss, 1992, for Kenya) suggest that the effects of global warming could be more pronounced in the short term for tropical glaciers like these than for glaciers of high and medium latitudes. Other studies also indicate that tropical glaciers are excellent indicators of short-term climatic variations, as seen in the Andes by their response to anomalies of short duration and of variable intensity resulting from the El Niño phenomena (ENSO) (Thompson *et al.*, 1984; Francou *et al.*, 1995; Ribstein *et al.*, 1995). The sensitivity of tropical glaciers to climatic variability is still inadequately explained, mostly due to the lack of data available on net mass balances and on the climatic variables that control these glaciers. This sensitivity could be due, at least in part, to a particular condition affecting the mass balance which is that, on these glaciers, the accumulation period stretches throughout the summer and is therefore synchronous to the period in which ablation is at its maximum. Thus, a deficit in precipitation during this season translates directly into strong ablation tied to the increase in direct radiation. It could also be suggested that, with the summer rainfall, an increase in the temperature would shift the snow-rain boundary to higher altitudes, causing a significant contribution in sensible heat over the majority of the glacier (Lliboutry *et al.*, 1977*a*).

8.3 EXISTING INVENTORIES

Early glacier inventories of Chile and Argentina have been summarized by Mercer (1967). The distribution of glaciers in Chile has been described by Lliboutry (1956), who was the first to conduct detailed glaciological work in Chile. In the late 1970s, the Dirección General de Aguas of the Ministry of Public Works started a detailed glacier inventory programme. As part of this programme, glaciers have been inventoried in the Río Maipo basin (33° to 34°S; Maragunic, 1979), in the Río Cachapoal basin (34°S; Caviedes, 1979), in the Río Aconcagua catchment (32°S; Valdivia, 1984), in the Río Mataquito basin (35°S; Noveroy, 1987), in the north of Chile (18° to 32°; Garín, 1986) and in the Lake District (37° to 41° 30′S; Rivera, 1989). In addition, a preliminary glacier inventory based on 1:250,000 maps has been compiled for the Northern Patagonia Ice Field by Valdivia (1979*a*, 1979*b*). A more precise and complete glacier inventory for the Northern Patagonia Ice Field based on 1:50,000 cartography has been compiled by Aniya (1988). Table 8.2 summarizes the glacier inventory in Chile. The total area covered by detailed glacier inventories is 5,515 km², representing approximately only one-fourth of the total glacier area in Chile (cf. Table 8.3).

In Argentina, Helbling and Reichter explored the high Cordillera between Mt. Aconcagua and Mt. Tupungato from 1907 to 1912 (Reichter, 1929; 1967). Helbling (1919; 1935; 1940) published an accurate map of the Río del Plomo valley (33°S) on a scale of 1:25,000, describing the glaciers and fluctuations of the glaciers' termini since 1909–1934; Groeber (1947*a*; 1947*b*; 1951; 1955) studied the geology and described the glaciers in the Central Andes; Lliboutry (1956) published maps and measured the englaciated area between 32°30′S and 35°S and the glaciers in Patagonia. Feruglio (1957) described the glaciers in the Cordillera Argentina between 21°S and 51°S. An inventory of Patagonian glaciers was first undertaken by Lliboutry (1956) and Bertone (1960). In the provinces of Mendoza (33°S) and San Juan (31°S), the Instituto Argentino de Nivología y Glaciología (IANIGLA-CONICET) has compiled glacier inventories (IAHS(ICSI)/UNEP/UNESCO, 1989) in the following basins: Río Mendoza: Corte and Espizua (1978; 1981) identified 1,025 ice bodies bigger than 0.02 km² that

TABLE 8.2 Inventoried glaciers in Chile

Region	*Basin*	*No. of Glaciers*	*Area (km²)*	*References*
I	*		29.70	
II	*	14	12.13	
III	*	49	66.83	
IV	*	11	7.02	
V	Aconcagua	267	151.25	Valdivia 1984
Metropolitán	Maipo	647	421.90	Marangunic 1979
VI	Cachapoal	146	222.42	Caviedes 1979
VI	Tinguiririca	261	106.46	Valdivia 1984
VII	Mataquito	81	81.91	Noveroy 1987
VIII-IX	Bío Bío	29	52.37	Rivera 1989
IX	Imperial	13	18.72	Rivera 1989
IX-X	Toltén	14	68.48	Rivera 1989
IX-X	Valdivia	6	42.33	Rivera 1989
X	Bueno	11	19.35	Rivera 1989
X	Maullín	1	2.84	Rivera 1989
X	Chamiza	1	1.05	Rivera 1989
X	Petrohué	12	60.57	Rivera 1989
XI	Northern Patagonia Icefield·	**28	4,200.00	Aniya 1988
Total		***1,600	5,515.33	

* Only a few glaciers in the North of Chile drain into distinct rivers and they have not been classified into basins.

** This number reflects the outlet glaciers of the Northern Patagonia Icefield. Contiguous mountain glaciers, although included in the total area, are not counted individually (Aniya 1988).

*** Five glaciers in Regions VIII, IX and X drain into two different basins. This is accounted for in the total.

TABLE 8.3 Uninventoried glaciers in Chile

Latitude	*Region*	*Description*	*Estimated Area (km²)*
35° to 37° S	VII & VIII	Río Maule and Rio Itata basins, glaciers limited to a few high peaks (e.g., Volcán San Pedro, 3,499 m)	50
41° 30′ to 45° 30′ S	X & XI	Continental Chiloé and northern Aisén, three main ice caps on Mt. Minchinmávida (2,470 m), Mt. Yanteles (2,042 m), and Mt. Melimoyu (2,400 m) (Lliboutry 1956), plus many smaller mountain glaciers	*250
45° 30′ to 49° S	XI	North, East and South-east of Northern Patagonia Icefield, and North-east of Southern Patagonia Icefield. Many mountain glaciers, e.g., Volcán Hudson, Cerro San Lorenzo	400
48° 15′ to 52° S	XI & XII	Southern Patagonia Icefield, total area 13,000 km² (Aniya *et al.*, 1992), of which 90% is claimed by Chile (Martinic 1982)	11,700
51° 45′ to 52° S	XII	Cordillera Sarmiento, ice caps and mountain glaciers, first explored by Miller	100
52° 40′ to 53° S	XII	Península Muñoz Gamero, icefield on south-western part of peninsula	*200
52° 50′ to 53° 20′ S	XII	Isla Riesco mountain glaciers	100
53° 45′ to 54° S	XII	Isla Santa Inés, icefield on eastern part of island	*250
54° 20′ to 54° 50′ S	XII	Cordillera Darwin, 150 km-long icefield, 25 km wide in central part	2,000
55° 10′ S	XII	Isla Hoste, ice cap located on Península Cloue, western part of the island, plus smaller glaciers to the East	150
Total			15,200

Note: * indicates areas estimated by Lliboutry (1956). Otherwise, area estimations are made in this study by inspection of 1:500,000 scale maps.

covered 304 km² of bare ice and 344 km² of debris-covered ice. Río Tunuyán: eastern slopes of Cordón del Plata and Cordón del Portillo (the glacierized area is 144 km², 40% of which corresponds to uncovered ice and 60% to covered ice) (Espizua, 1983*a*; 1983*b*). In the Río Atuel: Cobos (1979, unpublished IANIGLA report) revealed a glacierized area of 186 km² consisting of 80% of uncovered ice and 20% of covered ice. Río Malargüe: in this basin, the englaciated area covered 12 km², 2% of bare ice and 9.5% of ice covered by debris (Cobos, 1987). The Río San Juan catchment, with Ríos de los Patos, Blanco, Calingasta, Ansilta and Castaño (Aguado, 1983; 1986 (unpublished IANIGLA report); Espizua and Aguado; 1984) has a total glacierized area of 556 km² (42% of uncovered ice and 58% of ice covered by debris). Glacier inventories from the Patagonian Andes, Argentina, have been published by Rabassa *et al.*, 1975; Rabassa *et al.*, 1978*a*; 1978*b*; 1978*c*; Rabassa, 1980; 1981; Rabassa *et al.*, 1981; Rabassa, 1983. At 55°S, Lendaro and Iturraspe studied the Martial Glacier and Roig (1990) studied the geomorphology and hydrology of de cirque glaciers in the Tierra del Fuego Andes. Rabassa *et al.* (1981; 1982) compiled a glacier inventory of the James Ross and Vega Islands in the Antarctic Peninsula.

Detailed glacier inventories (cf. IAHS(ICSI)/UNEP/UNESCO, 1989) were also compiled in Bolivia (Cordillera Occidental and Cordillera Oriental, Jordan, 1991), Peru (Northern and Southern Cordilleras, Ames *et al.*, 1988) and partially in Ecuador, Colombia (Sierra Nevada de Santa Marta and various volcanoes) and Venezuela (Sierra Nevada de Mérida). In Peru, the Glaciology Division of Electroperú in Huaraz completed an inventory which is mainly based on aerial photogrammetric flights conducted between 1955 and 1970. The flights in 1970 were undertaken by NASA and the Servicio Aerofotográfico Nacional (SAN) after the catastrophic earthquake on May 31. These photographs were recorded on infrared film and were used for the inventory of the Cordillera Blanca. The 1955 material was used for Cordillera Ampato and some zones of the Cordilleras Chila and Huanzo (cf. Fig. 8.3). The rest of the inventory is based mainly on 1962 pictures. The bulk of the Peruvian ice masses are located in the Cordilleras Blanca, Vilcanota, Ampato and Central (Ames *et al.*, 1988; cf. Table 8.4). The total glacier area (excluding the insignificant amounts in the Cordilleras Barroso and Volcanica) is 2,042 km². Altogether, 3,044 glaciers were recorded. Fig. 8.4 shows some interest-

Figure 8.3 Distribution of glaciated Cordilleras (high mountain ranges) in north, central and southern Peru. The true extent of glaciated areas cannot be shown on this scale.

TABLE 8.4 Number, total area and estimated total ice volume in the Peruvian Cordilleras according to the Peruvian Glacier Inventory

Cordillera	*Number of glaciers*	*Total area (km²)*	*Total volume (km³)*
1 Blanca	722	723	22.60
2 Huallanca	56	21	0.43
3 Huayhuash	117	85	2.99
4 Raura	92	55	1.33
5 La Viuda	129	29	0.43
6 Central	236	117	2.54
7 Huagoruncho	80	23	0.40
8 Huaytapallana	152	59	1.15
9 Chonta	95	18	0.26
10 Ampato	93	147	5.12
11 Vilcabamba	98	38	0.72
12 Urubamba	90	41	0.78
13 Huanzo	115	37	0.60
14 Chila	87	34	0.58
15 La Raya	48	11	0.16
16 Vilcanota	469	418	12.00
17 Carabaya	256	104	1.96
18 Apolobamba	109	81	2.11
19 Volcanica	-	-	-
20 Barroso	-	-	-
Total	3,044	2,042	56.15

ing statistical results from the Cordillera Blanca. Although this mountain range appears most spectacular to any visitor to the Santa Valley, no glacier is larger in surface area than 16.5 km² (Jankapampa). The longest glacier is Copap (7.0 km). Of particular interest is the distribution of average glacier altitudes (Fig. 8.4b). On the eastern side, average altitudes are sometimes lower than 4,800 m a.s.l. and typically around 5,000 m. On the western side of the Cordillera, they are significantly higher. Here, average equilibrium lines are usually above 5,100 m and sometimes even above 5,400 m. This can be interpreted as an effect of precipitation distribution since the main advection of moisture is from the Amazon basin in the east. Nevertheless, quite a large part of the glacier surfaces are oriented towards the southwest (cf. Fig. 8.4c) because the high valleys on this side of the Cordillera Blanca are somewhat less steep than on the other, thus providing relatively extensive accumulation areas at high altitudes. Similarly interesting relationships can also be found when comparing the Cordilleras La Viuda, Central, Huaytapallana and Chonta (IAHS (ICSI)/UNEP/UNESCO, 1989). It is to be expected that a general glacier retreat as a result of continued climatic warming will soon lead to a total loss of glaciers in the Cordilleras Chonta, Huallanca or La Raya, where only a few small glacierlets can be found.

8.4 EXISTING LONG-TERM OBSERVATIONS (LENGTH CHANGE, MASS BALANCE, MAPS)

In Chile and Argentina, there are several cases of surging glaciers near Mendoza (Glaciar Horcones Inferior and Glaciar Grande del Nevado del Plomo) and Santiago (Juncal Sur, Museo and Colina). A few glaciers in the Southern Patagonia Ice Field are presently advancing and Moreno Glacier is largely stable. In contrast to the above glaciers, most glaciers in Chile and Argentina are presently retreating, as indicated by recent moraines and their present behaviour. The only glacier in Chile presently monitored for mass balance is Echaurren Norte Glacier at 33°S. Monitoring has been done by the Dirección General de Aguas routinely in spring, summer and autumn since 1975. Results have been published for the periods 1975–1983 (Peña *et al.*, 1984) and 1975–1994 (Escobar *et al.*,, in press) and the programme still continues today. With respect to frontal variations, the only systematic studies based on aerial photographs and satellite imagery are on glaciers of the Northern Patagonia Ice Field (Aniya, 1988; 1992; Aniya and Enomoto, 1986; Warren, 1993) and the Southern Patagonia Ice Field (Aniya *et al.*, 1992; Warren and Sugden, 1993). Sporadic observations of glacier variations elsewhere in Chile are described in Lliboutry (1956), Mercer (1962; 1967) and Marangunic (1964*a*; 1964*b*).

In Argentina, some glaciers are presently monitored by IANIGLA for mass balance in the Río Mendoza basin. Mass-balance measurements in the Cajón del Rubio area began in 1979. Balances were calculated continually up to 1987 but no measurements were taken in 1986, 1988 and 1990. Balance measurements re-started in 1991. The authors reconstructed the accumulation data at Piloto Glacier for the missing years and analysed the mass balance behavior of Piloto and Alma Blanca Glaciers (Leiva *et*

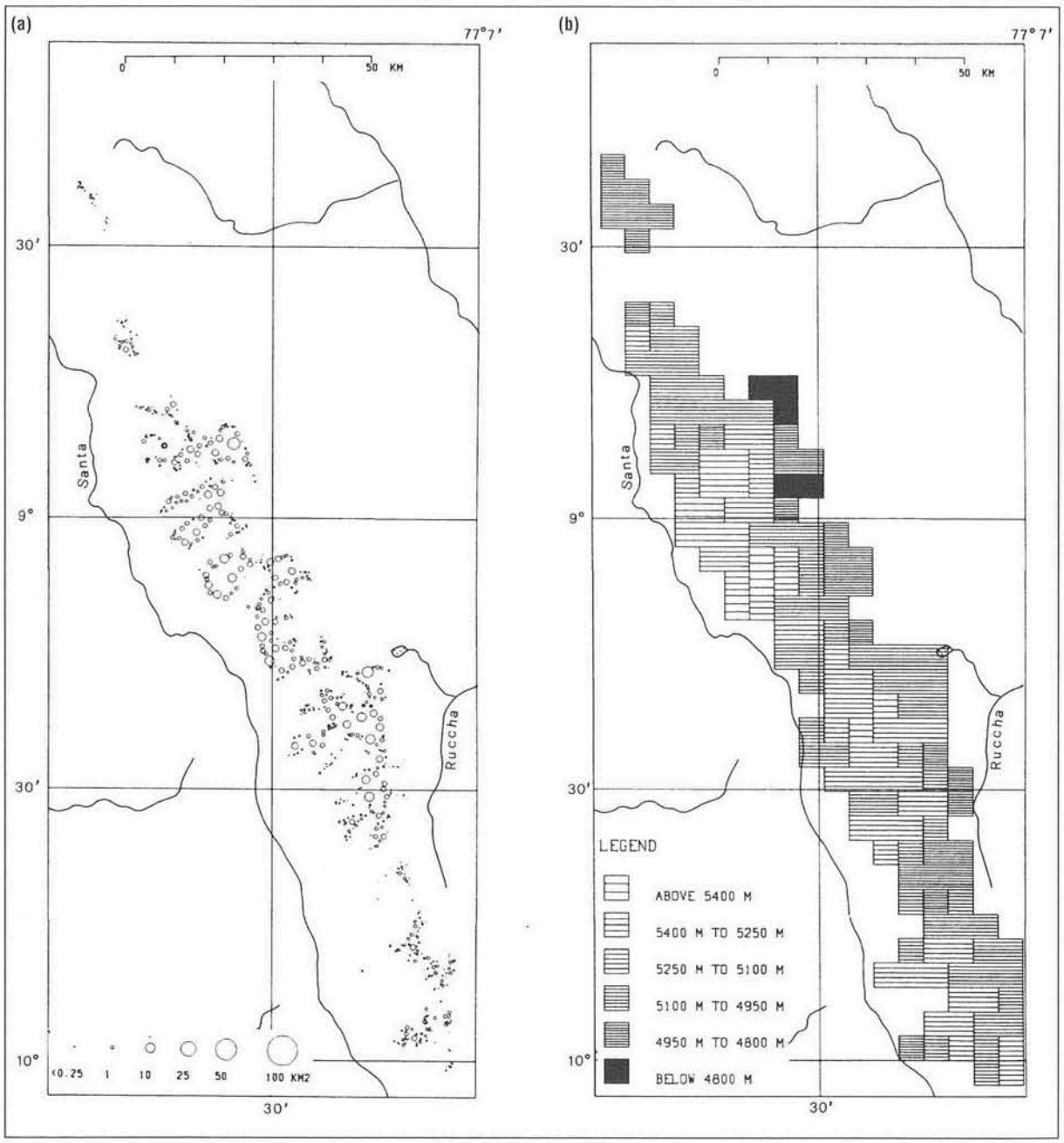
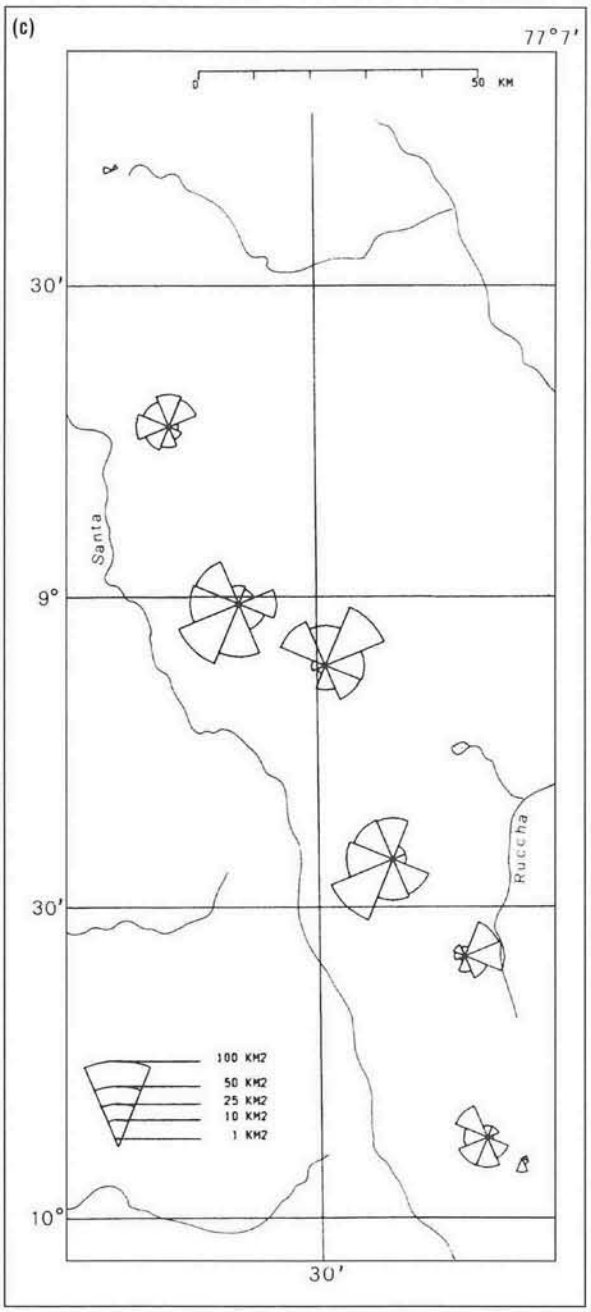


Figure 8.4 (a) Glaciers in the Cordillera Blanca according to surface area; (b) mean glacier elevations in the Cordillera Blanca (notice that the mean elevation is somewhat lower on the eastern side compared with the western side of this mountain range);

al., 1986; Leiva and Cabrera, in press). Regarding glaciers' frontal variations, studies have been made in the following basins: Río del Plomo basin (33°S) (Espizua, 1986; 1987; Espizua and Bengochea, 1990; Leiva *et al.*, 1989; Llorens and Leiva, in press); Río Tunuyán basin (33°15'S) (Llorens and Leiva, 1994); Rio Atuel basin (34°S) (Cobos and Boninsegna, 1983); Río Frías basin (41°S) (Villalba *et al.*, 1990). Mercer (1965; 1968) studied the fluctuations of ice margins in Patagonia. Fluctuations on Castaño Overo Glacier, Mount Tronador, in Northern Patagonia have been studied by Bertani *et al.*, 1985; 1986; Brandani *et al.*, 1986 and on Martial Glacier in Ushuaia (55°S) by Lendaro and Iturraspe. A Glacier Research Project in Patagonia that included *Characteristics of Recent Glacier Variations in Patagonia, Southern Andes* was carried out during the summer of 1993–94 at and around the Upsala, Ameghino and Moreno Glaciers (Skvarca *et al.*, 1995; Naruse *et al.*, 1995; Takeuchi *et al.*, 1995; Aniya and Sato, 1995).

Until recently, there were very few mass balance studies available for the Central Andes comparable to those undertaken for tropical African glaciers, notably Lewis Glacier (Hastenrath, 1984). This lack of work is all the more regrettable in that the Central Andes represent more than 95% of the surface area of glaciers found in the Tropics. Over the last fifteen years or so, only the Cordillera Blanca (Peru) has provided data on the balances of ablation areas of two glaciers, Uruashraju and Yanamarey (Ames, 1985; Kaser *et al.*, 1990), in addition to data over several decades concerning the oscillations of the gla-



(c) summarized surface areas and aspects of glaciers according to catchment basins. After IAHS(ICSI)UNEP/UNESCO, 1989.

cier's termini. From ice cores taken from the Quelccaya Ice Cap (Peru), we know that the Little Ice Age commenced in the Central Andes around 1480 A.D. and ended around 1880 A. D. (Thompson *et al.*, 1986). According to photographic documentation analysed by Broggi (1945) in the Peruvian Andes, the retreat at the end of the 19th and beginning of the 20th centuries is interrupted by a phase of advance between 1909 and 1932. This advance is itself followed by a significant retreat in the period 1932–1945.

In Peru, systematic length measurements of glacier tongues and mass balance measurements began in 1968 in the Cordillera Blanca. The aim was to estimate the contribution of meltwater runoff to the water available for hydroelectric power production. Since 1940, a general tendency towards glacier retreat has been documented by aerial photographs taken in 1948 and 1962 and by topographic surveys carried out every year since 1968 at the terminus of about half a dozen glaciers of the Cordillera Blanca (Ames, 1985; Kaser *et al.*, 1990). In the early 1970s, an ablation stake network was installed and monitored on the tongues of Uruashraju and Yanamarey. Since 1980–81, the stakes have been measured at the end of both the wet and dry seasons (Kaser and Ames, 1990). Owing to the problems associated with very high altitudes, stakes could only be maintained up to 4,900 m a.s.l. It is noteworthy that accumulation mainly takes place during the rainy season, which usually lasts from October until March or April. There is hardly any seasonal temperature variation. Therefore, ablation takes place at any time of year. The longest record of length fluctuations is detained by the Broggi (located north of Nevado Huascarán, 1.1 km long), Uruashraju (2.5 km) and Yanamarey (1.7 km) Glaciers. The latter two are located in the southern part of Cordillera Blanca. Later on, the Gajap, Huarapasca and Pastoruri Glaciers completed the observational programme. The selection of regularly visited glaciers had to be done mainly from the point of view of accessibility. Additional glacier snout positions could be reconstructed using aerial photographs (1948 and 1962). The cumulative length changes as recorded by field measurements (Fig. 8.5) indicate a relatively modest retreat from 1948 until the end of the 1970s. Uruashraju and Yanamarey even recorded a very slight readvance around 1975. Between 1948 and 1980, Broggi Glacier retreated, on average, 11 m per year. Nevertheless, the total length loss of 366 m during this period amounted to one-third of the glacier's length in 1970! From 1980 to 1991, the retreat accelerated. Broggi lost an average of 26.2 m per year. In 1991, it retreated by 53.2 m, the biggest single-year retreat ever recorded. The glacier was then 654 m shorter than in 1948. The other glaciers, including the newly-measured ones, retreated, on average, between 17.2–21.1 m per year. Clearly, this pronounced ice loss is a general phenomenon in the Cordillera Blanca: maps and terrestrial photogrammetric pictures produced by German-Austrian expeditions in the 1930s (Kinzl, 1940*b*; 1942; Kinzl and Schneider, 1950) show various small glaciers which have since disappeared. Until now, it has been difficult to relate the observed glacier retreat to local climatic data, although a simple meteorological station has been maintained at Querococha, a site 3,980 m a.s.l. and 8.5 km below Yanamarey. Precipitation measurements commenced in 1954 and temperature recordings in 1964. During this interval, the recordings do not show a clear, systematic increase in temperature. However, at least the small readvance of Yanamarey and Uruashraju, as well as the decelerated retreat of Broggi around 1975, correlates with slightly lower temperatures and somewhat increased precipitation in the early 1970s. Given that the analysed glaciers were very short, their length varying between

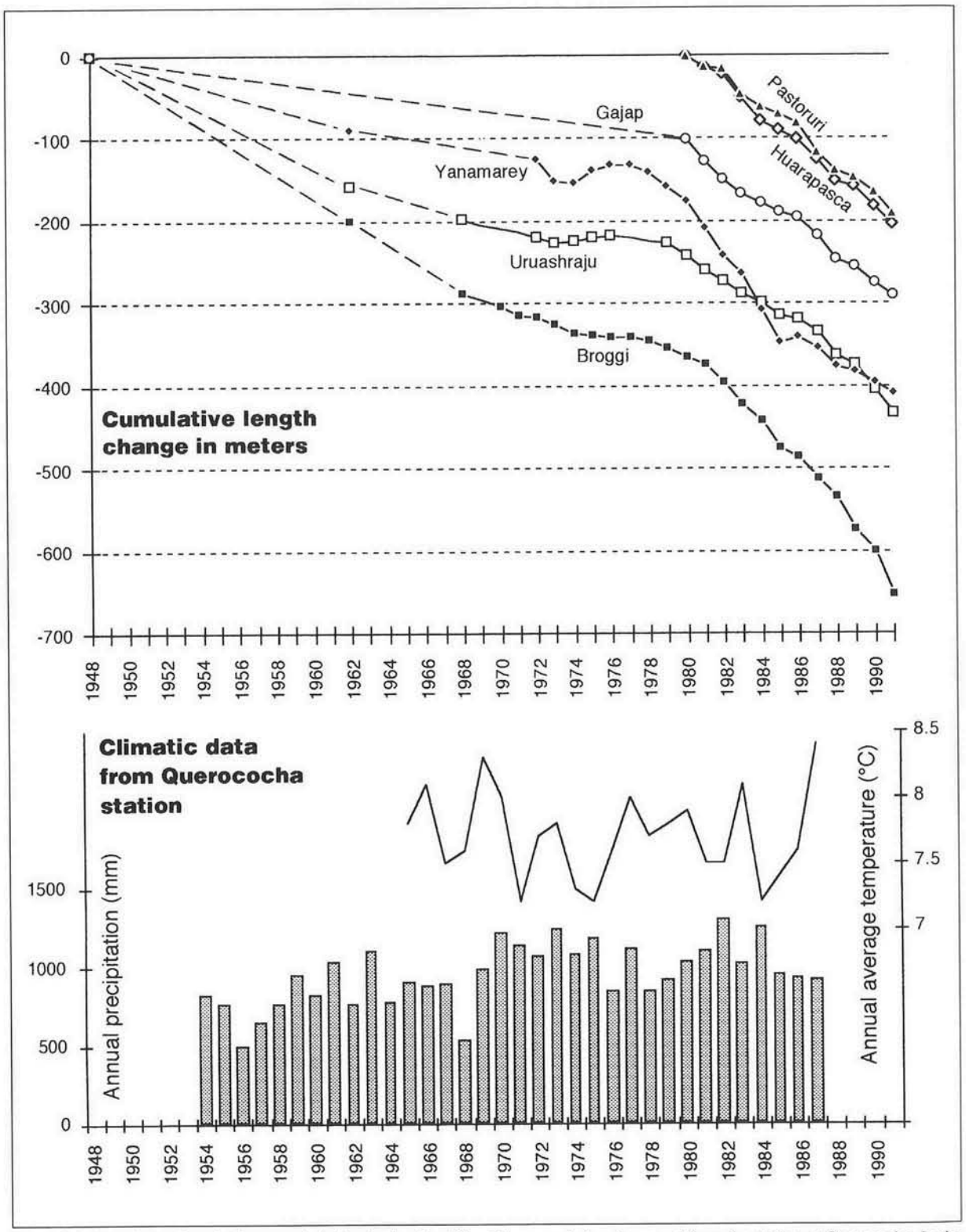


Figure 8.5 Cumulative length changes of glaciers in the Cordillera Blanca and climatic record from the station at Querococha. In the upper diagram, dashed lines show length changes inferred from aerial photographs. The 1948 snout positions are taken as zero.

1 km to 2.5 km, changes in balance were almost immediately followed by changes in terminus position. An acceleration in the rate of retreat is evident for all these glaciers starting from 1982 (Table 8.5): the rate of the retreat is three times that of the average retreat in the preceding decades. It is possible to compare this acceleration to that measured by photogrammetry on the terminus of Quelccaya Ice Cap. According to Brecher and Thompson (1993), the rate of retreat was three times faster between 1983 and 1991 than between 1963 and 1978, while the mass volume loss was seven times greater.

Since 1991, there has been an ongoing programme of glacier monitoring and hydrology in the Cordillera Real of Bolivia, the Cordillera Blanca of Peru and some volcanoes in Ecuador, under the auspices of the French Scientific Research Institute for Development in Cooperation (ORSTOM) (Francou *et*

TABLE 8.5 Length variation of 3 glaciers of Cordillera Blanca (1948–1993)

Length Variation	*Broggi*	*Uruashraju*	*Yanamarey*	
1948–1993	–16.0	–11.0	–9.0	(1)
1948–1981	–11.0	–8.0	–6.0	(1)
1982–1993	–30.0	–19.6	–18.6	(1)
Total Length in 1982	1.2	2.5	1.6	(2)

(1) metres per year
(2) kilometres
Source: Oficina de Recursos Hidricos; Electroperú, Huaraz, Peru.

al., 1995). The results obtained in Bolivia from monthly measurements of glacial and hydrological balances of the Zongo and Chacaltaya Glaciers are sufficiently definitive to be taken as references for other glaciers in the Tropical Andes. These results present the only measurements of mass balances and hydrological balances of glacierized basins available to date in the Central Andes. Despite the exceptional historical documentation dating from the 18th century available on glaciers in Ecuador (Hastenrath, 1981), it is still impossible to provide a precise explanation for the glacial retreat which started at the end of the Little Ice Age. All we can ascertain from the work of Meyer (1907), confirmed by information collected by Broggi (1945) in Peru, is that glacial retreat has been occurring since at least 1870. The disappearance of numerous glaciers from a number of volcanoes which were ice-covered during the 19th century, such as Pichincha and Sincholagua, suggests that the snow-line on the Western Cordillera rose, between 1800 and 1975, from 4,650 m to 4,950 m (Hastenrath 1981). The readvance of Peruvian glaciers around the mid-1920s is equally documented in Bolivia (Müller, 1985; Jordan, 1991). The analysis of the compelling ground-based photographic archives assembled by Kinzl (1940*a*) in 1930 and 1940 will allow, in the near future, precise determination of the variations of glaciers' termini in the Cordillera Blanca twenty years before the first aerial photographs.

On the Chacaltaya Glacier (Francou, unpublished data), according to photographic documentation, a guide mark dating from 1982 and topographic measurements at the terminus every year since 1991, the retreat is estimated at 2.0 m/year between 1940 and 1993. It ranges from an average of 0.95 m/year for the period 1940–1982 to 6.05 m/year (a ratio of 1 to 5) for the period 1982 to 1993. These results tally with those obtained for the African glaciers in the Ruwenzori (Kaser and Noggler, 1991) and on Mount Kenya (Hastenrath and Kruss, 1992). They suggest that, if the tendency is to persist, numerous minor glaciers may disappear in the next two or three decades. The balance of glaciers at medium and high latitudes is determined, above all, by the temperature level of the ablation season (summer), which lasts approximately 3–4 months (Martin, 1977; Lefauconnier and Hagen, 1990). In the Tropics, because the warm season (which favours ablation) is synchronous to the rainy season (which favours accumulation), the effect of seasonal climatic variability, which could influence the net yearly balance, stretches over at least six months. To analyse the influence of seasonal variability, the present authors have measured ablation and accumulation on a monthly basis on the Zongo and Chacaltaya Glaciers, two glaciers under study in the Cordillera Real of Bolivia (Fig. 8.6 and 8.7). The results obtained over three years and partially published (Francou *et al.*, 1995) show that, with respect to net balance, a year can be divided up into three periods (Fig. 8.8).

1) The early summer months of October, November and December, preceding the rainy period, are months of strong ablation. Ablation is also strong during the 1 or 2 months at the end of the rainy season, March and April. In November-December, the firn line is situated at more than 5,500 m, well within the accumulation zone. The heavy rate of ablation during this period is due to a combination of factors, amongst which: the large quantity of energy at the top of the atmosphere (months around the summer solstice); the mostly cloud-free skies; the not very high albedo because the ice is still incompletely covered with snow, and the contribution of sensible heat owing to the increase in the hygrometry.
2) The months of accumulation, January-February (sometimes March), correspond to the highest levels of rainfall. It should be noted that the energy available for melting, by direct radiation or by sensible heat, remains sufficiently high to induce strong ablation in the lowest part of the glacier, close to the terminus. Since, at the same time, a period of high accumulation dominates over the greatest part of the surface, the activity coefficient of the glacier is high for this season, close to 0.25 m / 100 m in water-equivalent.
3) The months of winter, May–August, which are generally dry and cold, are months in which the net balance ought to remain stable. Cloud cover is low but the energy at the top of the atmosphere is 30 % less than in summer and it is just able to melt or sublime the snowfall accumulation over the period. The energy available for melting is all the more insufficient in that the amount of sensible heat is also very low. It should be noted, however, that winter is the season in which the quantities of energy received by the opposite facing

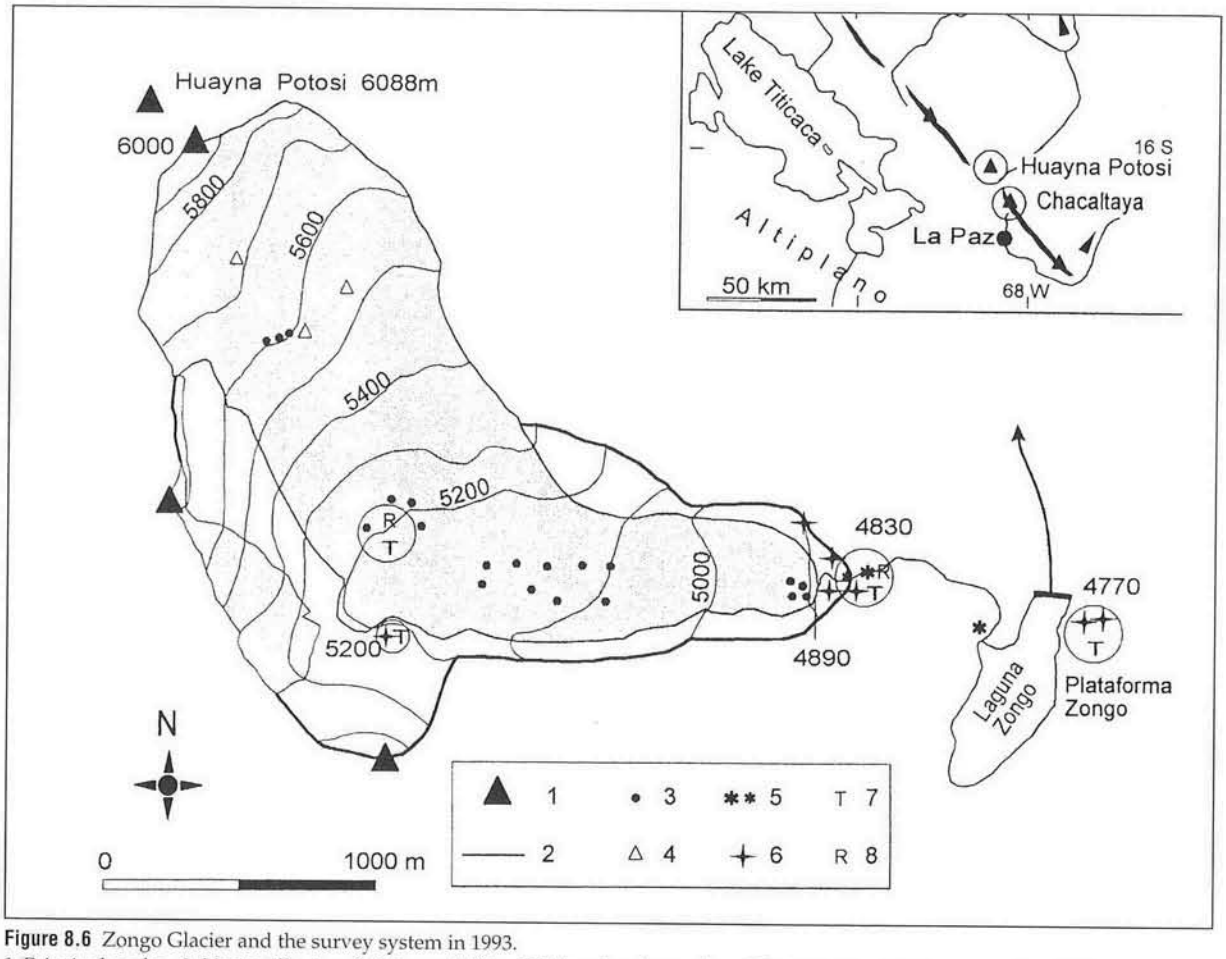


Figure 8.6 Zongo Glacier and the survey system in 1993.
1. Principal peaks – 2. Limits of basin – 3. Stakes – 4. Pits – 5. Water-level records – 6. Rain gauges – 7. Thermographs – 8. Pyranometers.

slopes present the greatest contrasts. This explains a difference of approximately 300 m in the ELA of glaciers of most contrasting slopes, which are the NE and SW slopes. However, in August 1994 and in July–August 1995, notable ablation rates were measured on Zongo Glacier (Francou and Ribstein, unpublished data).

The highest instantaneous discharges from glacier torrent of Zongo Glacier have been measured in the summer, usually in November-December, at the end of dry periods (Fig. 8.9). These events represent the highest values of ablation registered on a daily basis. The frequency and length of these 'dry periods' in the warm season seem to be the important factors in determining the value of the annual net balance.

Using ice cores taken from Quelccaya Ice Cap (Peru), Thompson *et al.* (1984) demonstrated the effects of the El Niño phenomena (ENSO events) on the accumulation balances of glaciers at high altitudes; every ENSO event is marked in the Central Andes by a reduction in accumulation rates. This result tallies with the significant decrease in precipitation during the ENSO episodes (Francou and Pizarro, 1985). During the ENSO event of 1991–1992, monthly measurements of balance taken from the Zongo and Chacaltaya Glaciers also show that the ablation rate during these events is significantly high, even at great altitudes (Fig. 8.8 and 8.9).

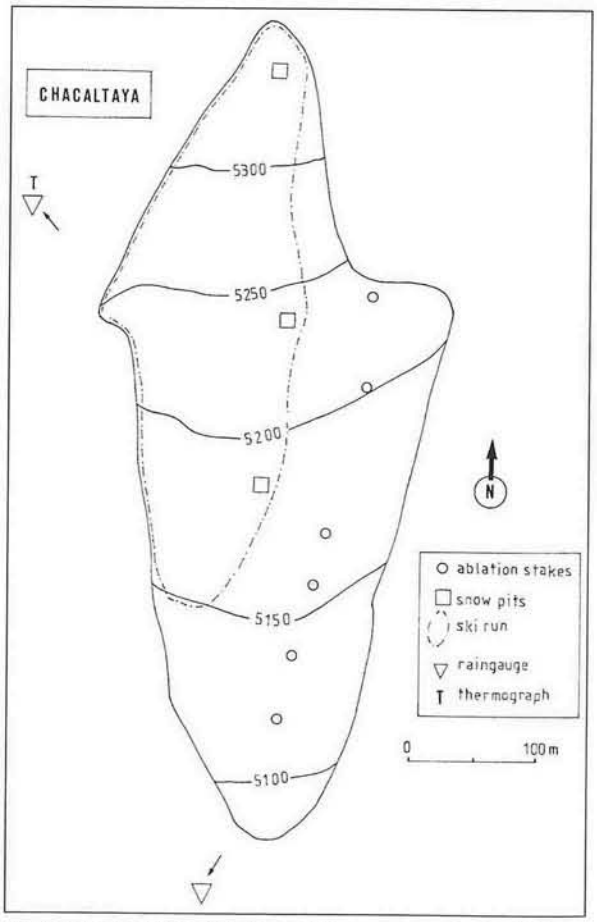


Figure 8.7 Chacaltaya Glacier and the survey system in 1993.

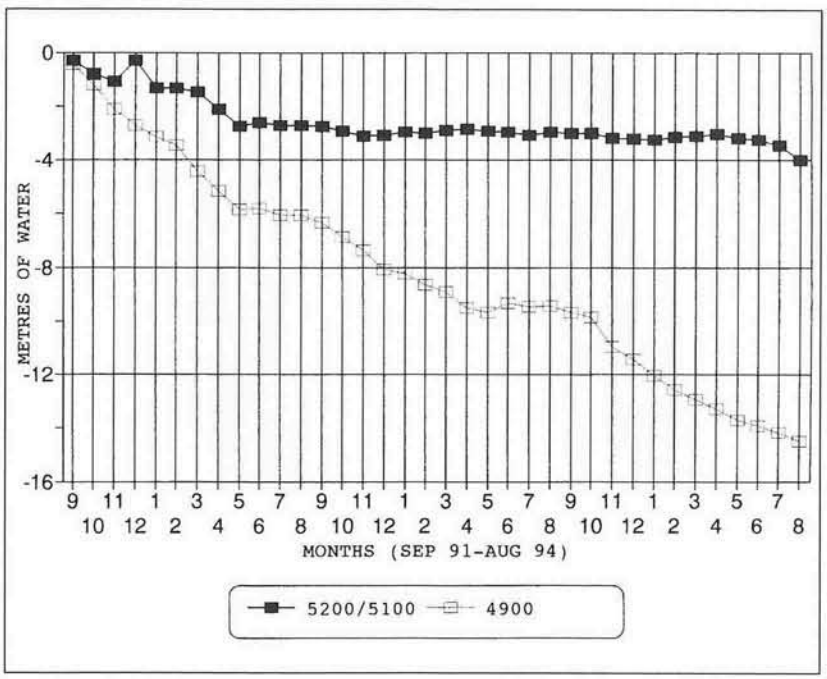


Figure 8.8 Cumulative net balance for the 36 months from September 1991 to August 1994 at Zongo Glacier: 5,200-5,100 m elevation range (black rectangles); 4,900 m (white rectangles).

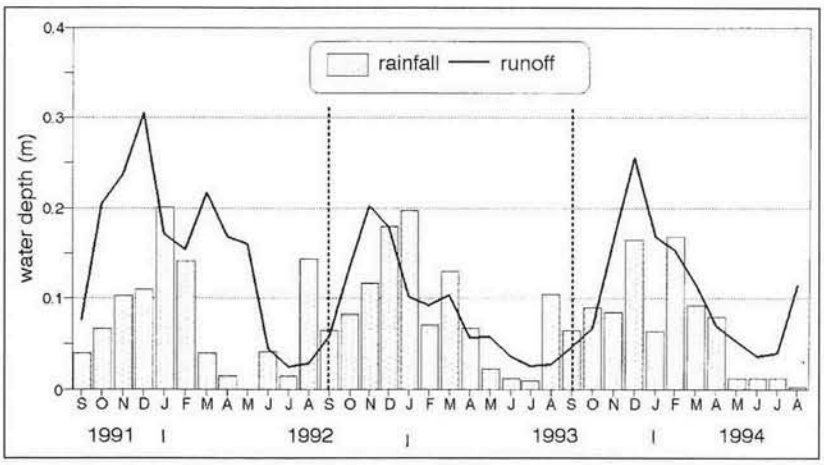


Figure 8.9 Monthly precipitation (average from 4 rain gauges) and runoff on Zongo Glacier.

Therefore, both terms of the balance, accumulation and ablation are influenced by the phenomena. The ENSO effect can be appreciated on Zongo Glacier by comparing the values obtained for specific net balance, precipitation, specific ablation, specific runoff, the Equilibrium Line Altitude and the Accumulation Area ratio for 1991–92, 1992–93 and 1993–94 (Table 8.6). It is noteworthy that 1992–93 and 1993–94 are not ENSO years: the specific net balance was equilibrated during the first period and negative during the second.

A comparison of the three years reveals that the ablation values are significantly different, while the accumulation values, given by the amount of precipitation, are more uniform. More than by a reduced amount in total precipitation, the ENSO year 1991–92 has been marked by 1) a rainy season shorter than normal – at Plataforma de Zongo (4,770 m), only 4 months received more than 50 mm instead of the usual 7 months during a normal period and 2) a reduction in cloud cover during the warm season, which resulted in high average radiation values linked with above-average maximum temperatures (values between 1 and 2 standard deviation). The short rainy season, restricted to January-February, resulted in two periods of strong ablation, a first one in November-December, stronger than normal, and a second one, more attenuated, in March. These two peaks are well reflected in the runoff values of the glacier torrent (Fig. 8.9).

Table 8.6 also shows that the ablation (Ag) and runoff (Q) values for the three years are similar, which implies that the sublimation rate would be relatively weak on this type of glacier. It is, however, unquantifiable, given the imprecision of the measurement.

The balance of Zongo Glacier has been reconstructed using measurements taken at the limnimetric station in 1991–93 and using readings taken every day since 1973 (that is, over 20 years) in a canal that recollects the waters of the glacier (Ribstein *et al.*, 1995). Fig. 8.10 shows that, for every ENSO event, given by a significant negative value of the South Oscillation Index, there is a correspondingly high negative balance value with a possible delay of a few months. This suggests that, in this part of the Andes, ENSO events directly control variations in glacier balances. ENSO events explain 4 periods with very negative balances: 1977/78, 1982/83, 1987/88 and

TABLE 8.6 Zongo Glacier (1991–1993): net balance, ablation, runoff, precipitation, ELA and AAR

Year	*Bn (1)*	*P (2)*	*Ag (3)*	*Q (4)*	*ELA (5)*	*AAR (6)*
1991–92	–1.38	0.92	2.30	2.25	5,300	58
1992–93	0.02	1.06	1.04	1.18	5,100	86
1993–94	–0.73	0.85	1.58	1.56	5,200	67

(1) specific net balance (m of water)
(2) precipitation measured near the glacier (4,800–5,200 m) (m of water)
(3) specific ablation: Ag = P – Bn (m of water)
(4) specific runoff (surface of the glacier: 2.1 km²) (m of water)
(5) Equilibrium Line Altitude (in m)
(6) Accumulation Area Ratio (in %)

1991/92 (only the negative balance of 1979–80 does not correspond to an ENSO event).

Subtracting the estimated average amount of rainfall received by the glacier every year (i.e., 1.062 m of water) from the average runoff value (i.e., 1.472 m of water) gives an average deficit balance of 0.41 m apparent during this period of 20 years for Zongo Glacier. Sublimation seems to be low but, if it were taken into account, this deficit would be even more negative. Over these 20 years, only three have had net positive balances, 1974–75, 1975–76 and 1986–87. For the same three years, the termini of the small glaciers being monitored in the Cordillera Blanca have not retreated, indicating that their balances have been positive or in equilibrium (Fig. 8.5).

This study suggests that the present retreat of Bolivian glaciers, such as those of Cordillera Blanca (Francou *et al.*, in press), is strongly influenced by ENSO events. In between these events, periods of positive or in-equilibrium balances may occur but their duration is not such as to reverse the tendency for these glaciers to retreat. At the most, they can slow down the process but only for a short time.

An accelerated glacier retreat has also been reported from the Andes de Mérida in the northwestern part of Venezuela (Schubert, 1992; 1993).

8.5 SPECIAL EVENTS

Some glaciers in the Central Andes of Argentina have experienced rapid advances over several hundred metres and are considered surging glaciers. Examples include the glaciers in the Mendoza basin. Horcones Inferior Glacier on the southern flank of Mt. Aconcagua surged in 1985 (IAHS(ICSI)/UNEP/UNESCO, 1993; Llorens and Leiva, 1994). In the Río del Plomo basin at 33°S, Grande del Juncal Glacier surged in 1910 and at some time between 1934 and 1955. Grande del Nevado Glacier surged at the end of 1933, between 1963 and 1974, and again in 1984. It dammed up a lake in January 1985 (Helbling, 1919; 1935; 1940; Espizua, 1986; 1987; Espizua and Bengochea, 1990; Bruce *et al.*, 1987; Prieto, 1986; IAHS(ICSI)/UNEP/UNESCO, 1988; 1993). The Glaciar Innominate advanced 2,900 m between 1986 and 1991 (Llorens and Leiva, in press). In the Río Tupungato basin, an unnamed glacier to the south of Tupungato Glacier known as Glacier B has advanced 1,100 m since 1985/1986, the position of the front remaining unchanged between 1991 and 1994. From the Mesón San Juan ice plateau, a glacier tongue advanced 750 m from 1985 to 1986 but did not present the characteristic chaotic surface (low reflectance through analysis of Landsat image) found on other surging glaciers (Llorens and Leiva, 1994).

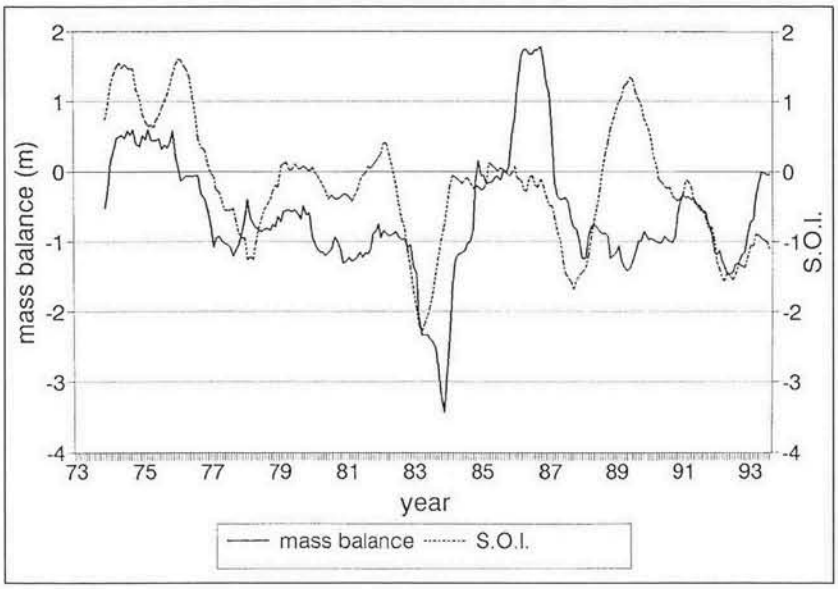


Figure 8.10 Mass balance of Zongo Glacier vs. South Oscillation Index (SOI). Mass balance is reconstructed using hydrological data. SOI is the difference in pressure between Tahiti and Darwin in standardized values. Both curves represent 12-month running means.

TABLE 8.7 Catastrophes produced by glaciers in Peru. All events are from the Cordillera Blanca unless otherwise noted. Where no literature reference is given, the source is personal records by Ames.
Abbreviations: IA = ice avalanche; LOF = lake outflow

Date	*Origin of catastrophe (mountain, lake, etc.)*	*Place of greatest damage (e.g., town)*	*Type of catastrophe*	*Approx. number of victims*	*Reference*
March 4, 1702	?	Huaraz	flood	?	Alba Herrera, 1969
January 6, 1727	Nevado Huandoy	Town of Ancash	earthquake, probably IA, flood	1,500	Alba Herrera, 1969; Silgado, 1978
January 6, 1727	?	Huaraz	earthquake, perhaps flood caused by IA	more than 1000	Alba Herrera, 1969; Silgado, 1978
February 27, 1869	Cerro San Cristobal	Monterrey (near Huaraz)	LOF, flood	11	Alba Herrera, 1969
June 24, 1883	Lake Tambillo (Rajucolta)	Macashca	LOF caused by IA	'many'	Alba Herrera, 1969
January 22, 1917	Nevado Huascaran	Villages Shacsha and Ranrahirca	IA	?	Alba Herrera, 1969
March 14, 1932	Lake Solteracocha (Cordillera Huayhuash)	planting fields in the Quebrada Pacllon	LOF	?	Alba Herrera, 1969; Kinzl, 1940*c*
January 20, 1938	Lake Artesa (Pakliashcocha)	bridge destroyed near Carhuaz	LOF	none	Kinzl, 1940*c*
April 20, 1941	Lake Suerococha (Cordillera Huayhuash)	planting fields destroyed	LOF	none	Alba Herrera, 1969
December 13, 1941	Lakes Acoshcocha and Jircacocha	Huaraz	LOF	4,000	Lowther and Giesecke, 1942; Track, 1953; Buse, 1957; Fernandez, 1957; Oppenheim, 1947; Heim, 1948
January 17, 1945	Lakes Ayhuiñaraju and Carhuacocha	Chavin de Huantar	LOF caused by rock slide from Nev. Huantsan	300	Indacochea and Iberico, 1947; Spann, 1946
October 20, 1950	Lake Jankarurish	Hydroelectric power station of Huallanca	LOF, possibly caused by IA	200, perhaps 500	Ghiglino, 1950
June 16, 1951	Lake Artesoncocha	no damage, water absorbed by Laguna Paron	LOF (first, see below)	none	Fernandez, 1957
October 28, 1951	Lake Artesoncocha	as above	LOF (second, see above)	none	Fernandez, 1957
November 6, 1952	Lake Milluacocha	minor damage in planting fields	LOF	none	
? 1953	Lake Tullparaju	no damage downvalley	landslide, waves erode artificially modified lake outlet	none	
December 8, 1959	Lake Tullparaju	minor damage in planting fields	as above	none	
January 10, 1962	Nevado Huascaran	Ranrahirca and 9 smaller villages	IA	4,000	Dollfus and Peñaherrera, 1962; Patzelt, 1985
December 22, 1965	Lake Tumarina	Quebrada Carhuascancha	LOF, probably caused by IA	several	
May 31, 1970	Nevado Huascaran Norte (west face)	Yungay, Ranrahirca and Matacoto	earthquake causes large rock and IA	20,000	Lliboutry, 1971, Plafker *et al.*, 1971; Patzelt 1985; Welsch, 1970
May 31, 1970	Huascaran Norte (north face)	Quebrada Llanganuco	earthquake causes rock and IA	14	
August 31, 1982	Lake Milluacocha	bridges and planting fields destroyed	lake outburst caused by IA	none	
December 16, 1987	Nevado Huascaran	road blocked	IA	none	
January 20, 1989	Nevado Huascaran	road and bridge damaged, planting fields destroyed	IA	none	

At 34°S, Laguna Glacier in the Río Atuel basin advanced 1,400 m between 1970 and 1982 (Cobos and Boninsegna, 1983). In the Southern Patagonia Ice Field, the position of Moreno Glacier has been studied in detail owing to the cyclical damming and periodic catastrophic drainage of the southern arm of Lago Argentino.

In historic times, most of the catastrophic floods and ice avalanches in Peru were recorded in the Cordillera Blanca and Huayhuash. This is only partially due to the relatively large extent of the ice masses. Perhaps equally important is the unusually high population density, in particular in the Santa Valley on the west side of Cordillera Blanca. Perhaps nowhere else in tropical high mountain ranges is there such an intense interaction between glaciers and man. Subsistence agriculture in the mountain areas of Peru is in part dependent on glacial runoff for irrigation (some people even make a living from transporting blocks of glacier ice down to local markets where it is used to produce ice cream and to cool drinks). Therefore, a significant part of the population lives within reach of glacial floods and, in some cases, even within the runout distances of ice avalanches. In other Peruvian Cordilleras, catastrophic events are noted less frequently but do occur. The origins are diverse. Lake outbursts have various kinds of triggering mechanisms. However, most of them have in common that the lakes initially formed as a result of the general retreat of the glaciers (Oppenheim, 1947; Heim, 1948; Track, 1953; Fernandez, 1957; Ames *et al.*, 1994). Typically, they are dammed by poorly consolidated morainic material and perhaps sometimes even by buried stagnant ice. Extreme rainfall may then weaken the dam until it fails spontaneously. In many cases, landslides and ice avalanches have fallen into the lakes, causing waves which erode the lake outlet. Exponential increase of discharge then leads to the sudden drainage. The so-called «aluvion», a turbulent mixture of rock, finer sediment and water then rushes downvalley causing destruction and often the loss of many lives. Table 8.7 summarizes all known major events (including ice avalanches) as far back as the beginning of the 18th century. Some lakes were made less dangerous by lowering the water level and strengthening the outflow. Successful examples of such work are Laguna Safuna (Lliboutry, 1977; Lliboutry *et al.*, 1977*a*; 1977*b*) or Laguna Llaca above Huaraz. A major problem during projects of this kind was the steep topography. Usually, the first task was the construction of a road in steep and difficult terrain from the Santa valley to the construction site. Not all lake modifications were without problems. On October 20, 1950, Lake Jankarurish in the Quebrada Alpamayo produced an aluvion which completely destroyed the installations of the Huallanca hydroelectric power plant under construction at the time on the lower Santa River. A tunnel was filled with debris and the main bridge and three railway bridges were washed away. Officially, 200 people were reported dead but other estimates give figures of 500. At the time of the flood, a group of workers employed by the official institution in charge of preventing floods was lowering the lake level. The immediate cause of the disaster was, most probably, a large wave triggered by an ice avalanche from a glacier on the western slopes of Nevado Alpamayo, causing progressive erosion of the overflow channel. The outburst volume was estimated at some 4 million m^3, increasing downvalley as material from the riverbed and sides got carried along. The material rushed towards Santa River at a speed of about 30 km/h (Ghiglino, 1950).

8.6 GAPS AND NEEDS

A basic task to be undertaken is the completion of the glacier inventory in Chile and Argentina, as much remains unknown. There is a need to know the mass balance of more glaciers along the Andes of Chile and Argentina, not only near Santiago and Mendoza as is presently the case, but also in other parts of the country.

Continuity in the inventorying of glaciers and in glacier mass balance studies has to be established. In addition, an interconnected research cooperative programme is needed to ensure the continuity and enhancement of present glaciological research. Training courses are another must.

8.7 SUGGESTED FUTURE DEVELOPMENT OF MONITORING ACTIVITY

The observed strong glacier retreat in the Peruvian Andes, as well as in other parts of South America (e.g., Central Andes of Argentina since the beginning of the century), is considered an impressive example of the importance of glacier fluctuation measurements, particularly in tropical or subtropical areas. In remote regions or areas with few or no systematic climatological data records, such measurements provide a highly sensitive indicator of recent or subrecent climatic changes. Continued glacier retreat is expected to further affect the local population: new terminal lakes will probably form, thereby creating new potential sources of catastrophic flooding. A further loss in ice masses will, at the same time, temporarily add to the amount of water available for agriculture and hydroelectric power production. In the long run, however, runoff must decrease. Efforts should be made to continue taking these valuable measurements, particularly in view of a potentially persistent trend of atmospheric warming. However, the very limited financial means, coupled with the small number of personnel trained to carry out this work, poses a serious problem.

The marked sensitivity of tropical glaciers to climatic changes means that they are particularly well suited as indicators in the current research on global

warming. The high resolution with which the glacier transmits this information makes it a unique instrument with few equivalents in the tropical continental environment. To follow these processes, it is essential to establish a long-term network of studied glaciers in the Central Andes. With this network as a starting point, research should be directed towards the following:

1) A better understanding of the functioning of tropical glaciers must be developed by studying their balances at shorter time intervals (days, months): energy balance, glacial balance, hydrological balance.
2) The possible effects on tropical glaciers of an increase in temperature, as estimated by global circulation models, must be analysed.
3) The impact of ENSO events according to the latitude (Equator–Tropics) and the areas of climatic influence in the Cordillera (Amazon and Pacific) must be quantified.
4) A rapid and broad glacial retreat for the high Andean catchments must be analysed from the viewpoint of potential consequences for the hydrological regimes and the effects on water resources of a possible increase in the risk of glacial hazards (avalanches, overflow of glacial lakes).

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9 Glaciers in Europe

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9.1 INTRODUCTION

The presently glacierized area in Europe of some 54,000 km^2 mainly occurs in Svalbard and Iceland (IAHS(ICSI)/UNEP/UNESCO, 1989). Smaller but very well-documented ice masses exist at higher altitudes in Scandinavia and the European Alps. The following text concentrates on Svalbard and Scandinavia but also refers to the Italian Alps as an example of low-latitude/high-altitude glaciers in central Europe. Additional information on Alpine glaciers can be found in Haeberli (1996). Brief comments on the (Spanish) Pyrenees are reserved for the isolated glaciers and ice patches found in the Mediterranean mountains.

9.2 GLACIER DISTRIBUTION AND CHARACTERISTICS

The main sources of information from mainland Norway, Svalbard and Sweden (Fig. 9.1) are different glacier inventories: 1) *Atlas over breer i Sør-Norge* (Atlas of Glaciers in South Norway) by Østrem *et al.*, (1988); 2) *Satellite Image Atlas of Glaciers of the World* edited by Williams and Ferrigno (1993) in which there are three chapters: Glaciers of Norway by Østrem and Haakensen, Glaciers of Sweden by Schytt and Glaciers of Svalbard by Liestøl; 3) *Glacier Atlas of Svalbard and Jan Mayen* by Hagen *et al.* (1993*a*). In the following text, Norway, Sweden and Svalbard will be described separately.

The total area covered by glaciers in Norway (Fig. 9.1) is greater than for most other countries in Europe but still less than 1% of Norway's total area. According to glacier inventories compiled in 1969 and 1988 in southern Norway and in 1973 in northern Norway, 2,595 km^2 of land is covered by glaciers (Østrem and Ziegler, 1969; Østrem *et al.*, 1973; 1988). The largest ice cap in Norway, Jostedalsbreen, covers 487 km^2 and is the largest continuous ice mass in continental Europe. Only Iceland and Svalbard have larger glaciers.

Glaciers occur as ice caps, outlet glaciers, cirque glaciers and small valley glaciers. The glaciers in

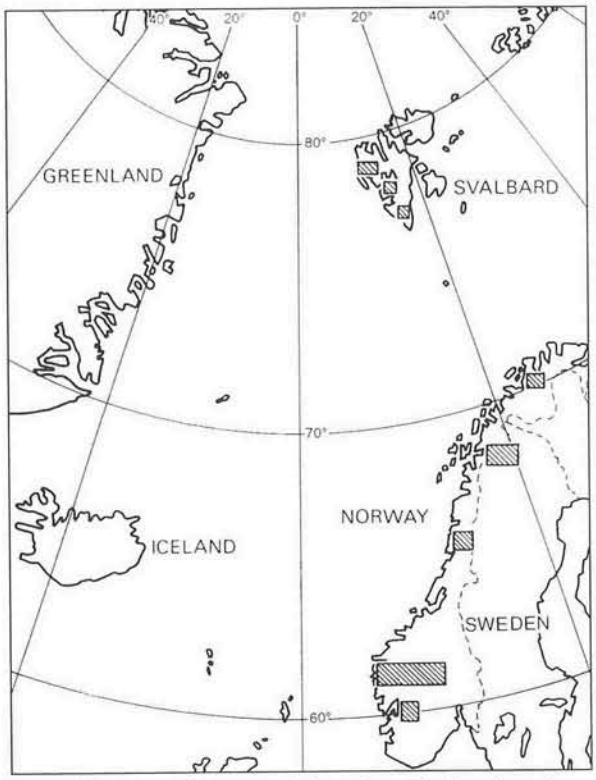

Figure 9.1 Areas where mass balance investigations have been conducted on glaciers in Scandinavia and Svalbard.

Norway are concentrated in two areas: 1) The glaciers in the central and western part of southern Norway occur in the high-mountain areas, with the dominant ice caps being Jostedalsbreen, Hardangerjøkulen and Folgefonni in the west. Numerous valley and cirque glaciers exist in the Jotunheimen mountain area further east. 2) Most of the glaciers in northern Norway are concentrated in Nordland County – the narrowest part of Norway – between approximately 66°N and 68°N. However, a number of valley glaciers and small ice caps are also distributed farther north, particularly on the Lyngen peninsula east of Tromsø. The northernmost glacier in continental Europe is found on the island of Seiland at 70°25′ N.

On the western coast of southern Norway, the critical glaciation level is about 1,200 m a.s.l., which means that mountains higher than this level normally carry glaciers. In the eastern part of Jotunheimen, the glaciation level is about 2,200 m; consequently, all glaciers there are situated at much higher elevations than those near the coast. The equilibrium line altitude (ELA) on the westernmost glacier in South Norway, Ålfotbreen, is about 1,000 m a.s.l., gradually increasing to about 1,550 m a.s.l. on Jostedalsbreen and 2,100 m a.s.l. on Gråsubreen in the eastern part of Jotunheimen. Similar conditions prevail in northern Norway, where glaciers form at considerably lower elevations in the coastal districts than in the more continental inland part of the Scandinavian peninsula. The lowest ELA has been observed on Langfjordjøkulen where the ELA is about 700 m a.s.l.

More detailed information about the length of glaciers, their orientation and the date of aerial photography, etc. can be found in the two glacier atlases of Norway (Østrem and Ziegler, 1969; Østrem *et al.*, 1973; 1988; Østrem and Haakensen, 1993). The largest glaciers are listed in Table 9.1. The glacier atlases also contain considerable geomorphological information, such as surface characteristics, morainal features, proglacial lakes and other data related to each glacier unit listed.

The most recent glacier inventory of Sweden (Østrem *et al.*, 1973) contains 294 glaciers, with a total area of 314 km^2. Thus, there are numerous small glaciers. The Swedish high mountains are situated in a much more continental type of climate than most of the Norwegian mountains. Therefore, only a few massifs rise above the present-day glaciation level and, where they do, the areas are too small to support any extremely large glaciers. Most of the Swedish glaciers are small cirque and valley type glaciers with well-defined area distribution. The largest of thse are listed in Table 9.2. Glaciers are concentrated in the Kebnekaise and Sarek mountain massifs in northern Sweden close to the Norwegian border. Sweden's largest glacier is Stourrajekna, with an area of 12.7 km^2. Swedish glaciers are all situated far from populated areas and, hence, did not attract much attention until the 1800s.

TABLE 9. 1 The size and location of the largest glaciers in Norway

Glacier name	*Area (km^2)*	*Height (m)*		*Location*	
		maximum	*minimum*	*Long. E*	*Lat. N*
1. Jostedalsbreen	487	2,000	350	7° 00′	61° 40′
2. Vestre Svartisen	221	1,580	20	14° 00′	66° 40′
3. Søndre Folgefonni	168	1,660	490	6° 20′	60° 00′
4. Østre Svartisen	148	1,550	208	14° 10′	66° 40′
5. Blåmannsisen	87	1,560	810	16° 00′	67° 20′
6. Hardangerjøkulen	73	1,850	1,050	7° 20′	60° 30′
7. Myklebustbreen (Snønipbreen)	50	1,830	890	6° 40′	61° 40′
8. Okstindbreen. . .	46	1,740	750	14° 10′	66° 00′
9. Øksfjordjøkelen	41	1,170	330	22° 00′	70° 10′
10. Harbardsbreen	36	1,950	1,250	7° 40′	61° 40′

Source: Østrem *et al.*, 1988

TABLE 9. 2 Area of the largest glaciers in Sweden

Glacier name	*Area (km²)*
Stourrajekna	12. 75
Almaljekna	12. 15
Salajekna	11. 10
Partejekna	11. 10
Jakatjkaskajekna	9. 96
Suottasjekna	8. 11
Mikkajekna	7. 62
Å. Ålkatj-jekna	6. 56
Å. Ruotesjekna	5. 41
Riukojietna	4. 99

Source: Østrem *et al.*, 1973

TABLE 9. 3 Area of the largest ice caps and ice fields in Svalbard

Spitsbergen	*Area (km²)*
Olav V Land Ice Field	4,150
Holtedalfonna	1,375
Åsgårdsfonna	1,230
Lomonosovfonna	600
Isachsenfonna	505
Balderfonna	345
Fimbulisen	320
Sørkappfonna	265
Løvenskioldfonna	265
Nordmannsfonna	250
Filchnerfonna	203
Ursafonna	125
Hellefonna	122
Nordaustlandet	
Austfonna m/Vegafonna	8,450
Vestfonna	2,455
Glitnefonna	174
Edgeøya	
Edgeøyjøkulen	1,365
Digerfonna	270
Storskavelen	190
Kvitisen/Langjøkulen	100
Kvalpyntfonna	85
Kvitkåpa	80
Barentsøya	
Barentsjøkulen	570
Kvitøva	
Kvitøyjøkulen	705

Source: Hagen *et al.*, 1993*a*

The total glacierized area of Svalbard is 36,600 km² of a total land area of 61,700 km². Thus, about 60% of Svalbard is covered by glaciers of various types. Large areas are characterized by great continuous ice masses divided into individual ice streams by mountain ridges and nunatak areas. This glacier type was called Spitsbergen by Ahlmann (1933). Small cirque glaciers are also numerous, especially in the high alpine mountain regions in the western parts of Spitsbergen. Several large ice caps are located in the relatively flat areas of eastern Spitsbergen, Edgeøya, Barentsøya and Nordaustlandet. Some typical piedmont glaciers are found along the west coast, resting on the strandflat of Prince Karls Forland. Ice shelves do not exist because all glacier fronts terminating in the sea are grounded. The majority of the glaciers belong to the subpolar type. The margins and parts of the ablation area are below freezing point and partly frozen to the ground, while the accumulation area and deeper parts of the ablation area are at the pressure melting point. Many of the small cirque glaciers could be classified as polar glaciers because the entire ice mass is below melting point. The largest ice caps, ice fields, outlet glaciers and ice streams in Svalbard are listed in Table 9.3 and Table 9.4.

TABLE 9. 4 Area of the largest outlet glaciers and ice streams in Svalbard

Name	*Area (km²)*
Hinlopenbreen	1,250
Negribreen	1,180
Bråsvellbreen	1,110
Leighbreen	715
Stonebreen	710
Kronebreen	690
Etonbreen	665
Hochstätterbreen	580
Nathorstbreen	490
Monacobreen	410

Source: Hagen *et al.*, 1993*a*

The Italian side contains about 20% of the total glacierized surface area in the European Alps, with a total of nearly 1,400 glaciers (Table 9.5). From the hydrographic viewpoint, these are mainly concentrated in the basins of the Dora Baltea (Po, 296 glaciers covering 193 km²), the Adige (412 glaciers; 188 km²) and the Adda (Po, 299 glaciers; 126 km²). The mountain groups with the largest glacier surfaces (Fig. 9.2) are Ortles-Cevedale (101 km²), Monte Rosa (48.2 km²), Monte Bianco (46.1 km²), Gran Paradiso (41.8 km²) and Adamello-Presanella (41.6 km²). It should also be recalled that the eastern slope of the Gran Sasso d'Italia (Vomano basin in the Abruzzo Appennines) contains the Calderone Glacier, the southernmost glacier of Europe (42°28′15′ N), now reduced to a surface area of a few hectares.

The valleys of the central Spanish Pyrenees, which still contain active glaciers, belong, from west to east, to the basins of the rivers Gallego, Cinca, Esera-Garona and Noguera Ribagorzana. The peaks of the mountain ranges home to these glaciers are often more than 3,000 m high but their different shapes, locations, orientations and height above sea level give rise to a different number of glaciers in each massif.

TABLE 9. 5 Major glacial systems of the Italian Alps

Name	*Mountain group*	*Surface area (km²)*	*max. length (km)*	*snout altitude (m a. s. l.)*
Forni	Ortles-Cevedale	13. 20	5. 50	2,350
Miage	Monte Bianco	13. 20	10. 35	1,720
Mandrone	Adamello-Presanella	12. 38	5. 38	2,450
Lys	Monte Rosa	11. 80	5. 36	2,350
Rutor	Rutor	9. 54	4. 80	2,504
Malavalle	Breonie	9. 42	4. 34	2,460
Brenva	Monte Bianco	8. 06	7. 64	1,415

Source: Serandrei *et al.*, 1993

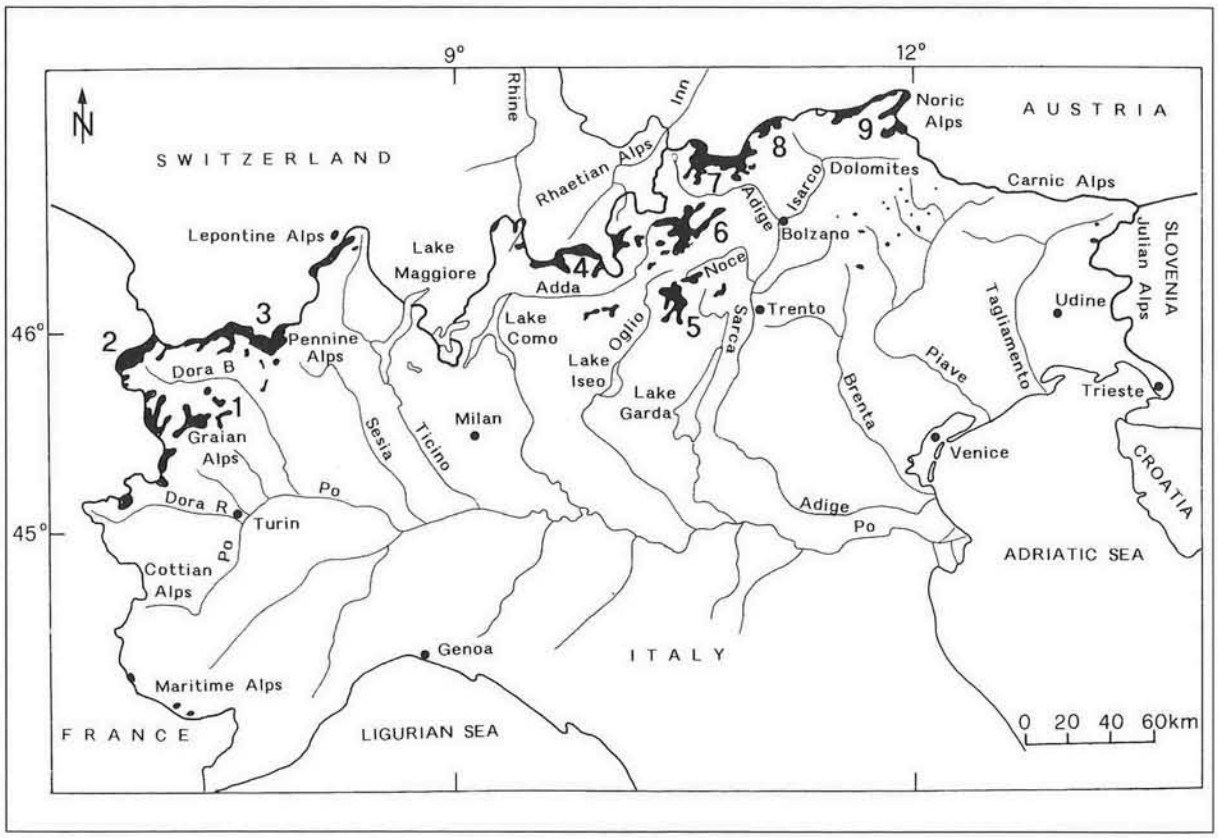

Figure 9.2 Sketch map of the drainage basins and glacier distribution in the Italian Alps. The numbers refer to the main glaciated mountain areas listed below (modified from Serandrei *et al.*, 1993):
1 Gran Paradiso **2** Monte Bianco **3** Monte Rosa **4** Bernina **5** Adamello-Presanella
6 Ortles-Cevedale **7** Alpi Venoste / Ötztaler Alpen **8** Alpi Breonie / Stubaier Alpen **9** Alpi Aurine / Zillertaler Alpen

9.3 EXISTING GLACIER INVENTORIES

Inventories of the glaciers in Norway have been made at various intervals. The first truly complete inventory was made in 1958, entitled *Glaciers and Snowfields in Norway* (Liestøl, 1962*a*). This inventory overestimated the number and area of glaciers, mainly due to the fact that both glaciers and snowfields were included. The topographic maps used as source materials did not distinguish cartographically between true glaciers and snowfields. Another, less detailed inventory prepared by Østrem (1963) contained an index map on a scale of 1:1,800,000 showing all glaciers in Scandinavia. An improved version of the southern part of this map was completed in 1963, when a glacier map of southern Norway was issued on a scale of 1:500,000 and printed in three colours (Liestøl and Østrem, 1963). Detailed glacier inventories were prepared of southern Norway in the late 1960s (Østrem and Ziegler, 1969) and of northern Norway and Sweden in the early 1970s (Østrem *et al.*, 1973). A second, revised glacier inventory for southern Norway was completed in the late 1980s (Østrem *et al.*, 1988).

In the compilation of these inventories, which were made in accordance with criteria recommended by UNESCO (1970), all glacier masses were divided according to a system based on hydrological factors. This system was chosen because the information obtained would be used mainly for hydrological pur-

poses. It was also decided that glaciers draining into different rivers should be subdivided into separate 'glacier units,' each unit draining into one river. The total number of these glacier units amounts to 2,113, whereas the number of 'glaciers,' continuous masses of ice regarded as one glacier body, amounts to 1,627, covering a total area of 2,595 km^2. High-quality aerial photographs are available for almost all glacierized areas in Norway and 31 modern glacier maps have been published on scales of 1:10,000 to 1:50,000. Landsat images have limited value in Norway for most glaciological studies but the satellite data have been used to qualitatively evaluate suspended sediments in lakes and fjords and to monitor the transient snow-line as an indication of the approximate net mass balance.

In *Die Gletscher Schwedens im Jahre 1908*, Axel Hamberg published a map of the Swedish glaciers (Svenonius *et al.*, 1910) but it was the areal distribution rather than the characteristics of the individual glaciers that was of prime importance in this work. At that time, it was next to impossible to prepare a good glacier map because the National Land Survey maps of the Lappish mountains had all been compiled around 1890 and published on a scale of 1:200,000. It is not surprising that many glaciers were not shown and that many snowfields were mapped as glaciers.

The next attempt to map the glaciers was made by Fredrik Enquist in 1918 when he made a study of the height of the glaciation level (Enquist, 1918). However, this map shows all glacierized mountainous areas in Sweden but not the individual glaciers themselves. The poor-quality topographic maps available made good glacier inventories more or less impossible. A few glaciers were mapped quite well during the early half of the century (Karsajøkeln in 1925 and 1943; Storglaciären in 1910, 1922, 1946 and 1949) but the topographic maps could not be adjusted to accept this detailed information. In 1958, however, the entire mountainous part of western Sweden was photographed on a scale of 1:65,000. These aerial photographs were used for preparing the first detailed compilation of a glacier inventory of Sweden (Schytt, 1959). However, the maps were based on the older topographic map sheets and could not possibly attain a very high degree of accuracy.

A modern inventory was finished in 1973, when the *Atlas over breer i Nord-Skandinavia* (Atlas of Glaciers in Northern Scandinavia) was published by Østrem *et al.* (1973). For the Swedish part of this inventory, a completely new set of 1:100,000 scale photogrammetric maps of fine quality were used. Most of the glaciers were already well-portrayed on these maps. The photographic coverage was also far better, both in terms of quality and scale, than that available in the late 1950s. Schytt's map (1959) was published on a scale of 1:600,000 and the glacier atlas contains maps with scales of 1:1,500,000, 1:600,000, 1:500,000 and 1:250,000. The glacier atlas and the ordinary topographic maps together provide a sound record of the shape and extent of Swedish glaciers during the 1960s. Besides the small-scale maps of the regional glacier inventory, several Swedish glaciers have also been mapped on larger scales for scientific purposes. Storglaciären was mapped by terrestrial photogrammetry from photographs acquired in 1910 and 1949 and by aerial photogrammetry from photographs acquired in 1959, 1969, 1980 and 1990. Bed topography was mapped by radio-echosounding methods in 1979 (Bjørnsson, 1981) and improved by Eriksson *et al.* (1993). Since 1960, the aerial photographic archives of Sweden's glaciers have expanded rapidly. Aerial photographs on a scale of 1:30,000 are now available for nearly all glaciers and many special aerial surveys of selected glaciers have been made.

In Svalbard, the first systematic inventory work started in 1980 at the Norwegian Polar Institute, Oslo, where all basic material is available. The work was not completed until 1994 after publication of the *Glacier Atlas of Svalbard and Jan Mayen* (Hagen *et al.*, 1993*a*). Much of the textual presentation in this atlas was also published in the *Satellite Image Atlas of Glaciers of the World*, in the chapter Glaciers of Svalbard, Norway (Liestøl, 1993). The inventory was carried out mainly in 1980–81. Since then, many of the small glaciers covering less than 1 km^2 may have melted away. The glaciers in general have had a negative mass balance during this period (cf.below). This is particularly true for glaciers with low mean glacier elevation close to the coast. Thus, many small patches have disintegrated, especially on the flat-lying islands on Kong Karls Land in the east.

The topographical maps used in this inventory in 1980–1981 were in the main a series of maps of Svalbard on a scale of 1:100,000 published by the Norwegian Polar Institute. Many of these maps were subsequently revised and redone, which led to the inventory itself being revised in Edgeøya, Barentsøya and Nordaustlandet in 1989/90. Some of the maps were constructed from aerial photographs dating from 1936 and the glacier areas and front positions had not been updated on the maps. As most glacier areas have been shrinking since the 1930s, aerial photographs on a scale of 1:50,000 taken in 1960, 1966, 1969, 1970 and 1971 – and a few from 1977 – were used in the inventory for updating the glacier extension and for investigating the moraine morphology associated with the glaciers. These aerial photographs are available at the Norwegian Polar Institute. Also used in the compilation are observations of glacier front position on calving glaciers from Landsat satellite images from August 1980. In the summer of 1990, a new aerial photography survey was carried out all over Svalbard. These pictures were not used in the inventory but they would probably confirm the general retreat of the glaciers because of the negative mass balance on most glaciers during the 20th century. The given volumes and areas in the inventory are thus probably slightly overestimated as compared with the 1990 situation.

The first example of systematic, descriptive cataloguing on a proper mapping basis of the Italian glaciers is the geographic work of O. Marinelli enti-

tled *Ighiacciai delle Alpi Venete* (1910), which followed and completed the fundamental work of E. Richter (*Die Gletscher der Ostalpen*) in 1888. The rapidly increasing interest in glaciers and glaciology, linked to hydrological research and exploitation of energy resources, later led to an *Elenco dei ghiacciai italiani* (List of Italian Glaciers (Porro, 1925) describing 774 glaciers, followed soon after by an *Atlante dei ghiacciai italiani* (Atlas of Italian Glaciers), on a scale of 1: 500,000 (Comitato Glaciologico Italiano, 1927).

Between 1959 and 1962, as a contribution to the International Geophysical Year (IGY, 1957–58), the Consiglio Nazionale delle Ricerche (National Council for Research) and the Comitato Glaciologico Italiano (Italian Glaciological Committee) jointly published the *Catasto dei Ghiacciai Italiani* (Inventory of Italian Glaciers). Its four volumes identify, list, describe and represent (scale 1:25,000) a total of 838 existing Italian glaciers, covering about 540 km^2, together with a further 190 glaciers which had disappeared in the 50 preceding years. In spite of the varying provenance of data and inadequate cartographic documentation, this work may really be considered the first organized attempt to create a standard base for the most important elements regarding the glaciers of a substantial region of the globe. Between 1980 and 1986, Italy collaborated with the newly constituted IAHS-UNESCO *World Glacier Inventory* (IAHS(ICSI)/UNEP/UNESCO, 1989) on the complete cataloguing of the glaciers of the Italian Alps. The differences in surface area and number of glaciers with respect to the preceding inventory are mainly due to the fact that the WGI also includes, for Italy, snowfields and glacierets not exceeding 5 hectares, which had not been listed previously. Moreover, the consequences of the glacier expansion in 1965–1985 (cf. below) meant that the glaciated surface area increased and that some of the glaciers that had previously disappeared came back into being.

The number of national and international inventories rose with later regional studies, such as *I ghiacciai del Gruppo Ortles-Cevedale* by A. Desio (1967), Les glaciers des Alpes Occidentales by R. Vivian (1975), *I ghiacciai del Veneto* (Zanon, 1990) and *Ghiacciai in Lombardia* by the CAI-Servizio Glaciologico Lombardo (1992).

Following several descriptions of the Pyrenean glaciers and a first cartography of their positions in the highlands during the 19th century, the first extensive work was published in 1894. This was carried out by the French geographer Franz Schrader (Schrader, 1894) and included details of the characteristics, position and extent of the glaciers. This makes available one hundred-year old maps, texts and figures for appraising the important features of glacier evolution in the Spanish Pyrenees during this period, although posterior data were not regularly obtained. Modern inventories (IAHS(ICSI)/UNEP/UNESCO, 1989) were compiled by D. Serrat (Spanish glaciers), P. Vautier (French glaciers) and by the ERHIN Programme (Martínez de Pisón and Arenillas, 1988).

9.4 LONG-TERM OBSERVATIONS

Since the Little Ice Age, culminating around 1750 in the greatest extent of Norway's glaciers in history, Norwegian glaciers have retreated almost continously and intermittent readvances have been relatively small. During the Little Ice Age, most glaciers had advanced considerably and, in most cases, this has led to the deposition of the outermost moraine at any glacier today. Minor glacier advances occurred during the first part of the 20th century but a major glacier recession started in about 1930. Systematic measurements of the front position of glaciers in Norway began in the late 19th century, although the earliest recorded observation of termini fluctuation was made in 1742 when it was stated that the Tverbre Glacier in the Jostedalen area had advanced and damaged a farm (Hoel and Norvik, 1962, p. 9).

The first written documentation of glacier measurements taken in Norway (published in 1803) was of the outlet glacier Nigardsbreen, which drains southeastward from the Jostedalsbreen Ice Cap. This glacier advanced so far in 1750 that it reached cultivated land and destroyed houses. Since this disastrous advance, the glacier's almost continuous retreat has been interrupted only by relatively small advances. From 1748 until the present time, the retreat amounts to about 5 km. The effect of the small advances can be seen easily in aerial photographs. The series of small morainal ridges has been dated and described in detail by Faegri (1933) and by Andersen and Sollid (1971).

A more systematic measuring of glaciers in Norway started about 100 years ago, when some scientists began to establish survey points in front of selected glaciers for annual position measurements of each terminus. This kind of systematic measurement was first made by the geologist P. A. Øyen, who published his figures in annual reports (Øyen, 1898; 1907; 1915), and by J. B. Rekstad (1905), who also photographed many glaciers. The famous French geographer, Charles Rabot, took many photographs in Scandinavia during the 1880s. Several of his photographs provide considerable information about the position of selected glacier termini at that time. Later, Professor W. Werenskiold would begin a long series of glacier-termini measurements in the Jotunheimen mountains. It is known that minor glacier advances occurred during the first part of the 20th century (Liestøl, 1962*b*) but that a period of major glacier retreat started around 1930.

Since the early 1900s, frontal positions have been measured annually, with some short gaps, on about fifteen glaciers (Bogen *et al.*, 1989; Nesje, 1989). Frontal positions have also been measured on other glaciers over shorter periods. The main period of retreat lasted from about 1930 until the end of the 1960s. Since then, the rate of retreat has declined and, in recent years, advancing glaciers have been recorded. Since 1988, Briksdalsbreen, one of the measured outlets, has advanced 200 m, 75 m of

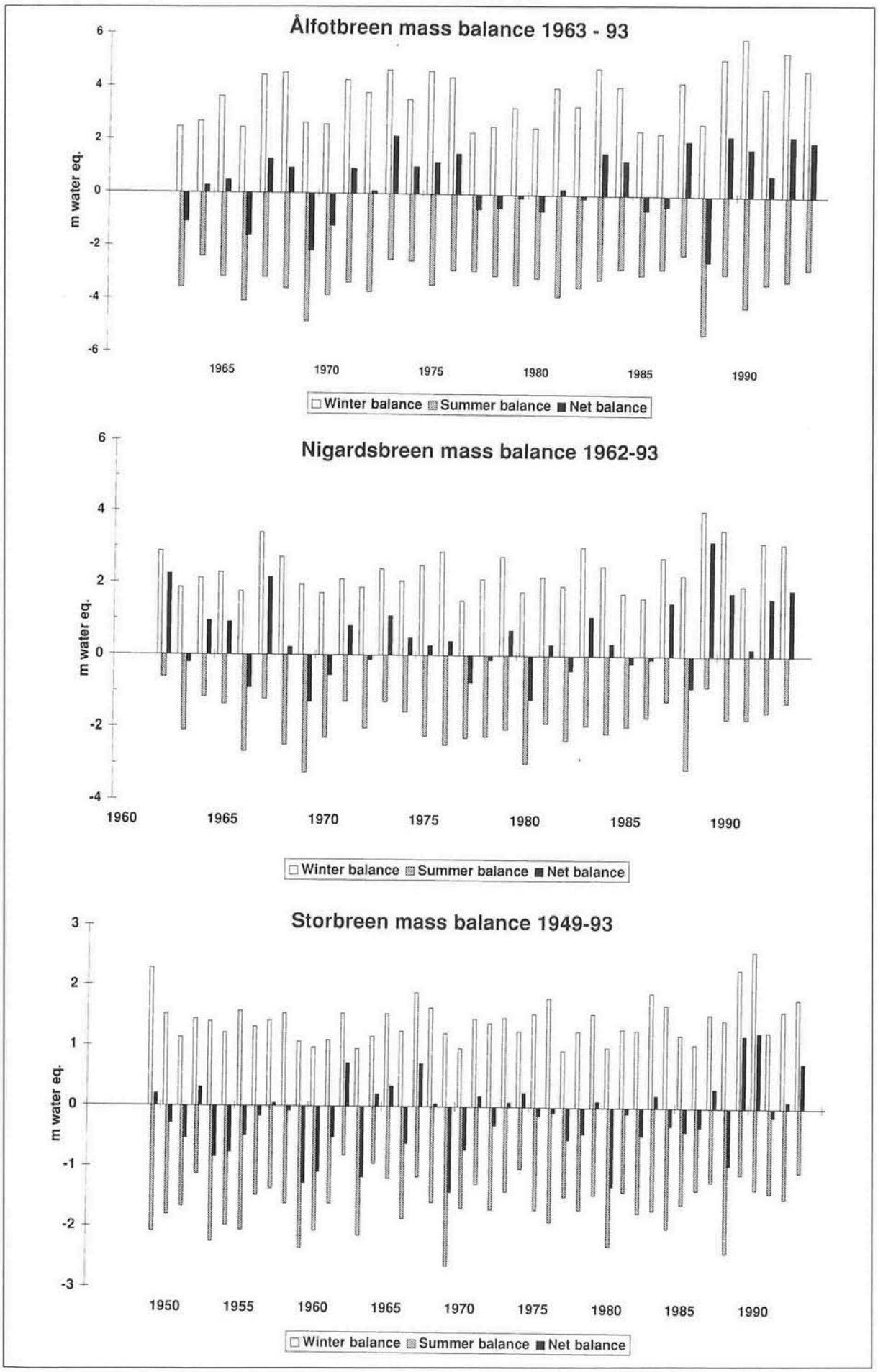

Figure 9.3 Mass balance results from three different glaciers in a transect west-east in southern Norway.

which was due to the advance in 1993. This is the largest annual advance ever measured on a glacier in Norway.

The longest continual mass balance time series in Norway dates from 1948 at Storbreen in Jotunheimen (Liestøl, 1967). The investigations were started by the Norwegian Polar Institute, which measured both winter and summer balance every year up until 1993. The Norwegian Water Resources and Energy Administration (NVE) is a governmental organization which is both an administrative body and responsible for developing all kinds of water and energy resources. The Hydrology Department of the NVE initiated long-term mass balance studies on selected glaciers in southern Norway in 1962 and 1963. Glaciers regarded as representative of certain areas were selected. The Ålfotbreen, Nigardsbreen, Hardangerjøkulen, Hellstugubreen and Gråsubreen Glaciers – in addition to Storbreen – comprise a west-east transect representing the change in climate from the maritime coast to the continental inland. The results are published in annual reports (Haakensen, 1982; Østrem *et al.*, 1991; Elvehøy and Haakensen, 1992). In northern Norway, long-term mass balance studies started on Engabreen in 1970.

In addition, some glaciers have been studied for three to five years on behalf of hydropower companies. Several hydroelectric power plants have been considered for construction near glacierized mountains because of the quantity of water available at high altitudes and the relatively short distances to sea level. In connection with this, mass balance investigations have been undertaken at a number of selected glaciers. Austdalsbreen has been studied for seven years. The studies comprise mass balance, change in frontal posititon and ice movement. At Spørteggbreen, mass balance studies were ended after four years of taking measurements. Austre Okstindbreen has been studied for several years by researchers from the University of Aarhus, Denmark. Mass balance studies started in 1987 and will continue at least until 1995. Investigations started in 1989 at Svartisheibreen and in 1990 at Trollbergdalsbreen, both close to the Svartisen Ice Cap. In 1991, mass balance measurements began being taken on Storsteinsfjellbreen in Skjomen. Trollbergdalsbreen and Storsteinsfjellbreen had been measured previously in 1970–75 and 1964–68 respectively. In 1989, measurements were initiatied on Langfjordjøkulen, one of the northernmost glaciers in Norway. The activity was stopped in 1993. In 1994, mass balance investigations were carried out on 12 Norwegian glaciers – seven in southern Norway and five in northern Norway.

In southern Norway, the results show a different trend for the western, maritime glaciers as opposed to the more continental glaciers 200 km inland. In the west, the glaciers have been increasing in volume while, in Jotunheimen in the east, the glaciers had a tendency to decrease up to 1987. This period terminated in 1988 with an extremely negative net balance for all glaciers in southern Norway. In the period from 1989 until the present, conditions have shifted towards a more positive net balance for all glaciers. The maritime glaciers have had a higher positive net balance. Even the continental glaciers, however, have known a mass surplus in the past few years. Most of the glaciers in southern Norway recorded their most positive net balance ever measured in 1989. Apart from 1967, four of the most positive years were measured in the period 1989–94. On Nigardsbreen (48 km^2), one of the main outlet glaciers from Jostedalsbreen, the mass balance has shown a positive trend since 1962 with a cumulative mass

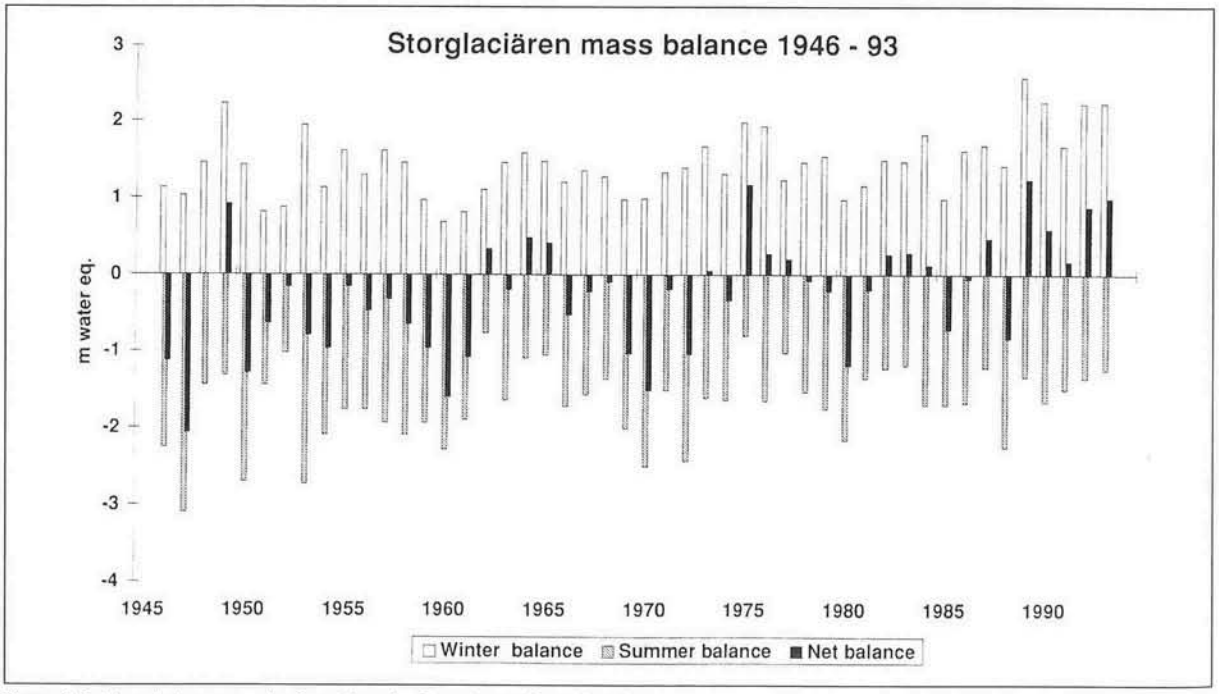

Figure 9.4 Mass balance results from Storglaciären in northern Sweden.

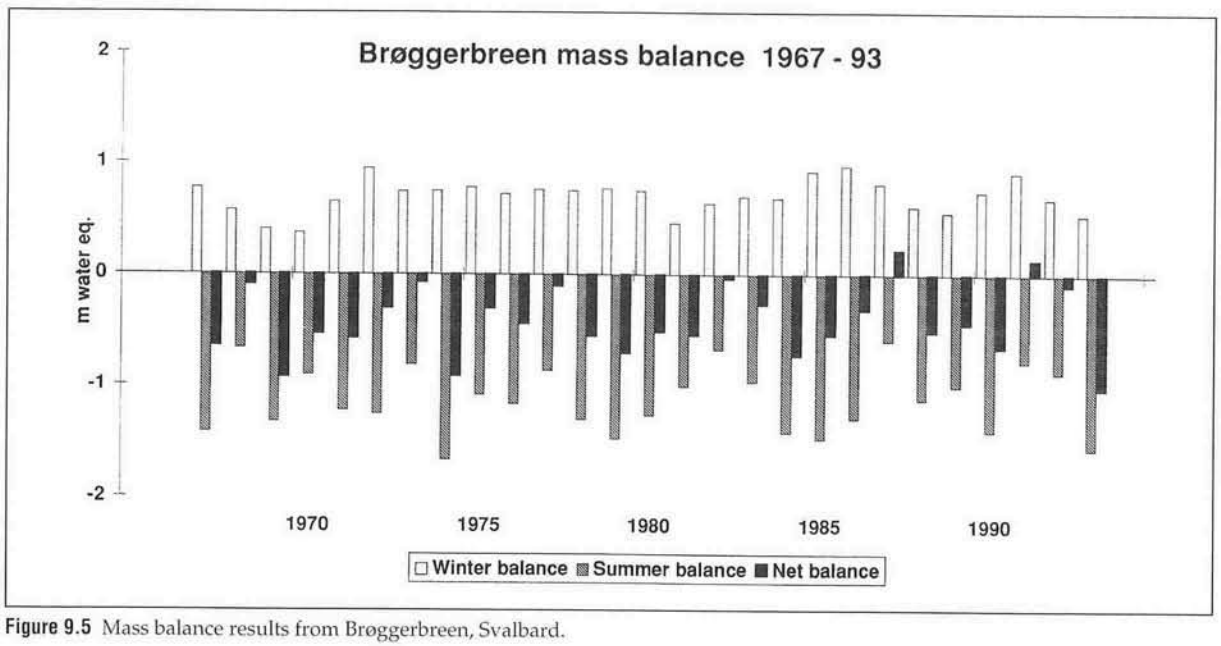

Figure 9.5 Mass balance results from Brøggerbreen, Svalbard.

increase of 6.5 m water equivalent in the 26-year period 1962–1988. During the five years up to 1993, the cumulative growth was an additional 8.5 m water equivalent, due mainly to higher winter precipitation, but also partly to cool summers. Results from the west-east transect are shown in Fig. 9.3, where Ålfotbreen represents the most maritime conditions, Nigardsbreen represents Jostedalsbreen Ice Cap and Storbreen represents the more continental type of climatic conditions.

In northern Norway, conditions are more variable. The tendency is more difficult to pinpoint owing to the lack of long measurement series. Engabreen, which is very close to the ocean and has been observed for 25 years, has shown a considerably positive cumulative mass balance amounting to 18 m of water equivalent, which is more than on any of the observed glaciers in southern Norway in 32 years. On the other hand, Engabreen has not shown a tendency to increase since 1988, unlike the south Norwegian glaciers. Short measurement series from other glaciers in northern Norway indicate that the more continental ones may be close to equilibrium or have been slightly decreasing.

At the beginning of the 20th century, most Swedish glaciers reached an extension close to their Holocene maximum. During this century, a dramatic recession of the glaciers has been recorded, mainly owing to a summer temperature increase of about 1°C around 1910 A.D. However, during the 1980s, the retreat showed a declining rate of recession. Some of the larger glaciers are still retreating but most glaciers are now in a near-balanced state with today's climate and a few small glaciers are even advancing (Holmlund, 1993).

The modern center for the study of Swedish glaciers is the Tarfala Glaciological Station in Kebnekaise, Lapland. The station is situated 70 km west of Kiruna, 30 km from the end of the road, at an elevation of 1,130 m, about 400 m above the tree line. Two glaciers in the Tarfala Valley were surveyed in 1897 and several times between 1908 and 1920. A series of mass balance studies started on Storglaciären in the spring of 1946 are still being pursued (Schytt, 1981; Holmlund, 1987; 1993; 1995). On Storglaciären, all three balance terms (winter balance, summer balance, net balance) are known for each year, starting in 1945/46 (Fig. 9.4). Several other studies have been, or are being, carried out on Storglaciären, such as flow dynamics, glacier hydrology/hydraulics, debris content/transportation, englacial temperature or glacier-bed topography and characteristics. To ensure that Storglaciären is reasonably representative of the glaciers in Lapland, 20 other glacier termini are also being quite regularly surveyed, from Kårsajökeln in the north (at 68°22′N) to Salajekna in the south (at 67°08′N).

In 1950, the Norwegian Polar Research Institute started the first systematic mass balance studies on Finsterwalderbreen on the south side of Van Keulenfjorden on Svalbard. Expeditions by the Norwegian Polar Research Institute to Svalbard were carried out every second year from 1950 to 1966. Therefore, net mass balance data can only be given as mean values for every second year during this period. The measurements showed a steady decrease of the glaciers, with a mean net balance of –0.25 m/year water equivalent. In 1966, investigatons started in the Kongsfjord area on Brøggerbreen and a year later on Lovénbreen; these measurements have been carried out every year since. Both accumulation and ablation have been measured by the direct glaciological method: snow-sounding profiles, density measurements and stake readings. The results for Brøggerbreen are illustrated in Fig. 9.5. The glaciers are not in balance with the existing climate as the ice masses have been steadily decreasing, with a mean net balance of about –0.40 m/year on Brøggerbreen.

Only two years (1987 and 1991) with a positive net balance have been registered. The average equilibrium line is about 100 m higher than the level that gives zero net balance. Steady state conditions would be obtained if the average summer temperature was lowered by 1°C or if the winter precipitation increased by about 50% (Hagen and Liestøl, 1990). There is a strong correlation between the annual net mass balance and the equilibrium line altitude (ELA). The ELA, in turn, is strongly correlated to the mean summer temperature or the sum of positive degree days during the melt season. Based on temperature recordings since 1912 in Svalbard, this high correlation has been used to reconstruct the net mass balance on Brøggerbreen since 1912 (Lefauconnier and Hagen, 1990). The total ice mass lost in the period 1912–1988 was 34.35 m of water equivalent, corresponding to a mean value of –0.45 m per year. This is almost 30% of the total ice volume of Brøggerbreen.

Russian glaciologists started systematic annual mass balance measurements in 1966 on Vöringbreen in Grønfjorden. In the years 1973–1976, they extended the programme to include three other glaciers, two in central-west Spitsbergen and one on the east coast. The results tally with the Norwegian recordings. Polish researchers have studied the mass balance on Hansbreen in Hornsund on southern Spitsbergen since 1988 and the frontal position has been mapped for thirty years. Both the Norwegian and Russian mass balance measurements have been carried out on relatively small (2–6 km^2), isolated cirque or valley glaciers close to the coast. The main parts of these glaciers are below 500 m a.s.l. Only sporadic measurements have been taken on large glaciers and ice caps. For this reason, mass balance investigations were initiated on Kongsvegen (105 km^2) in 1987. The results from seven years of study indicate that glaciers covering higher accumulation areas are closer to a steady state than the lower cirque glaciers closer to the coast.

The fluctuations of Italian glaciers began to be monitored just one hundred years ago, the year 1995 marking the centenary. In 1895, within the Club Alpino Italiano (CAI), a Glaciological Commission had been set up with this precise task; in 1914, the Comitato Glaciologico Italiano (CGI) was founded, with its head office in Turin, and the same year saw the publication of the first issue of the *Bollettino del Comitato Glaciologico Italiano*. However, systematic annual monitoring of glacier length variations only started in 1925 and has continued regularly ever since, results being published in the CGI's *'Bollettino'* and, since 1977, in a special section of the new journal, *Geografia Fisica e Dinamica Quaternaria-Bollettino del Comitato Glaciologico Italiano*. Data on the variations of Italian glaciers also appeared in the 'Reports' of the various international commissions published from 1895 until 1959 (Haeberli *et al.*, 1989). As from 1959–1965, they were inserted in Vol. I of *Fluctuations of Glaciers* of the Permanent Service on Fluctuations of Glaciers (PSFG), now the IAHS(ICSI)/UNEP/UNESCO World Glacier Monitoring Service (WGMS), and in later issues until 1985–1990.

Annual surveys carried out by CGI observers and, more recently, also by the CAI cover an average sample of about 100 glaciers, with measured variations from 1925 to 1992 (the number of glaciers on which simple qualitative observations are made is generally much larger). Surveys normally consist of snout measurements from fixed points, integrated with aerial photographs and satellite images (Serandrei *et al.*, 1993), and aerial and land-based photogrammetric surveys. In addition to data collection on snout variations, evaluations of the ELA and accumulation area ratio (AAR), selected annual meteorological data, verification of inventory entries and geomorphological observations are made. Repeated photographs are taken from fixed stations and this has allowed the creation and updating of an important photographic archive at the CGI. The CGI periodically organizes glaciological meetings, the most recent of which, on an international level, was held in Gressoney (Valle d'Aosta) in 1991. The proceedings are published in Volume 15 of *Geografia Fisica e Dinamica Quaternaria-Bollettino del Comitato Glaciologico Italiano*.

Information on the historical variations of some of the major Italian glaciers dates back centuries and covers several glaciers in valleys which were easily accessible and had long been populated. The sometimes violent manifestations of such glaciers could not escape attention and were often accompanied by fear because they lay in an environment which was generally alien to, and beyond the scope of, man's activities and economy, and because they were even able to cause serious natural calamities. After the varied and sometimes contrasting events of glacier activity in the second half of the 19th century and the early years of the 20th, collectively representing the steady decline of the Little Ice Age, the recent behaviour of Italian glaciers shows several phases of opposite sign. From the mid-1910s, numerous snout advances took place. In some cases, snouts reached the maximum positions they had occupied in the 19th century, although the volumes of the ablation tongues were not comparable with those of the Little Ice Age. Small but often clearly defined and recognizable end moraines bear witness to this small positive variation.

This phase terminated around the early 1920s and was followed by a period of intense retreat with more or less generalized withdrawal of monitored glacial snouts until the late 1950s. The corresponding effects are reflected in the 1959–62 *Catasto dei Ghiacciai Italiani* listing a number of glaciers which had disappeared in the preceding 50 years (about 20% of the total) and documenting the transformation of many former valley glaciers into smaller mountain glaciers. The early 1960s showed a new tendency towards glacier growth (Zanon, 1985; Wood, 1988), the number of glaciers advancing in the Italian Alps reaching its maximum in 1980 (87% of the monitored glaciers), in accordance with the observed glacier fluctua-

tions on the Austrian and Swiss sides of the Alps, where the number of advancing glaciers also reached its peak in 1980 (Patzelt, 1985). Climatically, in the case of the Eastern Alps, this phase was due to a drop in the summer temperature of 1°C and an increased annual precipitation of 3–4% with respect to the previous 30 years (Patzelt, 1985). Central alpine stations on the Italian side recorded summer temperatures down by 0.7°C in the ten-year period 1960–1969 and 0.4°C in 1970–79, together with increased winter precipitation of 5 % (Zanon, 1991).

The remarkable readvance of Italian glaciers occurred at different times and in different ways. Though relatively slight, it is of great interest since it was thoroughly observed and links the previous 40-year retreat of 1920–1960 with that taking place since the mid-1980s. Several medium-sized glaciers, mainly in the Venoste/Oetztaler Alps and the Ortles-Cevedale and Monte Bianco groups, have been advancing actively since the early 1960s, with overall progress of several hundred metres. For their part, the major valley glaciers have begun to advance, with response times of many decades after the beginning of progress. These times have often been longer than the duration of the fluctuation; the effects therefore exhausted themselves even before being reflected as snout advance, giving rise to only slight changes in glacial mass. Of the exceptions to this behaviour, the Brenva Glacier (Monte Bianco group), notoriously 'anomalous' as regards environmental influences (debris cover), advanced by about 500 m between 1965 and 1989 (Cerutti, 1992). An important consequence was the reappearance of glacierets and snowfields in place of those glaciers which had vanished during the first half of the century. This is revealed at least partly by the difference in number of glaciers and surface area between the 1959–62 *Catasto dei Ghiacciai Italiani* and the IAHS(ICSI)/UNEP/UNESCO-WGI (838 and 1,397 glaciers respectively, and approximately 540 and 608 km^2).

At the beginning of the 1980s, advance rates began to decline rapidly. This tendency accelerated in 1986 and reached its lowest ebb in the early 1990s. Hence, a new retreat phase has been predominant in the Italian Alps since the end of the 1980s. The number of advancing glaciers, still prevalent in 1981, has fallen steadily. The consequences of this new retreat will probably not be apparent in the 20th century, mainly as regards its intensity over such a short period. On the Caresèr Glacier (Ortles-Cevedale Group), according to two aerial surveys in 1980 and 1990, an average ice level variation of –11.24 m was recorded, with a maximum value of –18.34 m between 2,860 and 2,900 m a.s.l. It is significant that 67% of the overall loss in volume (54.2565.10^6 m^3) occurred in the area between 3,000 m and 3,150 m, thus involving most of the old accumulation area. Similarly, with an overall surface area reduction of more than 20 % (about 1 km^2), the maximum (relative) percentual reduction with respect to 1980 involved the highest area of the glacier between 3,200 m and 3,350 m (Giada and Zanon, 1991). In the same period, with an unchanged minimum altitude of 2,860 m, maximum elevation fell from 3,350 m to 3,310 m, mean elevation from 3,094 m to 3,075 m, and median elevation from 3,092 m to 3,084 m.

In the Italian Alps, the first example of real assessment of glacier mass balance parameters by means of direct glaciological measurements was the research carried out by U. Monterin between 1930 and 1935 on the Bors and Lys Glaciers (Monte Rosa group). This work may be considered the indirect result of the stimulus provided by H. W. Ahlmann for this kind of research during the 1930s, which was further developed after the Second World War. However, systematic research on glacier mass balance only really began in 1964, first on the Marmolada Glacier (Western Dolomites) (Zanon, 1965); from 1966 onward, such measurements have continued regularly on the Caresèr Glacier (Central Alps, Ortles-Cevedale group). The Caresèr Glacier is located on the south-eastern slope of the Cevedale massif in the high Pejo Valley (Noce-Adige basin). It is a cirque glacier of about 4 km^2, with altitudes ranging between 2,860 and 3,350 m a.s.l.. An experimental series of radar sections taken by helicopter in March 1994 gave maximum thicknesses for the central part between 35–40 m and 60–80 m (unpubl. data by Tabacco and Zanon). After a course of 1.8 km, the glacial stream feeds the ENEL (Italian National Electricity Board) hydroelectric reservoir of 16.10^6 m^3 capacity at an altitude of 2,600 m a.s.l.. From the climatic viewpoint, the glacier may be considered representative of the glacierized areas of the transitional belt between the continental-type alpine environment and that of the northern Mediterranean region. Data from the totaliser network installed on the glacier surface between 3,000 m and 3,200 m allows comparisons to be made between snow accumulation measurements and also with the long-term data at the dam site (Zanon, 1992); there is an automatic meteorological station on the glacier surface at about 3,200 m, managed by ENEL, for collecting climatological data with special regard to glaciological and environmental studies as part of the ALPTRAC project (another similar station is installed at Colle Vincent, Monte Rosa, at 4,090 m a.s.l.). Mass balance parameters are measured by direct glaciological recordings over a series of fixed date balance years, by comparing net accumulation and ablation. For the whole period of activity from 1966–67 to 1992–93, the average mass balance of the Caresèr Glacier was –0.52 m/year water equivalent. During this period, however, the regime passed through two distinct phases corresponding to two climatic trends which have been clearly documented for the evolution of alpine glaciers in general. From 1966–67 to 1979–80, mass balance was practically in equilibrium with a mean value of –0.13 m/year water equivalent. From 1980-81 to 1992–93, this value changed to –1.08 m/year water equivalent. Absolute peaks were recorded in 1990–91 (–1.79 m water equivalent), 1981–82 (–1.68 m water equivalent), 1986–87 (–1.64 m

water equivalent) and a least negative value of –0.30 m water equivalent in 1991–92. This regime shows a good qualitative fit with those of other sample glaciers in the Alps as monitored by WGMS. During the period in question, the ELA on Caresèr Glacier (3,090 m a.s.l. for a zero mass balance) ranged between 2,667 m and 3,485 m, with a mean value of more than 3,200 m at the beginning of the 1990s. For several years between 1980 and 1992, the ELA was higher than the maximum elevation of the glacier, with the result that the entire glaciated area had an AAR of zero for about ten years, compared with a value slightly higher than 0.5 for a zero mass balance. This situation typically reflected that of all the glaciers on the southern slope of the Central Alps up to 3,500 m, subjected to intense deglaciation still in 1992. Between 1967 and 1990, elevation and volume variations in the Caresèr Glacier were recorded, as already mentioned, by comparing digital models of the glacier's surface obtained from periodic aerial surveys. During the period 1967–1990, the altitude of the ice surface lowered by an average of –13.76 m. In terms of mean water depth, these values practically coincide with those obtained by direct glaciological recordings for annual mass balance (Giada and Zanon, in press).

In the last few years, research on glacier mass balance has also been carried out on the Sforzellina and Fontana Bianca/Weissbrunn Glaciers (Ortles-Cevedale group) and, more recently, on the Lys and Indren (Monte Rosa) and Ciardoney (Gran Paradiso) Glaciers (Mercalli *et al.*, 1993; 1994). Research has been pursued on Sforzellina Glacier since 1986–87 and balances have been reconstructed for the period up to 1970 (Barsanti and Smiraglia, in press). Research on the Fontana Bianca/Weissbrunn Glacier lasted from 1983/84 to 1987/88 (Secchieri and Valentini, 1992) and from 1991/92 to 1992–93 (Kaser *et al.*, in press).

On the Spanish side of the Pyrenees, two phases of the Little Ice Age can be seen in some moraines. The descriptions of Pyrenean writers at the end of the 18th and beginning of the 19th centuries show the Pyrenean glaciers in what we can consider to be their historical maximum extent. By the end of the 19th century, however, a notable retreat is obvious. This continues until 1957 (with a limited readvance around 1912) then lessens a little, only to accelerate again during the 1980s. At the beginning of the same decade, one hundred years after Schrader's estimates, the glaciers and glacierets of the Spanish Pyrenees occupied a total surface area of 632 ha, a figure which, ten years later (July 1991), had been reduced to 568 ha. (Table 9.6, cf. Martínez de Pisón *et al.*, 1995) as a consequence of accelerated degradation. This accelerated degradation is especially noticeable with smaller glaciers, whose situation has become critical. This tendency caused the disappearance of glaciers in Balaitus (SE Glacieret) and even in Aneto (North Cregüeña, Llosás and East Salencas). The glaciers of Aneto and Maladeta are, nevertheless, still the most outstanding perennial surface ice bodies of the Spanish Pyrenees.

Since 1991, Spanish glaciers have been monitored within an extensive project aimed at quantifying hydric resources from accumulated snow in the highlands (Arenillas *et al.*, 1992). Within this programme financed by the Spanish Government, it has been possible to begin research into present-day glacial dynamics. Maladeta Glacier, situated at the head of the Esera river and the second-biggest (0.5 km^2) of all Spanish glaciers, was chosen to be the object of this research. Mass balance measurements have been taken on the glacier, the geodetic surveying of the stakes installed to perform the mass balance measurements have provided information on surface movements of the ice and, by means of seismic and georadar soundings, ice thicknesses of up to 52 m have been determined (Martínez, 1995; Martínez and García, 1994).

9.5 SPECIAL EVENTS

Advances of Norwegian glaciers were observed mainly in the mid-18th century, during the Little Ice

TABLE 9. 6 Estimated surface area of glaciers in the (Spanish) Pyrenees in 1980 and 1991 arranged in descending order of hectares in 1991

Massif	*1980*	*1991*	*Area Losses*	
			(ha)	*(%)*
Aneto	324	302	22	6. 8
Perdido	107	90	17	15. 8
Infierno	62	60	2	3. 2
Posets	66	48	18	27. 3
Viñemal	20	18	2	10. 0
Perdiguero	18	17	1	5. 5
Balaitus	18	15	3	16. 6
Munia	12	10	2	16. 6
Besiberri	6	6	0	0. 0
Taillón	10	2	8	80. 0
Total	632	568	64	10. 1

Source: Martínez de Pisón *et al.*, 1995

Age. As described above, some outlets from Jostedalsbreen destroyed cultivated land and even houses on some farms. Since then, only small advances have been observed after long periods of retreat. However, even such small advances may cause problems. In the summer of 1986, the readvance of the some small outlets of Jostedalsbreen caused an ice avalanche that killed three tourists in the valley downstream.

A number of glacier-dammed lakes have been observed in Norway (Liestøl, 1956). The best known is Lake Demmevatnet, which is a lateral lake dammed by Rembesdalsskåki, a westerly outlet glacier from Hardangerjøkulen Ice Cap. Several outbursts from this lake have been recorded since about 1750 A.D. Almost every year, the lake would empty in August but heavy floods occurred at about twenty-year intervals. In 1893, a catastrophic outburst flooded the Simadalen Valley downstream from the glacier. After that, a rock tunnel was constructed that lowered the lake level by twenty metres, preventing further floods until 1937 when a new outburst caused heavier damage than ever before. The reason for the new flooding was the retreat of the glacier, which had reduced the thickness of the ice barrier considerably and left more space for water in the lake. A tunnel of about 5 m in diameter was later observed under the glacier and the total volume of water was estimated at $11.5 \cdot 10^6$ m^3, with an average flow of 900 m^3/s. After this latest outburst, another rock tunnel was dug 50 m below the former tunnel. Since then, no flooding from Demmevatnet has been reported. The greatest ice-dammed lake of recent times in Norway is the lake at Austerdalsisen, an outlet from Svartisen Ice Cap (66° N/14°E). The lake has emptied every year in August since the 1940s and 1950s, draining a volume of up to $150 \cdot 10^6$ m^3. Flooding has never caused any damage. This is the best studied glacier-dammed lake in Norway (Liestøl, 1956).

In connection with hydropower plants, two different water intakes have been constructed under glaciers in Norway. Under Bondhusbreen, a westerly outlet from Folgefonni in south-west Norway, a subglacial water intake has been constructed under 160 m of ice. Several investigations have been carried out at the interface between the ice and the bedrock (Hagen *et. al.*, 1993*b*). Under Engabreen, an outlet from Svartisen Ice Cap, North Norway, another subglacial intake has recently been constructed under 200 m of ice. Here, a permanent subglacial observatory will be established giving access to the glacier bed throughout the year.

In the Sälka massif of Sweden, about 20 km west of Kebnekaise, an ice-dammed lake close to Säla Glacier has been draining annually since 1990 (Holmlund, personal communication). The outburst volume is about $1 \cdot 10^6$ m^3 and roughly corresponds to one year's ablation on the glacier. A deep gorge observed in bedrock at the tongue indicates that lake outbursts may have been common events for a long period of time. The outbursts are connected to the size and thickness of today's glacier in that a thicker glacier would prevent outbursts and a thinner glacier would allow for normal drainage. Thus, the glacier seems to have had an extension similar to the present one for a very long time in order to cause the erosion of the deep gorge.

Numerous glacier-dammed lakes have been observed in Svalbard. Glacier-dammed lakes are readily formed, both on the surface and laterally. Lakes of this kind vary in size and most of them are short-lived. They usually empty during summer either by reopening of a supraglacial channel or by reopening of a moulin at the bottom of the lake (Liestøl *et al.*, 1980). At most glacier fronts, the water drains subglacially and contains large amounts of suspended sediments. One subglacial lake, Setevatnet, at Kongsvegen in the Kongsfjord area on the north-west coast, has been observed to empty englacially. About $40 \cdot 10^6$ m^3 of water lifted part of the glacier and emptied along a 1 km-long crack across the glacier 2 km downstream from the lake reservoir (Liestøl, 1976).

Glacier surge represents a dramatic increase in ice flow velocity up to hundred times the normal flow rate resulting in the transportation of a great volume of ice from higher to lower parts of the glacier, usually accompanied by a rapid advance of the glacier front. The phenomenon repeats itself periodically, the intervals between surge periods, the quiescent phase, varying from 30 to more than 100 years and being characteristic for each individual glacier. The duration of the active surge averages about 1–3 years (Meier and Post, 1969). The phenomenon seems to be most common in subpolar glaciers (with temperate and cold parts) and is characteristic for Svalbard glaciers. Glacier surges occur independently of short-term climatic variation, which only affect the length of the period between surges. Most of the glaciers on Svalbard are surging. In fact, surges have been dated on nearly 100 glaciers from 1860 to 1992 but several surges may have occurred which have not been recorded. At some of these glaciers, the change in longitudinal profiles and frontal positions has been recorded. Surge occurs on all glacier types, from small inland glaciers to large calving, tidewater glaciers. A typical difference in behaviour has been observed between large glaciers ending on land and those ending in the sea. In the latter case, the entire glacier system is usually affected, the main stream triggering the surge of many minor tributaries. On glaciers ending on land, however, only the surging stream is involved, resulting in the formation of complicated, folded moraine systems frequently observed on Svalbard glaciers. Several authors have described the surge events in Svalbard, including observations recorded by different expeditions. In 1839, the French Recherche Expedition described the Recherche Glacier in Bellsund as being heavily crevassed, with the ice front extending 3 km beyond its present-day position. Fridtjovbreen, on the north

side of Bellsund, surged in 1858–61. The glacier then advanced 6 km, filling the entire Fridtjov Harbour. The sea floor was partly pushed up in front of the glacier and shell-bearing clay banks were observed.

The two largest surges known in Svalbard occurred at approximately the same time, Negribreen in 1935–36 and Bråsvellbreen in 1937-38. In one year, Negribreen advanced 12 km into the fjord along a 15 km wide section of the front. Bråsvellbreen advanced 20 km into the sea with a 30 km wide front (Liestøl, 1969). Detailed observations have been made on Hessbreen in Van Keulenfjorden (Liestøl, 1974), on Usherbreen in Storfjorden (Hagen, 1987; 1988) and on Bakaninbreen, a tributary glacier to Paulabreen in the inner part of Van Mijenfjorden (Dowdeswell *et al.*, 1991). A review of surging and calving glaciers in eastern Svalbard has been compiled by Lefauconnier and Hagen (1991) in order to establish a list of potential future surging glaciers that could be considered as a source of icebergs in the Barents sea and thus affect offshore operations. An updated list of observed glacier surges is published in the Glacier Atlas of Svalbard and Jan Mayen (Hagen *et al.*, 1993*a*). When a glacier surges into the sea, the glacier becomes heavily crevassed and numerous – if relatively small – icebergs are produced during the active advance period. However, during the years following the advance when glacier activity decreases, fewer but larger icebergs are produced. The duration of the active phase is significantly longer on Svalbard glaciers than for surge-type glaciers observed elswhere (Dowdeswell *et al.*, 1991). In Svalbard, the active phase may last as long as 3–10 years, while a duration of 1–2 years is more typical in other regions, such as observed on Alaskan glaciers. Ice velocities during the active phase are also considerably lower; mass is transferred down-glacier more slowly, over a longer period, and the termination of the active phase is not very abrupt. When a glacier surges, characteristic features remain on the glacier itself and on the landscape and, when preserved, these formations provide evidence of earlier surges. Folded median moraines and frontal changes are typical. Folded frontal moraines after a strong push may occur on glaciers ending on a sandur plain below the old marine limit, as on Usherbreen (Hagen, 1987).

On the Italian side of the alpine chain, glacier action, although not as intense as that on the northern side, has always been a cause of natural calamities in high- and medium-altitude mountain environments. In historical times, these calamities were mainly linked with the events of the Little Ice Age. One of the best-known and most significant examples is the Vedretta di Solda/Suldenferner (Ortles-Cevedale group): in 1818, its snout approached the first houses in the valley, advancing about 1,000 m in a year, or 3 m a day (Dutto and Mortara, 1992). Other examples are the Brenva Glacier (Monte Bianco group), which partially destroyed the chapel of Nôtre Dame de la Guérison, near Entreves in the Aosta Valley, again in 1818; the periodic emptying of the glacial lake of S. Margherita, near the Rutor Glacier (Northern Graian Alps), the cause of ruinous flooding in the La Thuile Valley downstream and along the valley bottom of the River Dora up to 30 km away, which was exorcized in vain with propitiatory masses and religious processions bearing sacral relics, and; the periodic emptying of the ice-dammed lake near the Cevedale Glacier, which influenced the history of the Martello/Martell Valley in the South Tyrol from 1881 to 1887, until a special containing dam was finally built. Dutto and Mortara (1992) have studied about 90 of the most important and significant episodes of glacier activity in the Italian Alps in historical times (cf. Table 9.7).

In more recent years, one example of a proglacial lake emptying is that of the lake near the snout of the Locce Glacier on the eastern side of Monte Rosa (Anzasca Valley, cf. Haeberli and Epifani, 1986). In 1979, about 300,000 m^3 of water flowed along the flank of the adjacent Belvedere Glacier and, through a breach in its lateral moraine, gave rise to extensive mass transportation for more than 3 km. Still more recently, in the summer of 1989 (Dutto and Mortara, 1992), about two-thirds of the entire glacial mass of the Upper Coolidge Glacier (Monviso) became detached and an avalanche of about 200,000 m^3 of ice and detritus slid down the mountainside, resulting in a difference in height of slightly less than 1,000 m, at seismographically recorded speeds of between 90 and 130 km/hour.

Although not included in the types considered by the above authors, one indirect consequence of the presence of glaciers and, in particular, of extraordinary changes in the parameters regulating their activity, is flooding, partly due to accelerated ablation. These episodes typically occurred in many areas of the alpine chain in August 1987. In the Pennine Alps, the glacierized basins making up part of the Grande Dixence hydroelectric system were especially badly hit (Rey and Dayer, 1990). In the Italian Alps, typical flooding occurred in Valtellina and Alto Adige/South Tyrol (unpubl. data by Zanon and Carollo). In the latter region, in some extensively glaciated valleys, such as Vallelunga/Langtauferer, Mazia/Matsch, Solda/ Sulden, Trafoi/Trafojer, Martello/Martell, Ridanna/ Ridnaun , and Aurina/Ahrn, rainfall in itself unexceptional was exacerbated by increased net ablation, estimated at approximately 40 mm/day water equivalent, as shown by the reduced albedo due to early melting of the seasonal snow cover and by an anomalous altitude of the zero isotherm. The result was heavy discharges along the main streams, with great changes in stream beds, right from the proglacial areas. In the case of the Martello/Martell Valley, meteoric waters and meltwater added to the effects of discharge from the Gioveretto/Zufritt hydroelectric reservoir, giving rise to flooding down to the confluence with the Venosta/Vinschgau Valley, where the most disastrous consequences were felt.

TABLE 9. 7 Italian Alps: glaciers exhibiting instability phenomena

Legend: 1) ice fall from snout of glaciers; 2) supraglacial debris fall/slide outside the lateral moraine; 3) landslide involving ice; 4) rapid advance of snout of glacier; 5) emptying of internal water-pocket; 6) emptying of proglacial lake; 7) emptying of ice-dammed lake; X = a case; [X] = more cases

Glacier	*1*	*2*	*3*	*4*	*5*	*6*	*7*
Belvedere		X			[X]		
Brenva		[X]	X	[X]	[X]		
Brouillard				X			
Cevedale							[X]
Cherillon	X						
C. Monticello	X						
Coolidge Sup.	X						
Coupé de Money	X						
Forni	X						
Freney	[X]				X		
Galambra							[X]
Gemelli di Ban						[X]	
Grandes Jorasses					[X]		
Gran Neyron					X		
Hohsand Merid.	[X]						
Lex Blanche	X			X			
Locce Sett.						[X]	
Becca di Luseney			X				
Lys				X	X		
Mandrone	[X]						
Malavalle							X
Miage		[X]				X	
Ormelune	X				X		
Patrì	X				X		
Perazzi			X				
Planpinceux	X				[X]		
Prasec					X		
Pré de Bar				X			
Rochefort					X		
Rutor							[X]
Scerscen						X	
Sissone					X		
Solda				X			
Triolet			X				
'Velan'					X		
Verra		X		X			

Source: Dutto and Mortara, 1992

9.6 GAPS AND NEEDS

At present, the mass balance investigations have good coverage in southern Scandinavia, with a profile extending from the maritime glaciers on the west coast to increasing continentality in the eastern part of the Jotunheimen area. In a south-north profile in Scandinavia, measurements are currently carried out in southern Norway (61°N–62°N – maritime and continental, in the Svartisen area (66–67°N – maritime) and in northern Sweden in the Kebnekaise area (68°N – continental). Further north, there are no ongoing measurements being taken, despite the large glacier areas. However, over a five-year period (1989–1993), Langfjordjøkulen Glacier (70 °N) was measured. This glacier showed a different pattern from the other Scandinavian glaciers, with no obvious correlation. Measurements taken over the five-year period indicate a stable glacier close to equilibrium. This is the only information from this northernmost part of Scandinavia, which makes Langfjordjøkulen a suitable glacier for long-term measurements. Investigations here would close a gap between the Scandinavian and Arctic stations in Svalbard and cover a latitude for which there is otherwise little information.

In the Svalbard Archipelago, mass balance investigations are running on three glaciers on the northwest coast by the Norwegian Polar Institute and one in south-Spitsbergen by Polish scientists. No mass balance investigations are being carried out in the eastern part of the archipelago. A corresponding pro-

gramme on Nordaustlandet could contribute to a better understanding of the climate in the North Atlantic region. It would also be part of a transect from maritime glaciers in the west to gradually dryer climatic conditions in the east and, thus, reflect part of the climatic gradient across the whole of the glacierized Eurasian High Arctic. There is very little data available from eastern Svalbard and the Russian archipelago, Franz Josef Land and Severnaya Zemlya further east.

9.7 FUTURE DEVELOPMENT OF MONITORING ACTIVITY

Systematic monitoring of glacier mass balance components and glacier extent is required to identify trends in the climate and understand the coupling between climate and glaciers. The uncertainty in identifying the causes of sea-level rise over the past century and in forecasting future changes stems in part from incomplete knowledge of the mass balance of the world's ice masses. Lack of continual observation is the main cause of uncertainty in the glacier mass balance. The current observations in Scandinavia and Svalbard are of great importance in this context because these are some of the few areas where the series are long enough to start analysing the climate. The challenge will be to obtain funding to continue the series in the coming years into a period when Global Circulation Models (CGMs) predict a warmer climate. This will be first detectable and have the greatest effect in arctic and subarctic areas. It will therefore also be important to continue the north-south transect and that from maritime to more continental glaciers.

For some additional glaciers, it might be advisable to establish a simple field programme for short periods. In Sweden, it is proposed that the mass balance gradient be obtained on several glaciers by means of a study over several years and that, thereafter, the equilibrium line altitude (ELA) or the accumulation area ratio (AAR) be measured, since the ELA is well correlated to the overall net mass balance of the glaciers. Observations of the ELA alone give the general climatic response of the net balance to the glaciers but do not separate the climatic effect on the summer ablation and the winter accumulation. It is, therefore, necessary to continue field measurements on some selected glaciers where individual time series of winter snow accumulation and summer balance/summer ablation are measured, to understand the effects of climatic change .

Remote sensing should be developed more in future monitoring of the ELA of glaciers. Radar images (SAR 7) seem to be able to distinguish wet snow and firn from glacier ice. Since SAR-images can be obtained even in cloudy weather conditions, the technique should be developed further for operational monitoring purposes. Modelling of the mass balance from degree-day models and energy balance models have proven to give very reliable results. An empirical degree-day model has been developed in Norway and applied on south-Norwegian glaciers where long-term mass balance studies have been carried out (Laumann and Reeh, 1993). The model could be tuned and tested on field measurements. The model gives reliable results using temperature and precipitation from nearby meteorological stations as input. These models should be used in combination with remote-sensing data in the future monitoring of glaciers, so that expensive field work can be restricted to a limited number of glaciers.

The currently growing interest in glaciological research in Italy, after a century of observations on glacier fluctuations, is in unique contrast with the reasons for the setting up, in 1895, of the Glaciological Commission of the CAI and, 20 years later, of the Comitato Glaciologico Italiano. The still imposing development of the alpine glaciers at the end of the 19th century as a consequence of the Little Ice Age and, to an even greater extent, their considerable progress in the first two decades of the 20th century, indicated their essential role not only in the regime of the rivers and groundwaters of the Po Plain but also as factors of prime importance in the growing hydroelectric industry and in the entire industrial development of northern Italy. One century after the end of the Little Ice Age, however, the onset of accelerated melting has given rise to glacier mass losses of more than 50% with respect to the mean of the 20th century (Haeberli, 1994). This phenomenon from the ten-year period 1980–1990 is expected to become even more intense in the mid-1990s. Apprehension at the prospect of the prolonged duration and consolidation of this trend not only concerns the very existence of this glacial phenomenon at medium latitudes but also the strong implications of this phenomenon for the climatic conditions which gave rise to it and the possible influence of man's activities. As regards environmental sciences, therefore, glaciological research emphasizes its identity, its methodologies and its goals. Together with the other earth sciences, it can supply valuable information for identifying the global problems which beset this centenary. In this regard, greater unity and awareness on a national scale and the strengthening of international collaboration are both fundamental.

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10 Glaciers in Africa and New Zealand

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10.1 DISTRIBUTION AND CHARACTERISTICS

On the African continent, glaciers still exist on three high mountains near the Equator (Fig. 10.1), namely Kilimanjaro in northern Tanzania, the Ruwenzori straddling Uganda and Zaire, and Mount Kenya (Figs. 10.2, 10.3, 10.4). Comprehensive accounts have been published of the glaciers of Equatorial East Africa (Hastenrath, 1984; Young and Hastenrath, 1991), including map documentation and bibliographic references.

On Kilimanjaro, ice cover now is limited to the volcanic cone of Kibo and represents the remnants of a formerly much larger ice cap. Twenty ice entities are recognized, with a total area of about 5 km^2. The ice cover is more extensive on the south than the north side, commensurate with the location of the mountain in the southern hemisphere. The eastern rim of the crater is now free of ice, contrasting with large glaciers to the west. This azimuth asymmetry can be understood as a consequence of the pronounced diurnal circulation systems typical of tropical high mountains: in the cloud-free mornings, the eastern flanks receive abundant solar radiation, whereas the west side remains shielded from the solar rays by the abundant cloudiness in the afternoon.

In the Ruwenzori, 43 glaciers existed recently, covering an area of about 4 km^2. There are small valley glaciers in addition to glaciers descending from larger ice caps. Azimuth asymmetries are not pronounced, presumably a consequence of the generally very cloudy conditions in this mountain massif.

On Mount Kenya, there are presently 11 valley glaciers, with an overall area of less than 1 km^2. The largest ice masses are found to the southeast, where precipitation is most abundant.

New Zealand has 3,153 inventoried glaciers, covering a total area of 1,159 km^2 and containing an estimated volume of 53.3 km^3 of ice. The glaciers are concentrated along the Main Divide of the central Southern Alps, with both the number and size of glaciers diminishing to the north and south (Fig. 10.5). In the North Island, glaciers occur only on the volcano of Mt. Ruapehu.

Most types are represented, dominantly alpine glaciers, with some large compound valley glaciers. Debris-covered glacier tongues are common on the large, low gradient glaciers, both in the high precipitation zone on the western side of the Alps and along the low precipitation areas to the east. Well-developed rock glaciers are formed in the more arid areas to the east of the Alps, with the most spectacular of

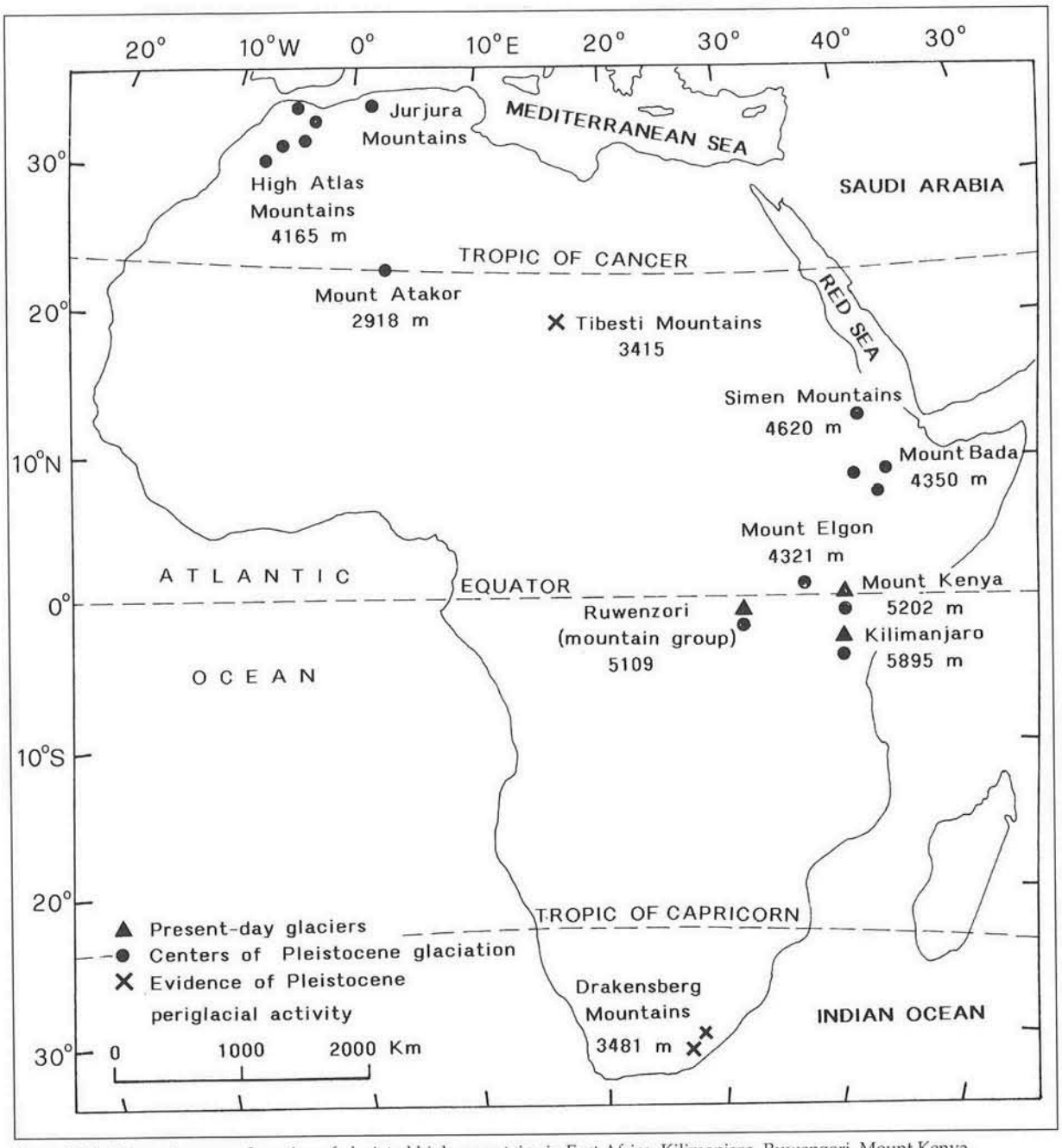

Figure 10.1 Orientation map. Location of glaciated high mountains in East Africa: Kilimanjaro, Ruwenzori, Mount Kenya.

these being on the Inland Kaikoura Range. One crater glacier is located on Mt. Ruapehu and one only small ice cap with radial ice flow is recognized.

10.2 EXISTING INVENTORIES

Glacier inventories for all three African mountains, mostly representative of the 1970s, are contained in a book by Hastenrath (1984) and have also been reported to the World Glacier Monitoring Service. In addition, for Mount Kenya, new maps have been produced for the 1987 and 1993 epochs and, based on these, updated inventories have been compiled (Hastenrath *et al.*, 1989; Rostom and Hastenrath, 1994).

The first attempt at an inventory of New Zealand's glaciers was made in 1967 with a *Southern Hemisphere Glacier Atlas* (Mercer, 1967). This inventory describes valley by valley the known larger glaciers, with references, photo sources and a map on a scale of approximately 1:1,000,000. No attempt was made by Mercer to estimate glacier lengths, areas or ice volumes. A provisional inventory compiled for the purpose of estimating water resources held as perennial snow and ice (Anderton, 1973) identified a total of 527 glaciers covering an area of 810 ± 40 km^2 with an ice volume of approximately 63 km^3. In 1971, the Water and Soil Division of the Ministry of Works and Development was contacted with the initial suggestion that New Zealand should become involved in compiling an inventory of the country's glaciers. This comprehensive inventory, containing about 24 pieces of data for each glacier, was completed in 1991 but has not been published as yet. The inventory recognized 3,153 glaciers covering a total area of 115.9 km^2, adopting the following definition for the minimum

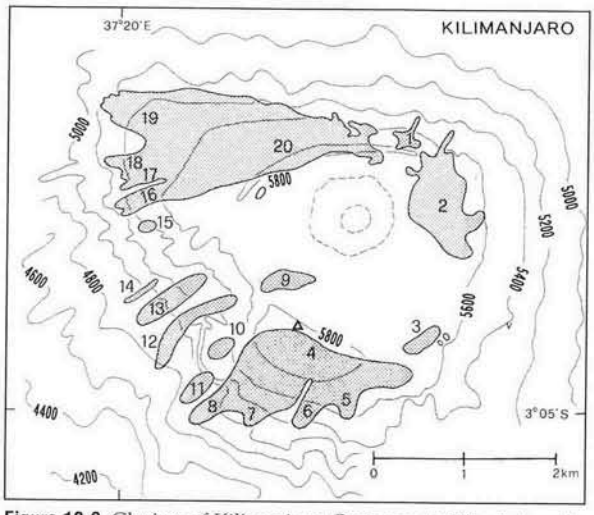

Figure 10.2 Glaciers of Kilimanjaro. Contours at 200 m intervals, ice areas stippled:1. Eastern Ice Field, 2. -, 3. Ratzel Glacier, 4. Southern Ice Field, 5. Rebmann Glacier, 6. Decken, 7. Kersten, 8. Heim, 9. Furtwängler (Western Crater), 10. Diamond, 11. Balleto, 12. Great Barranco (Great Breach), 13. Little Barranco (Little Breach), 14. Arrow, 15. Uhlig, 16. Little Penck, 17. Great Penck, 18. Drygalski, 19. Credner, 20. Northern Ice Field.

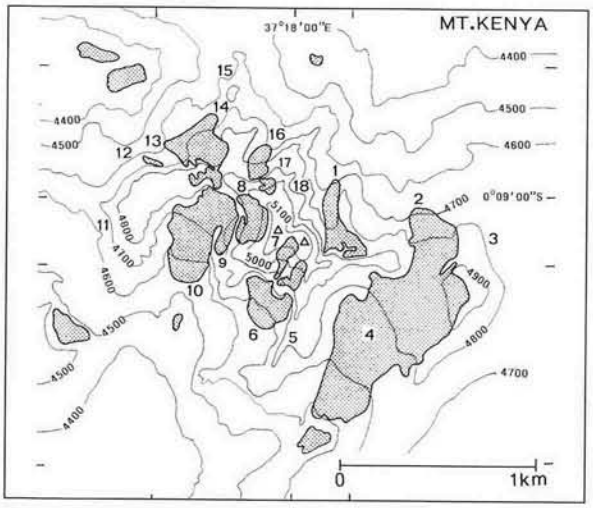

Figure 10.4 Glaciers of Mount Kenya. Contours at 100 m intervals. Large number denote the following glaciers: 1 = Krapf; 2 = Gregory; 3 = Colbe (disappeared after 1926); 4 = Lewis; 5 = Melhuish (disappeared after Feb 1978); 6 = Darwin; 7 = Diamond; 8 = Forel; 9 = Heim; 10 = Tyndall; 11 = Barlow (disappeared after 1926); 12 = NW Pigott (disappeared); 13 = Cesar; 14 = Joseph; 15 = Peter (disappeared after 1926); 16 = Northey; 17 = Arthur (disappeared); 18 = Mackinder (disappeared)

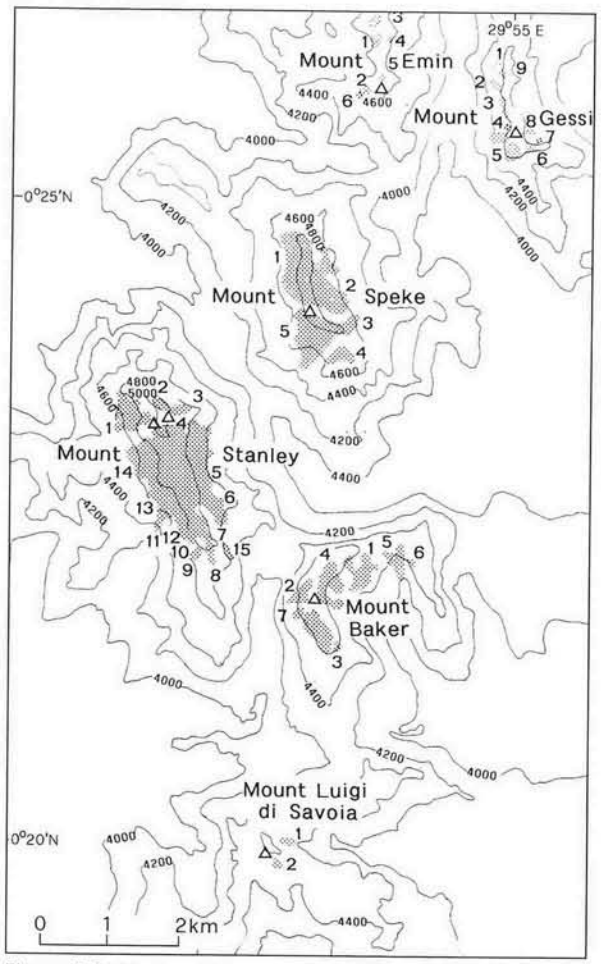

Figure 10.3 Glaciers of Ruwenzori

Mount Emin 4,802 metres:
1 = Kraepelin 1; Umberto (disappeared after1906); 2 = Emin 1; 3 = North Kraepelin; 4 = Kraepelin 2; 5 = Emin 2; 6 = Emin 3; one unnamed (disappeared after 1906)

Mount Gessi 4,769 metres:
1 = Gessi 1; 2 = Gessi 2; 3 = Gessi 3; 4 = Gessi 4; 5 = Iolanda 1; 6 = Iolanda 2; 7 = Iolanda 3; 8 = Iolanda 4; 9 = Gessi 0

Mount Speke 4,891 metres:
1 = Grant; 2 = Vittorio Emanuele; 3 = East Johnston; 4 = Johnston; 5 = Speke

Mount Stanley 5,111 metres:
1 = Alexandra; 2 = Albert; 3 = Northeast Margherita; 4 = Margherita; 5 = East Stanley; 6 = Elena; 7 = Coronation; 8 = Savoia; 9 = Philip; 10 = Elizabeth; 11 = West Elena; 12 = West Savoia; 13 = Moebius; 14 = West Stanley; 15 = unnamed

Mount Baker 4,873 metres:
1 = East Baker; 2 = Y Glacier; 3 = Edward 1; 3 = Edward 2; 4 = West Baker; 5 = Moore (Mubuku); 6 = Wollaston; 7 = Edward 3; Semper (disappeared after 1943)
Mount Luigi di Savoia 4,665 meters:
1 = Thomson 1; 2 = Thomson 2; Sellh (disappeared after 1906); Stairs (disappeared after 1906)

glacier size: 'Those ice bodies of 1 ha or greater in area which have remained in existence during the most negative balance years over the past two decades'.

10.3 LONG-TERM OBSERVATIONS (LENGTH CHANGE, MASS BALANCE, MAPS)

Reconstructions of long-term changes for all three African mountains are documented in Hastenrath (1984). Evolutions into the most recent decades have been traced in detail for Mount Kenya (Hastenrath *et al.*, 1987; Rostom and Hastenrath, 1994) and have been reported to the World Glacier Monitoring Service. Concerning the chronology of the glacier recession, two points must be noted: the earliest observations date back to the latter part of the 19th century and there is conclusive evidence that the ice began to retreat after 1880.

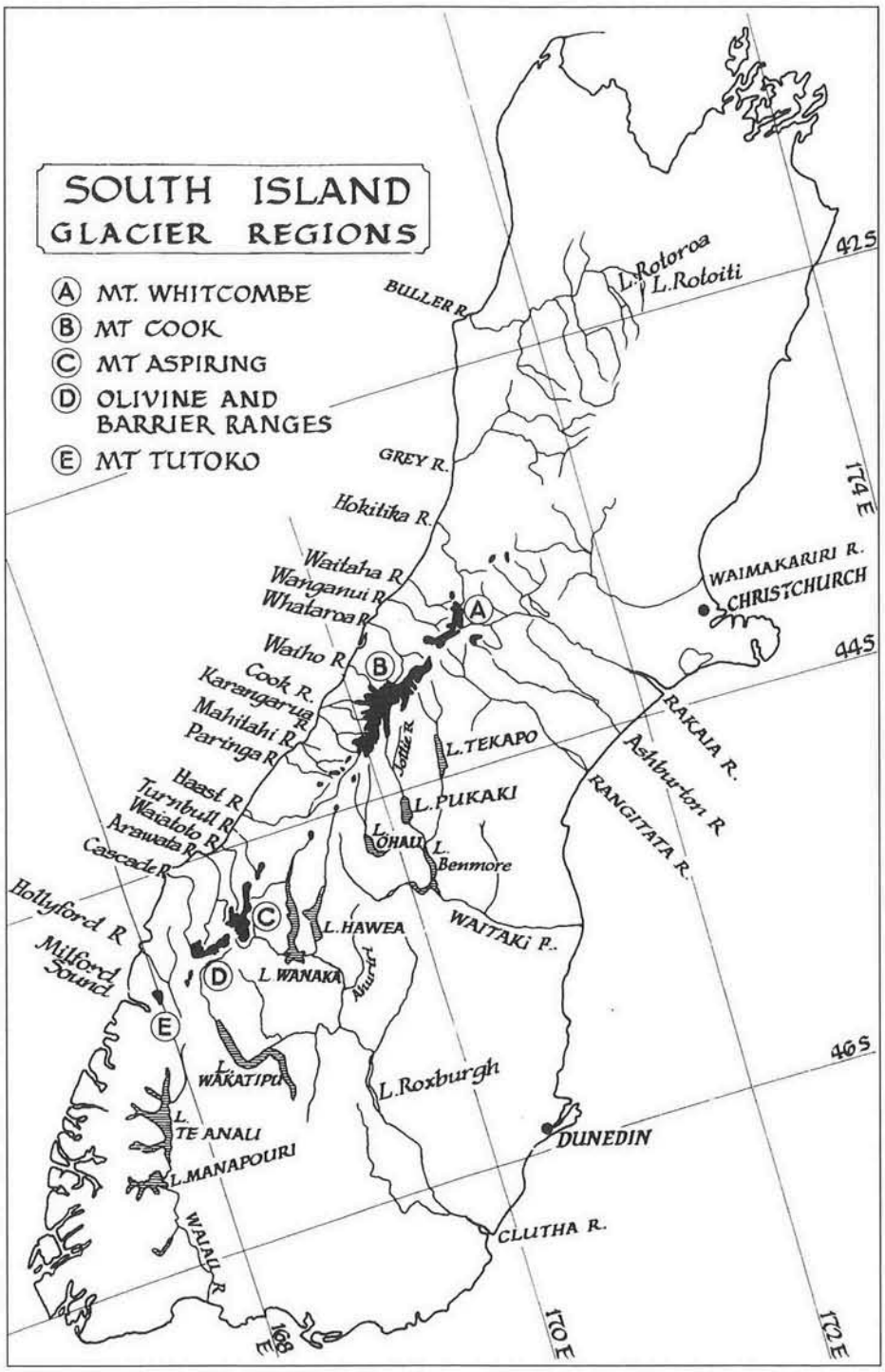

Figure 10.5 Distribution of the main glaciated areas of the Southern Alps, New Zealand.

Available maps and air photography are listed by Hastenrath (1984). For Mount Kenya, maps on a scale of 1:5,000 of the entire glaciated area exist for 1963, 1987 and 1993 (Forschungsunternehmen Nepal-Himalaya, 1967; Hastenrath *et al.*, 1989; Rostom and Hastenrath, 1994). For Lewis Glacier on Mount Kenya, in particular, maps on a scale of 1:2,500 have been produced at four-yearly intervals from 1974 to 1990 and again in 1993 (Hastenrath *et al.*, 1995), and also in 1983 (Patzelt *et al.*, 1984).

Net balance and ice flow velocity, as well as precipitation, have been monitored on Lewis Glacier on Mount Kenya continually since 1978 (Hastenrath, 1984; 1991). Measurements from the glacier observation programme on Mount Kenya, along with the reconstruction of changing ice extent and volume, have served as input to numerical modelling and sensitivity studies aimed at the inference of climatic forcing (Hastenrath, 1984; 1994; Kruss, 1984; Hastenrath and Kruss, 1992*a*; 1992*b*). In brief, the onset of glacier recession in East Africa was caused by a drastic decrease in precipitation and cloudiness from pre-1880 to post-1900 conditions, which in turn was due to an acceleration of the boreal autumn equatorial westerlies over the Indian Ocean. In the first half of the 20th century, warming accounted for the ice thinning on Mount Kenya. By contrast, from 1963 to 1987, enhanced atmospheric humidity was instrumental and this may in part be a consequence of enhanced evaporation in the Indian Ocean. Two

findings stand out from this work: the onset of glacier recession in East Africa occurred distinctly later than in New Guinea and the Ecuadorian Andes and; factors other than temperature are also instrumental in the shrinkage of tropical glaciers.

The earliest observations of glacier termini fluctuations in New Zealand were available from around 1860, with detailed observations and surveys made sporadically from 1890 onwards, the late Little Ice Age maximum, at a few of the more accessible gla-

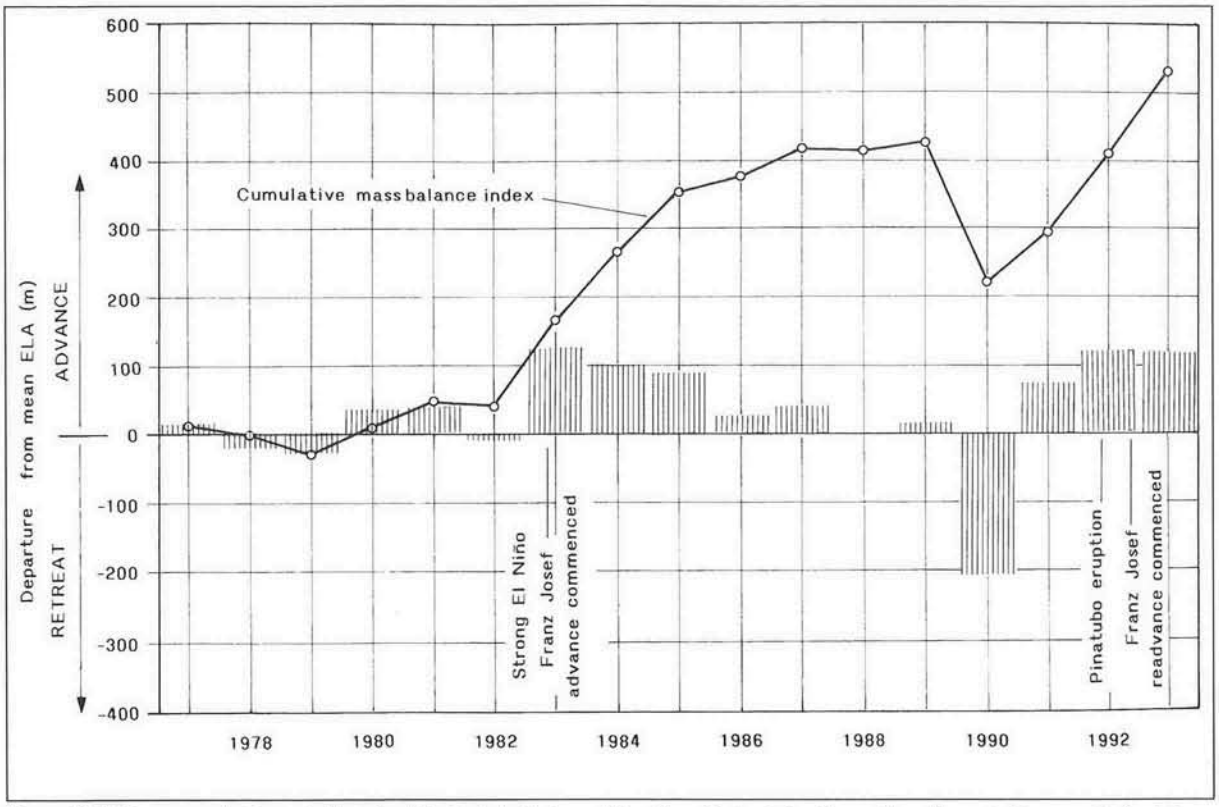

Figure 10.6 Mean mass balance indices for 47 selected glaciers of the New Zealand Southern Alps. These indices are derived from annual end-of-summer snow-line altitudes (ELAs). The mean annual and cumulative plots are annotated with significant glacial events.

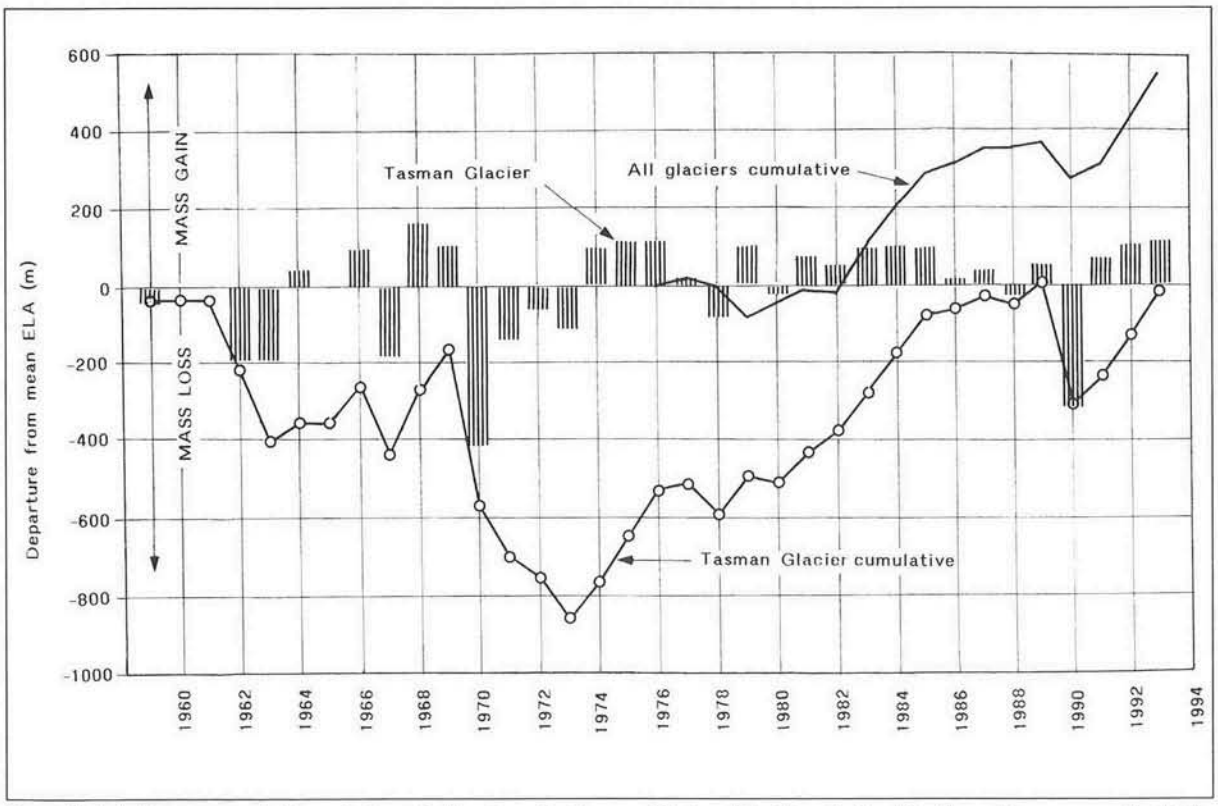

Figure 10.7 A 34-year record of mass balance indices from the Tasman Glacier of the New Zealand Southern Alps compared with the mean indices of 47 selected index glaciers surveyed since 1977. These indices, given as mean annual and cumulative plots, are derived from annual end-of-summer snow-line altitudes (ELAs).

ciers. Early observations include those made of the Franz Josef and Fox Glaciers; the glaciers of Mt. Cook, the Godley area and the glaciers of the Rangitata River. The only regular observations of frontal fluctuations are available from Franz Josef Glacier since 1865 (Sara, 1968) and Stocking Glacier (Salinger *et al.*, 1983).

The maximum extents for the Little Ice Age are frequently readily recognized by large sharp-crested moraines, identifiable in aerial photographs. A selection of 105 neoglacial moraines were identified in aerial photographs, the amount of length loss to the present being measured on the maps used for the inventory. The maximum late neoglacial extents were reached variously between 1750 and 1890 for different glaciers. The results show a mean loss in length of nearly 38%.

Few mass balance studies have been carried out on the NZ glaciers. Goldthwaite and McKellar (1962) carried out the first elementary mass balance studies on the Tasman Glacier, commencing in 1957–58 and continuing spasmodically for nearly a decade. Thompson and Kells (1973) took two seasons of balance measurements on the Whakapapanui Glacier on Mt. Ruapehu over 1968–69 and 1969–70. Meanwhile, following the investigations of Goldthwaite and McKellar, the Ministry of Works and Development commenced snow measurements on the Tasman Glacier in 1965 as an addition to a hydrological survey programme in the Waitaki Basin. Accumulation measurements were carried out at approximately bimonthly intervals until 1975. After 1971, measurements included ice ablation losses on the glacier tongue. Systematic mass and heat balance measurements have been carried out only on the Ivory Glacier, a small cirque glacier, from 1969 to 1975 as part of an IHD programme of representative basin studies (Anderton and Chinn, 1978). Partial mass balance studies have also been made on the Dart Glacier, on and off over the period 1975 to 1987 (Bishop and Forsyth, 1988). Latterly, the only systematic glaciological measurements being taken (apart from routine monitoring of the frontal position of the Franz Josef Glacier) have been annual surveys, since 1977, of the end-of-summer snow-line altitude (i.e. equilibrium line altitude, ELA) on some four dozen selected glaciers. As shown in Figs. 10.6 and 10.7, this provides a useful surrogate for mass balance over a wide coverage of the Southern Alps (Chinn, 1995). These surveys have detected a series of positive balances some time before the majority of glaciers commenced advancing.

10.4 SPECIAL EVENTS

Perhaps the most significant recent glaciological event in New Zealand has been the reversal of glacial retreat, which has persisted for the past century. This was first manifest in 1983 at the snout of the very reactive Franz Josef Glacier (Fig. 10.6). Analysis of the snow-line survey data shows that equilibrium to positive balances returned around 1978 (Fig. 10.7). Both the positive balances and glacial advances are continuing. Despite the current period of positive glacier balances, the majority of the large low-gradient glaciers have recently entered a phase of pro-glacial lake formation and are thus decoupled from the climate signal. The Godley Glacier has had the largest lake development, followed by the Classen, Maud and Grey Glaciers in the same area. The Tasman Glacier lake is accelerating in growth and this is likely to accelerate further, as the adjacent Murchison River breached a moraine wall to enter the lake during a flood in January 1994, changing the temperature regime of the lake. Pro-glacial ponds at the fronts of the Murchison, Hooker, Mueller, La Perouse and Balfour Glaciers are expected to expand rapidly into large pro-glacial lakes.

On 14 December 1991, a large rock avalanche fell from the summit of Mount Cook (3,764 m), New Zealand's highest mountain, taking some 20 m off the mountain's height (McSaveney *et al.*, 1992). The avalanche was estimated to contain 14 million cubic meters of rock, and flowed 7.3 km to cross the trunk of the Tasman Glacier and swash up the moraine wall on the far side. In its descent, each cubic metre of rock picked up another 3–4 cubic metres of ice and snow and reached speeds of 400–600 km per hour. The avalanche delivered a mantle of rock and snow to the already debris-covered trunk of the glacier. It is not expected that the event will have any lasting effect on the glacier balance.

Another rock avalanche fell from Mt. Fletcher onto the Maud Glacier in the Godley Valley, on the night of 2–3 May 1992 (McSaveney, 1992). This fall was of a lesser volume of 5–10 million cubic metres of rock and it also travelled over a debris-covered glacier trunk. However, the bulk of the deposit ended in the pro-glacial lake at the glacier front. Later, on 16 September 1992, a second rock avalanche fell from the same mountain, letting loose approximately the same volume of rock.

10.5 GAPS AND NEEDS

The most urgent need is for data for modelling glacier/climate relationships. In New Zealand, there is a particular need for ice thickness and terminus ablation data and for a programme of long-term mass-balance monitoring.

10.6 SUGGESTED FUTURE DEVELOPMENT OF MONITORING ACTIVITY

Regarding future glacier monitoring activity in East Africa, it would serve no purpose to propose ambitious schemes. Other glacier regions have witnessed the demise of long-term monitoring programmes. The observation programme on Mount Kenya, now in its seventeenth consecutive year, is the only such contribution for the tropical half of the Earth. It

would be a fair enough accomplishment to see to it that this effort continues into the third millennium.

In New Zealand, the glacier snow-line altitude (i.e. ELA) monitoring programme should continue, with supplementary data for modelling from a proposed radio-echo-sounding programme for ice-depth data. This activity should be accompanied by photographing of selected glacier front positions and pro-glacial lake development during snow-line monitoring flights.

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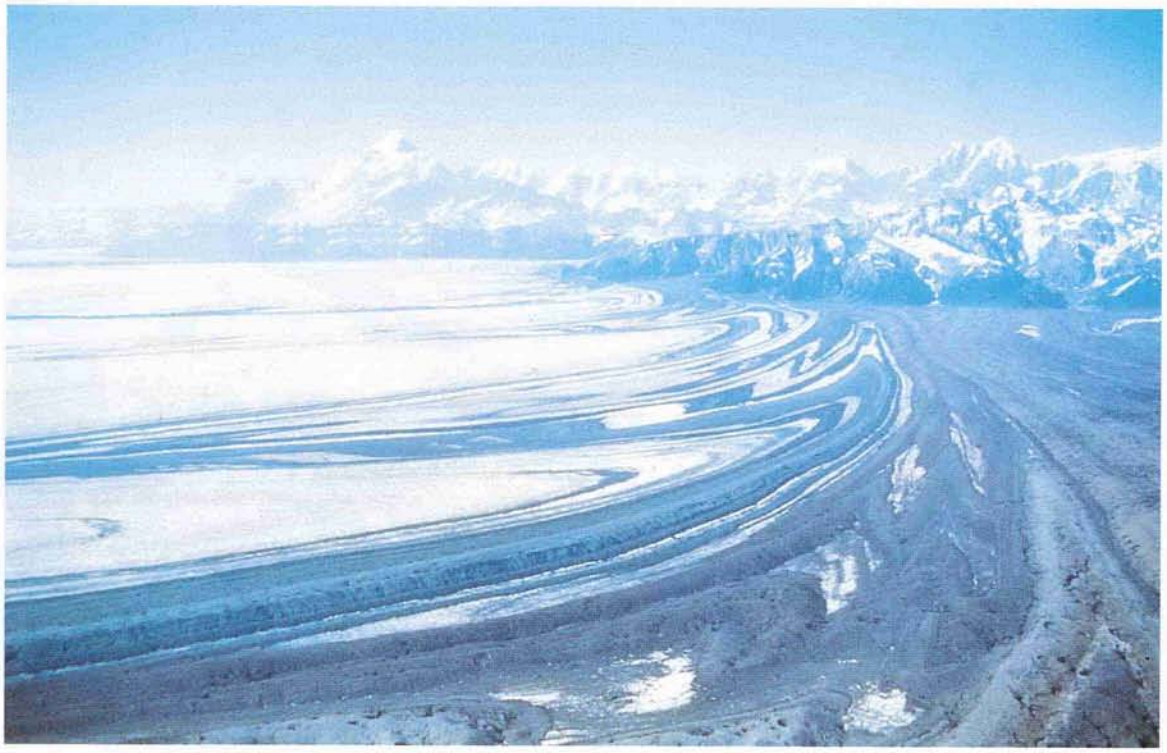
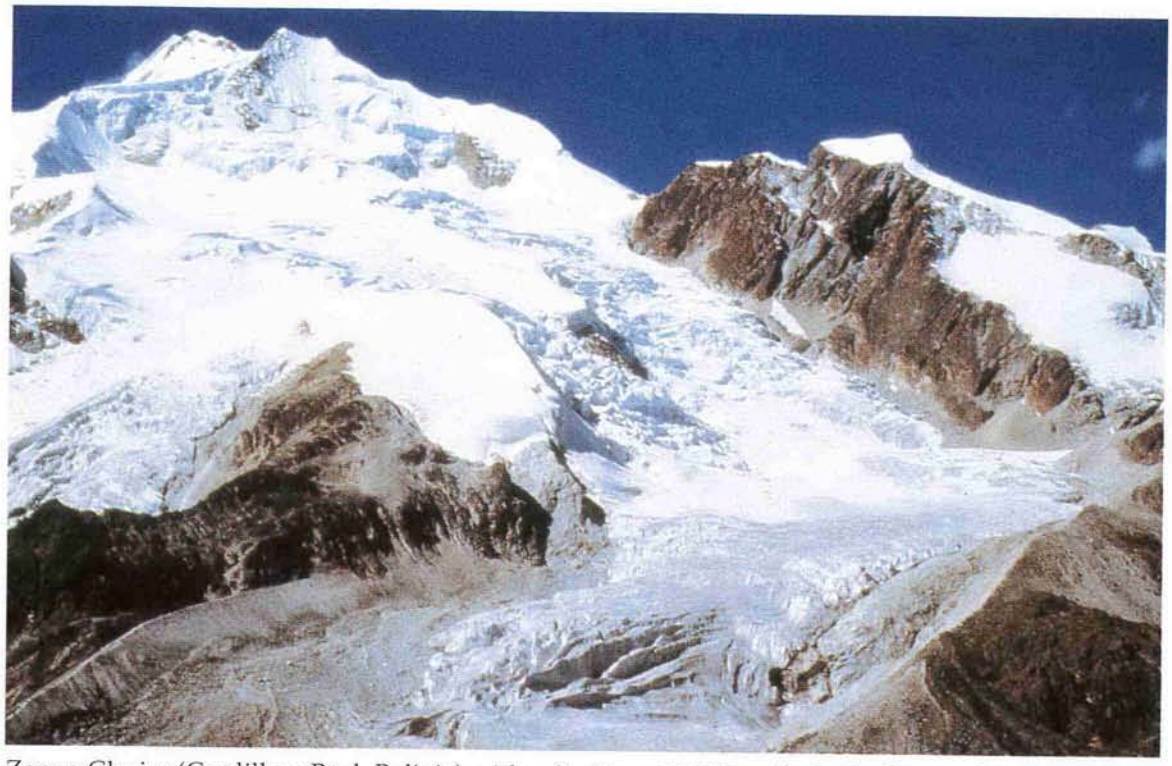
Aerial view of White Glacier (Axel Heiberg Island, Canadian Arctic).
Photo: C. S. L. Ommanney



Portion of the ablation area of the Seward-Malaspina Glacier, Alaska/Yukon Territory. This glacier system has an area of about 5,200 km². Note the repeated medial moraine structures due to surges occurring at intervals of about 40 years. The prominent peak in the background to the left is Mount Saint Elias, which rises from sea level to an altitude of 5,489 m. In the background to the right is the Mount Logan massif, the second-highest peak in North America (6,050 m). *Photo: M. F. Meier, 1985*

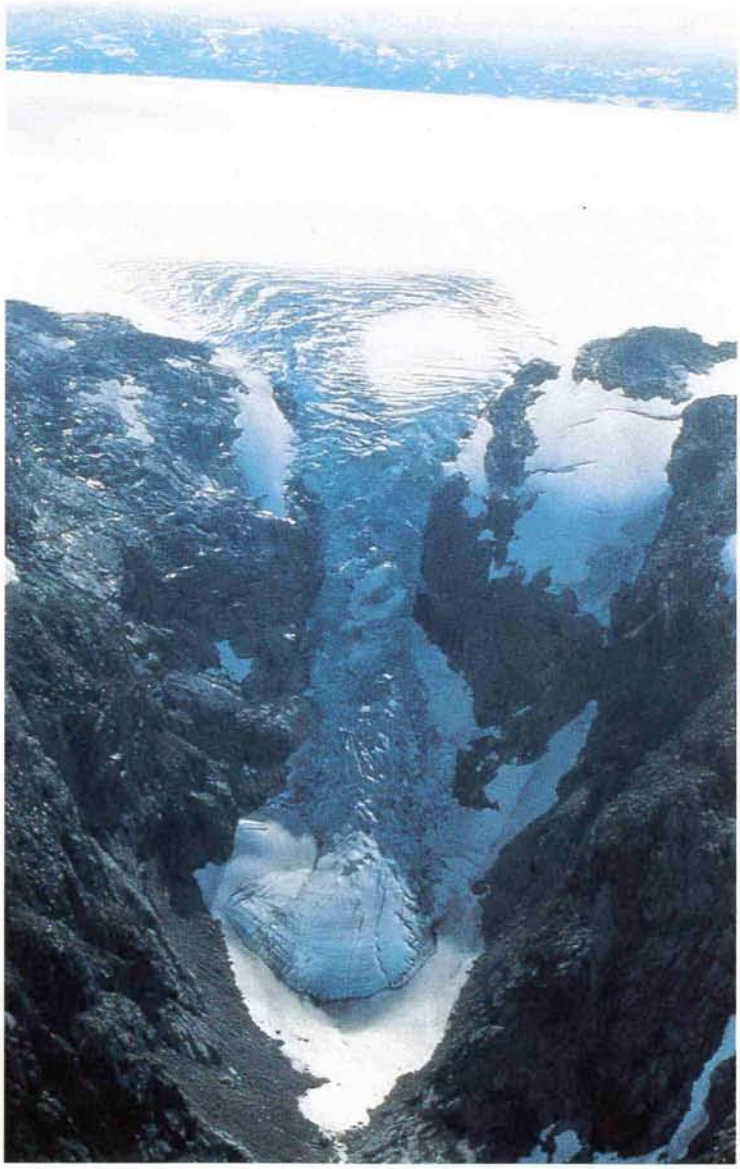
Snow- and ice-capped volcano Nevado del Ruiz, 5,320 m (Central Range, Colombia).
Photo: L. Guarnizo, 1994



Zongo Glacier (Cordillera Real, Bolivia) with a firnline at 5,200 m. In the background to the right is Huayna Potosí (6,088 m). *Photo: B. Francou, April 1995*

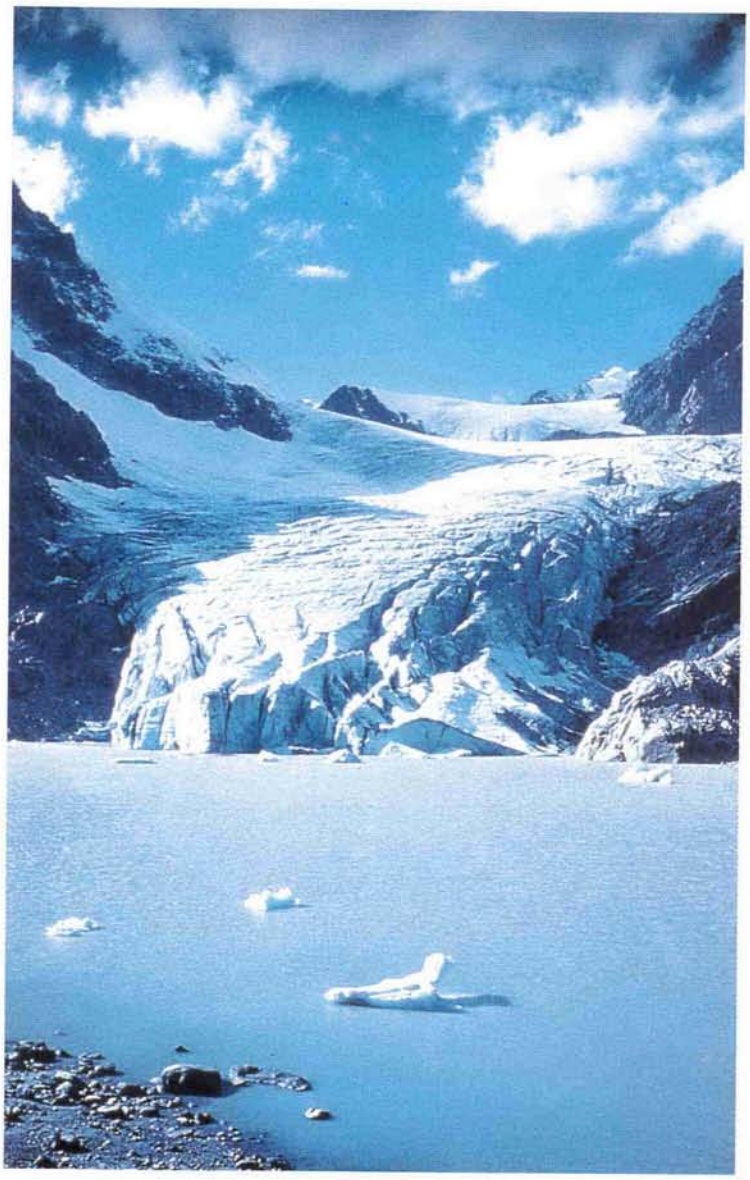
View towards south-east on volcano Eyjafjallajökull (Iceland) with a 78 km². ice cap of the same name. The summit of the volcano reaches 1,666 m a.s.l. The last eruption was in 1821–1823. The outlet glacier has advanced about 350 m in the last 25 years diminishing the lake area to one-third. The lake level is at 180 m a.s.l. *Photo: O. Sigurdsson*

Jemenialovbreen (Svalbard, Norway). *Photo: J. O. Hagen*



Outlet glacier of Folgefonni Ice Cap (Norw
Photo: A. Tvede

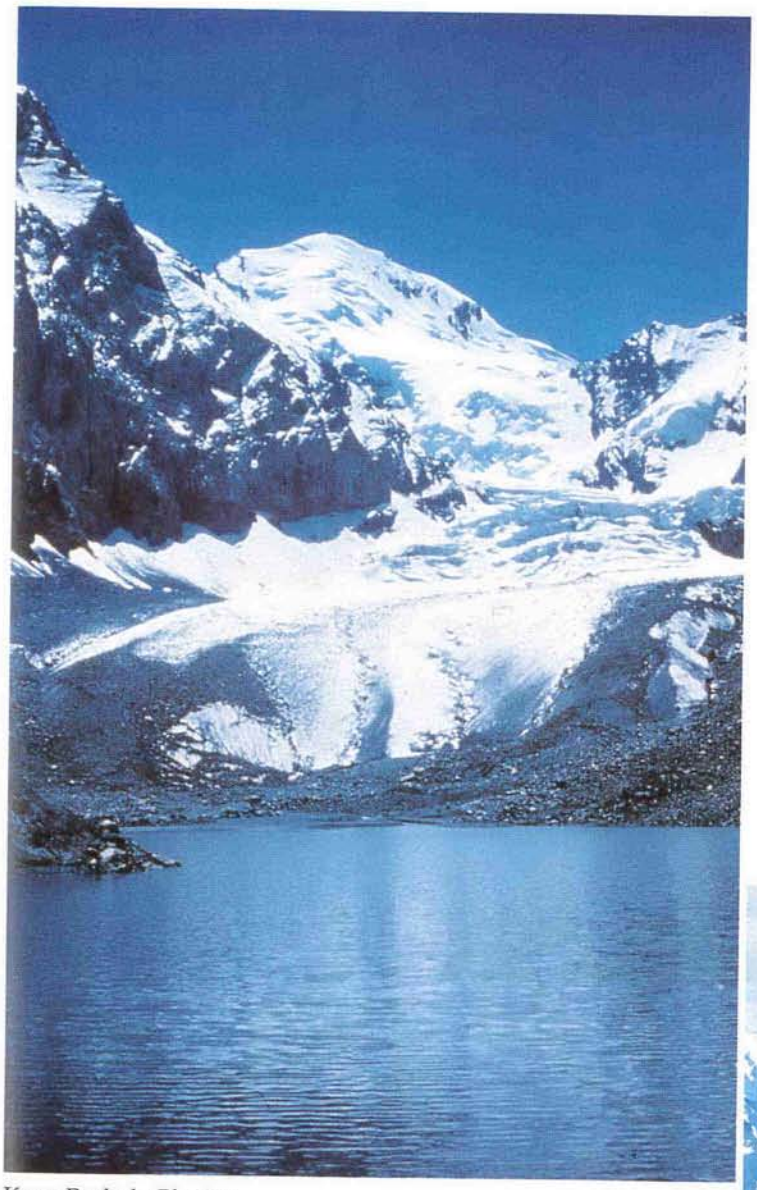
The Forni Glacier (Ortles-Cevedale Group, Italian Alps) is the largest valley glacier in the Italian Alps. *Photo: C. Smiraglia, 1989*



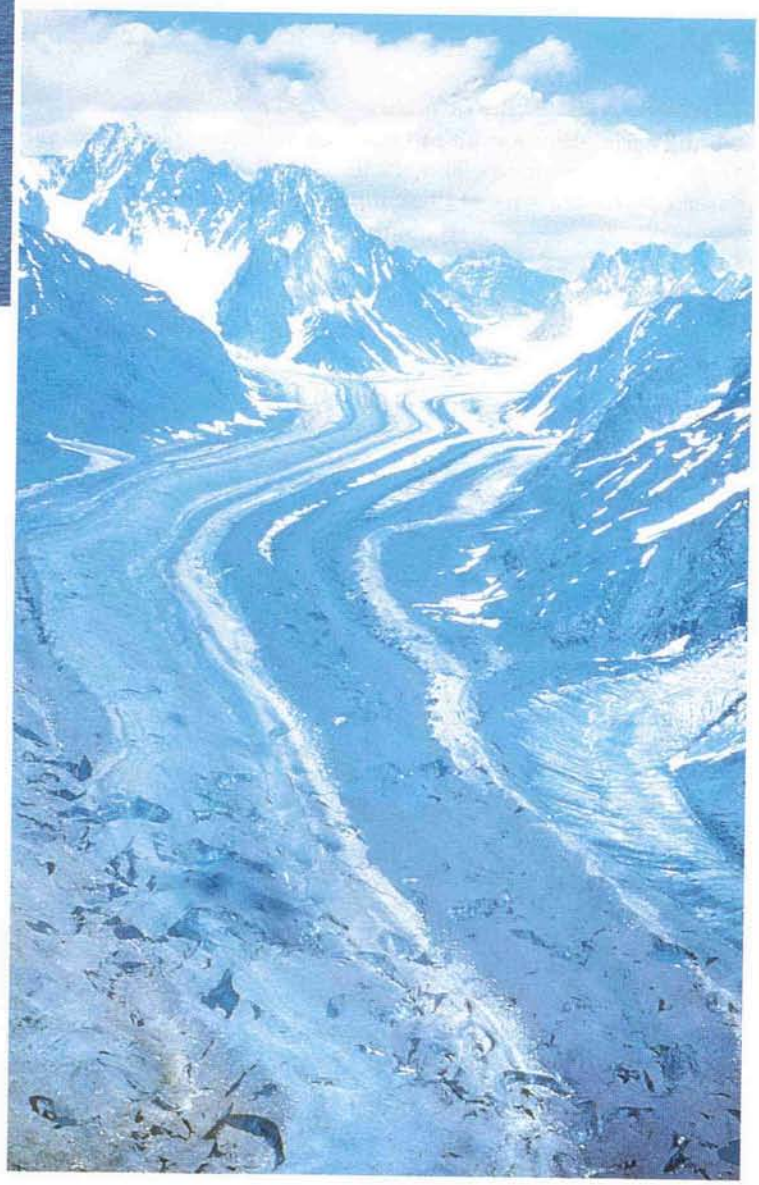
Bolshoy Dalar Glacier, Northern Caucasus, Kuban' River (1.5 km^2). *Photo: K. Rototaev, 1987*

Tongue of Franz Josef Glacier (New Zealand). *Photo: R. Zergenyi, December 1994*

Lewis Glacier with Mount Kenya (5,202 m) (Kenya). *Photo: S. Ardito*

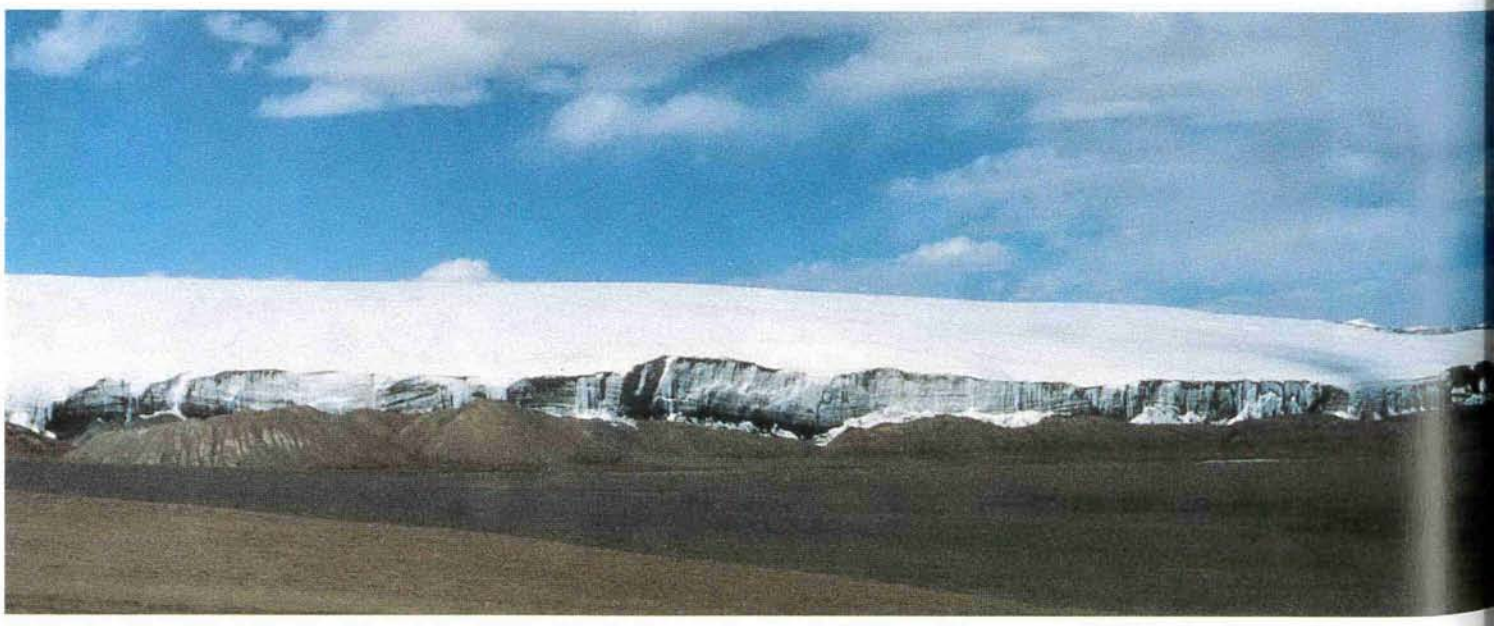


Kara-Batkak Glacier,
Tien Shan.
Photo: K. Rototaev, 1984



Zeravshansky Glacier,
Hissaro Alai.
Photo: K. Rototaev, 1981

Oblique aerial photograph of the expanded foot of an outlet piedmont glacier from a local ice cap in the central part of Nares Land (north Greenland). The glacier is viewed from the northwest. The well-defined margin and the smooth surface are characteristics of many of these outlet glaciers in north Greenland. *Photo: J. Lautrup, 1984*



Guliya Ice Cap, West Kunlun Mountains (China) is the largest ice cap in China with an area of 376 km^2 and a maximum altitude of 6,700 m. *Photo: Huang Zhong, 1992*

11 Glaciers in Asia

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11.1 INTRODUCTION

The presently glacierized area in Asia is estimated at some 185,000 km^2, with the bulk of this ice mass being found in the Commonwealth of Independent States (CIS, i.e. in the territory of the former Soviet Union), China and India/Pakistan. Considerable ice masses also exist in Bhutan, Nepal and Afghanistan, whereas glacier volumes are small in Turkey, Iran, Mongolia and Indonesia. The present chapter illustrates conditions in the CIS, in China, in the Himalayas and in Mongolia [which, in fact, is missing in IAHS(ICSI)/UNEP/UNESCO, 1989].

11.2 DISTRIBUTION AND CHARACTERISTICS

The territory of the CIS contains nearly 30,000 glaciers, with a total area of about 78,000 km^2 representing an estimated ice volume of some 16,500 km^3 of water (Table 11.1). Glaciers are widely distributed over the territory, from the Arctic Islands in the north to the southern spurs of the Vakhan Range (Tadjikistan) in the south and from the Khibini Mountains in the west to the Chukot Peninsula in the extreme north-east (Fig. 11.1). They exist under arctic, subarctic, temperate and subtropical climatic conditions (Krenke, 1982); they produce icebergs in the Arctic ocean, descend below the timberline in the Altai and Caucasus mountain ranges and lie at altitudes of more than 5,000 m a.s.l. in desert mountains of the eastern Pamirs.

Such variability of natural conditions causes a great range of morphological types, sizes and

TABLE 11.1 Distribution of glaciers in the territory of the former Soviet Union (CIS)

Region	*Number of glaciers*	*Glacierized area (km²)*	*Ice volume (km³)*	*Elevation (m a.s.l.) of glaciers max*	*min*	*of firn line*
ARCTIC ISLANDS	2,085	56,126	14,962	1,500	0	150–1,050
Novaya Zemlya	685	23,645	8,100	1,500	0	180–820
Franz Josef Land	995	13,735	2,100	620	0	150–350
Severnaya Zemlya	287	18,326	4,700	905	0	300–800
Others	118	420	62	1,070	0	180–1,050
URALS, KHIBINI	147	29	1	1,400	400	550–1,000
MOUNTAINS OF SIBERIA	3,863	1,699	74	4,510	100	320–3,850
North–eastern Siberia	2,118	706	33	3,140	100	320–2,600
South Siberia	228	66	2	3,570	1,200	2,000–3,400
Altai, Saur	1,517	927	39	4,510	1,970	2,100–3,850
KAMCHATKA	405	874	49	4,600	410	600–2,900
CAUCASUS	2,092	1,428	75	5,642	1,710	2,000–4,000
HISSARO–ALAI	3,890	2,328	135	5,560	2,400	3,100–4,920
PAMIRS	6,729	7,493	660	7,120	2,320	2,950–6,720
Zaalaiskiy range	550	1,329		7,120	3,200	3,920–6,720
Western and Central Pamirs	2,187	3,480		7,000	2,320	2,950–6,260
South Eastern Pamirs	1,873	1,310		6,700	3,420	4,460–5,680
Eastern Pamirs	2,119	1,374		6,110	3,900	4,330–5,680
TIEN SHAN	7,590	7,311	530	6,920	2,720	3,240–5,200
Western Tien Shan	1,456	649		4,840	2,760	3,240–4,490
Nothern Tien Shan	1,695	1,493		5,020	2,780	3,520–4,390
Inner Tien Shan	3,732	3,482		5,980	3,020	3,630–5,200
Central Tien Shan	707	1,687		6,920	2,720	3,600–5,160
DZHUNGARSKIY ALATAU	1,369	1,000	43	4,560	2,720	3,190–4,060
TOTAL	28,170	78,288	16,529	7,120	0	150–6,720

1. *Other Arctic Islands:* Victoriya, Ushakov, De-Long, Vrangel
2. *North-eastern Siberia:* Byrranga, Putorana, Orulgan, Kharaulakh, Suntar Khayata, Chersky range, Koryak Upland
3. *Southern Siberia:* Sayan, Kuznetsky Alatau, Kodar, Barguzin range
4. *Kamchatka:* Sredinniy range, Eastern Kamchatka
5. *Western and Central Pamirs:* Akademii Nauk, Petra 1, Darvazskiy, Vanchskiy, Yazgulemskiy, Severniy Tanymas ranges
6. *South-Eastern Pamirs:* Rushanskiy, Shugnanskiy, Shakhdar'inskiy, Iskhashimskiy ranges
7. *Eastern Pamirs:* Muzkol, Zulumart, Sarykolskiy, Severo-Alichurskiy Yuzhno-Alichurskiy, Vakhanskiy ranges
8. *Western Tien Shan:* Talasskiy, Pskemskiy, Ugamskiy, Chatkalskiy, Ferganskiy *et al.*, ranges
9. *Northern Tien Shan:* Kirgizskiy range, Zailiyskiy Alatau, Kyungey Ala-Too
10. *Inner Tien Shan:* Terskey Ala-Too, Kyooluu Too, Akshiyrac, Kakshaal-Too *et al.*, ranges
11. *Central Tien Shan:* Sary-Dzhas, Tengry-Tag, Engil'chek-Too, Meridional'niy ranges.

regimes of contemporary glaciers. Glacierization of islands in the Arctic is presented mainly by ice caps and outlet glaciers; the ice cover of the northern island of Novaya Zemlya, for instance, has a length of 413 km and maximal width of 95 km. The small corrie glaciers in the Urals, of which Igan Glacier (1.25 km²) is the largest, chiefly reflect redistribution of snow by wind. Small corrie glaciers also characterize the Siberian Mountains. Glaciers on Kamchatka relate to the monsoon climate and active volcanism.

Large centres of contemporary glacierization are situated in the Elbrus and Kazbek massifs of the Caucasus, in the Matcha mountain group of the Hissaro-Alai, in the Central Pamirs where there are a great number of large glaciers, in the region of Pobeda and Khan-Tengri peaks of the Tien Shan and in the Byelukha mountain massive of the Altai. Besides large dendritic glaciers like Fedchenko (length 77.0 km, area 649.6 km²) in the Pamirs or Engil'chek (length 60.5 km, area 567.2 km²) in the Tien Shan, practically all types of mountain glaciers – from complex valley glaciers to glaciers of flat tops, small corrie and hanging glaciers – exist in the mountain regions of the CIS.

There are presently about 45,000 glaciers in China, with a total area of 58,000 km² (Table 11.2). The largest ice volumes are concentrated in the high mountain ranges of central Asia (Tien Shan, Pamirs, Karakoram, Himalaya etc.) and on the Tibetan Plateau (Kunlun, Tanggula, Gangdisi, Nianqintanggula Shan), with 31 glaciers being more than 100 km² in area. Equilibrium line altitude increases from 2,800 m a.s.l. in the Altay Mountains to more than 6,250 m a.s.l. in the Himalayas. On the Tibetan Plateau, equi-

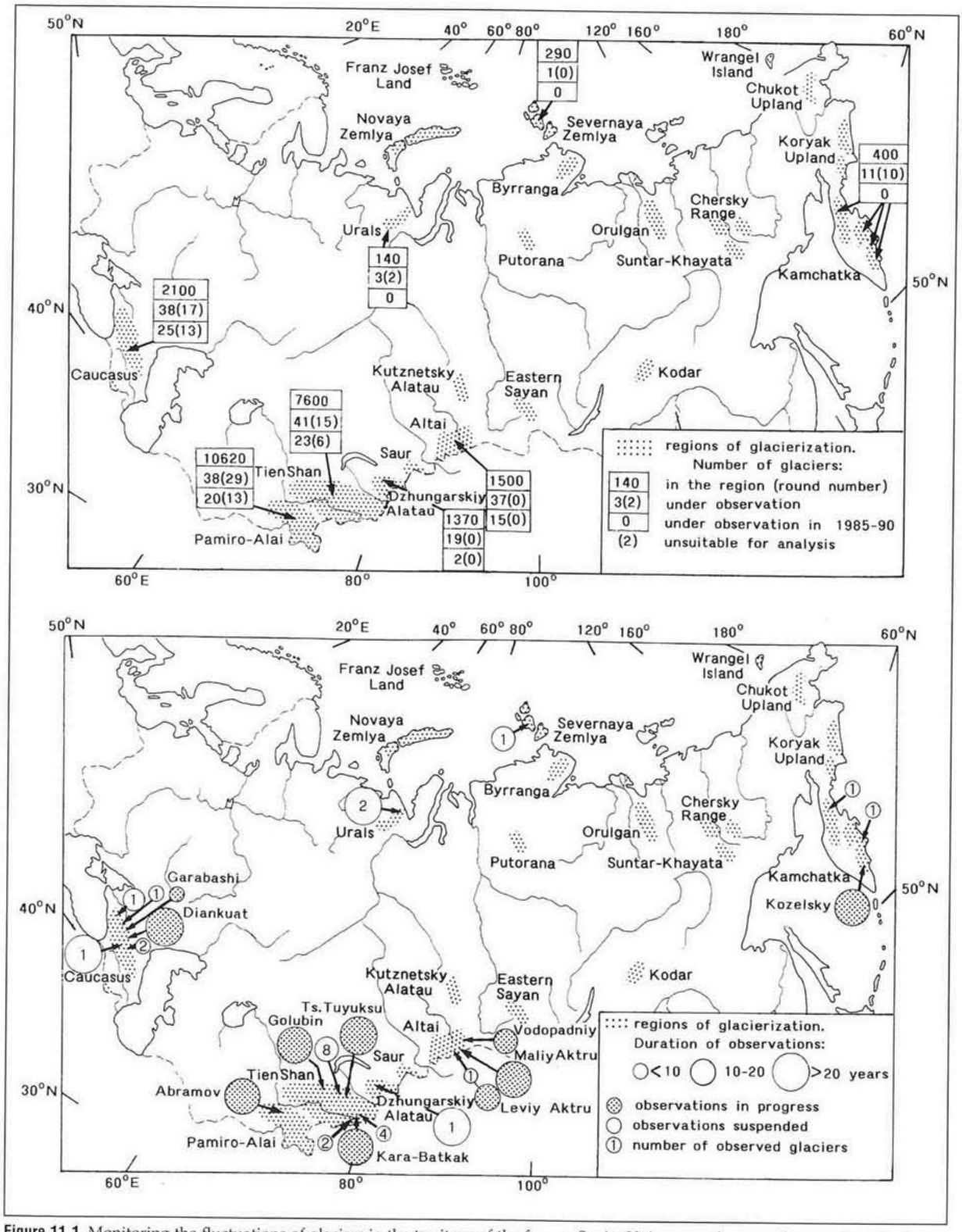

Figure 11.1 Monitoring the fluctuations of glaciers in the territory of the former Soviet Union over the past 40 years. Length-change observations (top) and mass-balance observations (bottom).

librium line altitude also rises from 4,200 m a.s.l. in the humid south-east to 6,200 m a.s.l. in the dry north-west (Wang and Yang, 1992).

The distribution of glaciers in the Himalayas is shown in Fig. 11.2 taken from Fujii (1977), as based on Field (1975), whose main source was the series by the U.S. Army Map Service (AMS, 1:1,000,000: 1945–1962; 1:250,000: 1958–1959). Glacier distribution in Nepal obtained from Landsat images is shown in Fig. 11.3 (Higuchi *et al.*, in press). Area distribution of perennial ice and snow cover in the Himalayan regions was compiled from various maps and references by Fujii and Watanabe (1983), as shown in Table 11.3. However, the accuracy of estimated areas is limited, since only a few regions were covered with exact mapping.

Characteristics of glacier distribution in the Himalayas can be seen in the difference of termini

TABLE 11.2 Glaciers in various glacierized areas of China

Mountainous Region (1)	*Mountainous Area (km²)*	*Highest Peak (m a.s.l)*	*Number of Glaciers*	*Glacierized Area (km²)*	*Ice Volume (km³)*	*Snow-line (m a.s.l)*	*Rel. Glacier Area (%)*	*Mean Glacier Area (km²)*	*Glacierized Ratio (%)*
Altay	28,800	4,374	403	280	159	2,800–3,350	0.48	0.69	0.97
Tarbagatay	4,400	3,835	21	17	7	3,310–3,380	0.03	0.80	0.38
Tien Shan	211,900	7,435	9,128	9,257	10,122	3,620–4,450	15.76	1.01	4.37
Pamirs (2)	23,800	7,649	1,823	2,623	2,176	4,400–5,860	4.47	1.44	11.10
Karakoram (3)	26,600	8,611	1,926	4,769	6,114	4,900–6,010	8.12	2.48	15.64
Kunlun	478,100	7,167	7,615	12,263	12,857	4,500–6,080	20.87	1.61	2.56
Altun	56,300	6,295	229	266	151	4,800–5,360	0.45	1.16	0.47
Qilian	132,500	5,827	2,815	1,930	935	4,300–5,160	3.29	0.69	1.46
Qiangtang Plateau (4)	441,900	6,822	2,283	3,355	2,739	5,700–6,200	5.71	1.45	0.76
Tanggula Range (5)	141,300	6,621	1,570	2,206	1,264	5,300–5,800	3.76	1.41	1.56
Gandise Range (5)	158,300	7,095	2,982	1,615	635	5,000–6,200	2.75	0.54	1.02
Nyainqentanglha Range (5)	110,600	7,162	3,900	7,536	4,900	4,200–5,700	12.83	1.93	6.81
Himalayas (6)	202,500	8,848	9,000	11,000	10,000	5,400–6,250	18.73	1.22	5.43
Hengduan	356,300	7,556	1,680	1,618	1,070	4,600–5,100	2.75	0.96	0.45
Total	2,373,300	8,848	45,375	58,735	53,129	2,800–6,250	100.00	1.29	2.47

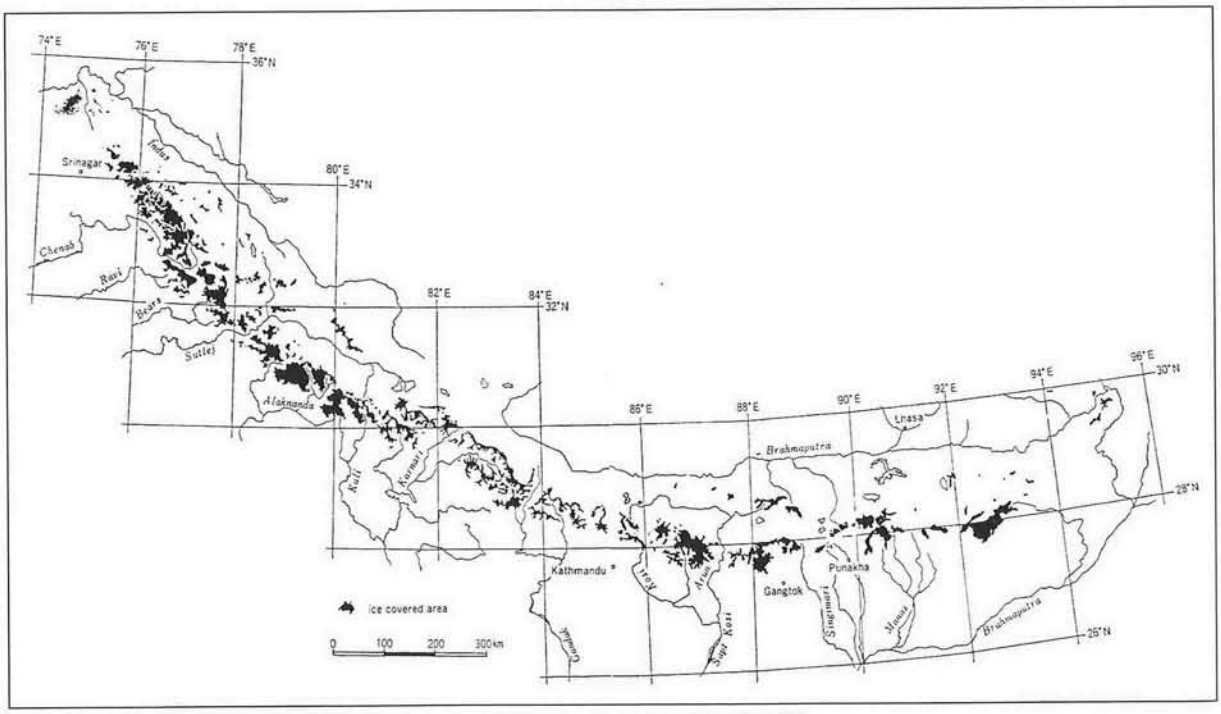

Figure 11.2 Glacier distribution of the Himalayas (Fujii, 1977 on the basis of Field, 1975).

altitudes, as shown in Fig. 11.4 taken from Fujii (1977). The termini altitudes are low in the western Himalayas as they are located at higher latitudes than those in the central and eastern Himalayas. Distinct differences can also be seen between the termini altitudes on the south slope and those on the north slope: lower altitudes of the termini on the south slope of the Himalayas reflect the higher accumulation due to the Indian monsoon.

The effect of the monsoonal precipitation provides much accumulation in summer when strong ablation occurs simultaneously. These glacier regimes have been called the summer-accumulation type. Their characteristics have been described by Ageta and Higuchi (1984).

Most of the glaciers in Mongolia are located in relatively high, uninhabited mountain areas very difficult to get to. The influence of glaciers on human life primarily relates to the processes concerning glacial meltwater, which provide essential freshwater for the local population and livestock and for the development of agriculture and hydropower planning in the western part of Mongolia. The country has 38 mountains with 180 glaciers at altitudes

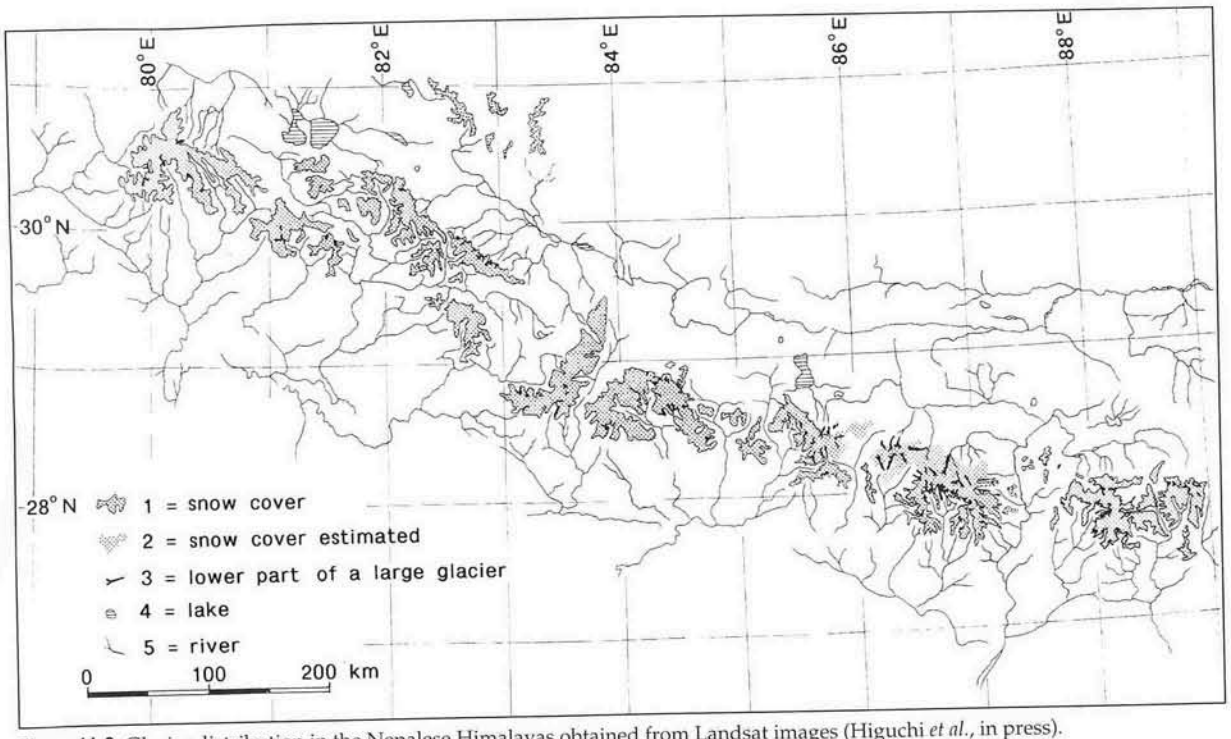

Figure 11.3 Glacier distribution in the Nepalese Himalayas obtained from Landsat images (Higuchi *et al.*, in press).

TABLE 11.3 Perennial ice and snow cover in the Himalayas

Region	*Area (km²)*		
Indus Basin	(700)		
Jhelum Basin	370		
Chenab Basin	3,820	Punjab Himalaya	6,210 km²
Ravi, Bears Basin	720		
Sutlej Basin	600		
Sutlej Basin	5,790		
Jemuna Basin	340	Kumaon Himalaya	13,320 km²
Ganga and Kali Basins	7,190		
Karnari Basin	460		
Gandaki Basin	510	Nepal Himalaya	2,620 km²
Kosi Basin	850		
Tibetan Side	(800)		
Tista Basin	630	Sikkim Himalaya	630 km²
Bhutan Himalaya	2,730		
Northern Assam	2,060	Assam Himalaya	5,180 km²
Namcha Barwa Massif	390		
Whole Himalayas			27,960 km²

Source: Fujii and Watanabe, 1983

between 2,800 and 4,600 m a.s.l. The glacierized surface area is 530 km². Around 80% of the glaciers are situated in the western part of Mongolia, between 87° 40′ and 71°E. Approximately 62% of the glaciers are small, with a total surface area of 42 km², which is only 8% of the total glacierized area. There are only 11 glaciers with a surface area exceeding 10 km² in Mongolia but their total surface area represents more than half the total glacierized surface area of the country. The largest glacier, Potanin, with a surface area of 53 km², is located in the western part of the country.

Glaciers in Iran, Turkey and Afghanistan exist under rather dry subtropical conditions, whereas the small Indonesian glaciers belong to the wet tropical climate type.

11.3 EXISTING INVENTORIES

As early as the beginning of the 20th century, the inventorying of glaciers started in the Russian Empire, on the initiative of the Russian Glacier Commission. The first inventory of glaciers was compiled in the

TABLE 11.4 (contd.)

Name of Glacier	*PSFG Nr. (SU)*	*Length (km)*	*Total Period of Observations*	*Number of Observations*	*Total ΔL (km)*	*Other Sources*
ALTAI						
Bolshoy Maashey	07104	8.2	1924–1990	15	–0.3±?	Soviet Geophysical Committee 1988
Gebler	–	9.0	1833–1985	19	–1.6	Soviet Geophysical Committee 1988
Korumdu	07103	4.7	1936–1990	24	–0.3	
Maliy Aktru	07100	4.4	1911–1990	37	–0.3	
Praviy Aktru	07101	5.3	1936–1980	15	–0.2	Soviet Geophysical Committee 1988
Rodzievich	–	7.8	1850–1986	15	–1.9	Soviet Geophysical Committee 1988
Sapozhnikov	–	10.5	1850–1986	10	–1.6	Soviet Geophysical Committee 1988
KAMCHATKA						
Erman	–	16.5	1935–1986	7	+2.3	Soviet Geophysical Committee 1991
Kozelskiy	08005	4.6	1948–1990	11	+0.7	Soviet Geophysical Committee 1991

Conventional signs:
* = surging glacier
8.0.11 (in column 'Other sources') = volume, part, issue of glacier inventory of the USSR (Gidrometeoizdat 1967–1982)
±? = gaps in observations: Aksu Zapadniy 1941–1958; Bolshoy Maashey 1974–1980

Notes:
1. Length of glacier is given by glacier inventory of the USSR (Gidrometeoizdat 1967–1982)
2. There are additional qualitative estimations of ΔL: Mayli 1835–1879: +X; Fedchenko 1878–1913: +X, 1914–1926: –X; Mushketov 1904–1914: +X
SOURCES: IAHS(ICSI)/UNEP/UNESCO 1988, 1993; IAHS(ICSI)/UNESCO, 1967, 1973, 1977, 1985; and others.

TABLE 11.5 Coverage of different glaciated regions of the former Soviet Union (CIS) by length change observations since the 1950s.

Glaciated region	*Number of glaciers with observation time series of 25–40 and 10–25 years*			
	up to 1990 (inclusive)		*suspended before 1990*	
	25 – 40	*10 – 25*	*25 – 40*	*10 – 25*
Polar Urals	–	–	[3]	–
Caucasus	13 + (3)	3 + (2)	2 + (9)	–
Pamiro-Alai	10 + (6) + [1]	3 + [1]	1 + (2)	2
Tien Shan	9	4 + (1)	2 + [187]	1
Dzhungarskiy Alatau	–	1	–	–
Altai	4	1 + (1)	(1)	1 + (1)
Kamchatka	1	–	1	–
Total	37 + (9)+ [1]	12 + (4)+ [1]	6 + (12)+ [190]	4 + (1)

Designation: 10 = number of glaciers with continual observations at intervals of 1 to 5 years
(6) = as above with gaps in observations of more than 2 years
[1] = as above but from repeated maps

In the Nepalese Himalayas, a pilot study for an inventory of glaciers in the Mount Everest region had been prepared by Müller (1970). An inventory of the Dudh Kosi region, including the region covered by Müller's inventory, was later compiled by Higuchi *et al.*, (1978, 1980) but the full data set of this inventory has not been published. Brief glacier inventories of Langtang Valley were reported by Iida *et al.* (1984) and Zheng *et al.* (1984). A detailed glacier inventory of the Langtang Valley was compiled by Shiraiwa and Yamada (1991). Higuchi *et al.* (in press) provide a report on glacier distribution of the Nepalese Himalayas and the Karakoram from Landsat images. Detailed inventories are also available from the Panjshir and Safed Kirs regions in Afghanistan, from the Nanga Parbat and Chitral regions in Pakistan and from the Baspa River basin in India (IAHS(ICSI)/UNEP/UNESCO, 1989).

In Mongolia, a preliminary glacier inventory was compiled on the basis of the existing topographic

TABLE 11.6 Long-term mass balance observations in the former Soviet Union (CIS)

Glacier Name	*PSFG No.*	*Country*	*Location*	*Area (km²)*	*Length (km)*	*Exposition*	*Period of Observation*	*No. of bal. years*	*Mean Annual for the Whole Observation (g/cm² year)* b_w/c_t	b_s/a_t	b_n
No. 104 (Vavilov)	01001	Russia	Severnaya Zemlya	1,817.0	60.0	ice cap	1974–1988	14	(28)°	(-34)°	(-6)
Obruchev	02002	Russia	Polar Urals	0.3	0.8	E	1957–1981	24	267.5°	–280.9°	–13.4
IGAN	02001	Russia	Polar Urals	0.7	1.2	E	1957–1981	24	232.8°	–251.5°	–18.7
Marukhskiy	03001	Russia	Western Caucasus	3.3	4.0	N	1966–1982	16	223.4°	–277.0°	–53.6
Khakel	03033	Russia	Western Caucasus	2.7	3.9	N	1959–1966	7	-	–	+12.4
							1973–1979	6	137.3°	–141.8°	–4.5
Garabashi	03031	Russia	Elbrus	4.5	5.8	S	1986–1993	7	128.8^	–111.4^	+17.4
Djankuat	03010	Russia	Central Caucasus	3.1	4.2	N	1967–1993	26	227.8°	–233.6°	–5.8
Tbilisa	03012	Georgia	Central Caucasus	3.8	3.0	SE	1967–1980	13	215.2°	–238.3°	–23.1
Abramov	04101	Kirghizstan	Pamiro-Alai (Alaiskiy R.)	26.2	9.4	N	1967–1993+	26	157.3^	–211.8^	–47.7
Ts. Tuyuksuiskiy	05075	Kazakhstan	Tien Shan (Zail. Alatau)	2.8	3.2	N	1956–1993	37	99.0^	–135.1^	–36.1
Zoya Kosmodemyanskaya	05092	Kazakhstan	Tien Shan (Zail. Alatau)	0.4	1.2	NE	1964–1980	16	98.5^	–113.8^	–15.3
Manshuk Mametova	05091	Kazakhstan	Tien Shan (Zail. Alatau)	0.4	0.6	W	1956–1964	8	–	–	+11.2
							1964–1980	16	96.0^	–133.4^	–37.4
Mayakovskiy	05094	Kazakhstan	Tien Shan (Zail. Alatau)	0.2	0.8	W	1964–1980	16	96.3^	–100.8^	–4.5
Molodezhniy	05090	Kazakhstan	Tien Shan (Zail. Alatau)	1.4	1.7	NE	1956–1964	8	–	–	+12.2
							1964–1980	16	98.2^	–150.8^	–52.6
Ordzhonikidze	05093	Kazakhstan	Tien Shan (Zail. Alatau)	0.3	1.2	W	1964–1980	16	96.6^	–107.4^	–10.8
Partizan	05095	Kazakhstan	Tien Shan (Zail. Alatau)	0.1	0.8	W	1964–1980	16	88.8^	–56.3^	+32.5
Visyachiy	05096	Kazakhstan	Tien Shan (Zail. Alatau)	0.3	0.6	NE	1964–1980	16	100.8^	–135.6^	–34.8
Igli Tuyuksu	05076	Kazakhstan	Tien Shan (Zail. Alatau)	1.7	2.2	NW	1956–1964	8	–	–	+9.8
							1964–1980	16	98.1^	–134.9^	–36.8
Kara-Batkak	05080	Kirgizstan	Tien Shan (Terk. Alatau)	4.6	3.5	N	1956–1993+	37	59.3^	–103.3^	–43.1
Golubin	05060	Kirgizstan	Tien Shan (Kirghiz. R.)	6.2	5.1	NW	1971–1993	22	63.6°	–99.5°	–35.9
Shumskiy	06001	Kazakhstan	Dzhungarskiy Alatau	2.8	3.6	N	1966–1991	25	60.0°	–73.2°	–13.2
Maliy Aktru	07100	Russia	Altai (Sev.-Chuiskiy R.)	3.8	4.4	NE	1961–1993+	32	90.5^	–95.3^	–3.9
Leviy Aktru	07102	Russia	Altai (Sev.-Chuiskiy R.)	6.5	5.9	SE	1976–1993+	17	85.8^	–89.7^	–3.1
Praviy Aktru	07101	Russia	Altai (Sev.-Chuiskiy R.)	4.8	5.3	NE	1979–1990	11	90.4^	–76.7^	+13.7
No. 125 (Vodopadniy)	07105	Russia	Altai (Sev.-Chuiskiy R.)	0.8	1.4	N	1976–1993+	17	53.5^	–56.1^	–1.8
Kozelskiy	08005	Russia	Kamchatka (Avachinsk. Gr.)	1.8	4.6	S	1972–1993	21	337.0°	–361.8°	–24.8

Conventional signs: ° = b_w, b_s (winter balance, summer balance)
^ = c_t, a_t (total accumulation, total ablation)
+ = components (b_w/c_t, b_s/a_t) are given only to 1991 inclusive

Notes: No. 104 (Vavilov): There is a gap in observations from 1981 to 1985. The approximate data for the whole period of observation are given in brackets.
Kara-Batkak: Balance is received by hydrological method.
Kozelskiy: Data for 1981–1984 and 1985–1987 are calculated.

TABLE 11.7 Special glaciological (large-scale) maps published in the former Soviet Union (CIS)

Title of map, scale, year of survey	*Author, institution, method of survey*	*Character and place of publication*
	POLAR URALS	
1. Obruchev Glacier 1:5,000; 1963	D. G. Tsvetkov Institute of Geography	2 black and white maps (2 lists) Tsvetkov 1970
2. IGAN Glacier 1:5,000; 1963	Russian Academy of Sciences Terrestrial photogrammetry	
	CAUCASUS	
1. Glaciation of Elbrus 1:10,000; 1957–1959	Yu. F. Knizhnikov, A. V. Bryukhanov, F. V. Nikulin, I. A. Moscow University Labutina, V. I. Kravtsova Terrestrial and partly aerial photogrammetry	Colour map (14 lists) Atlas of Elbrus' glaciers Moscow University 1963
	PAMIRO-ALAI	
1. Fedchenko Glacier and other Glaciers in its basin 1:50,000; 1928	R. Finsterwalder (Germany), I. G. Dorfeev (USSR) Terrestrial photogrammetry Soviet-German expedition at the Pamirs in 1928	12 maps Finsterwalder 1932
2. Fedchenko Glacier and other glaciers in its basin 1:50,000; 1958 Tongue of glacier in 1:25,000, the terminus in 1:10,000	G. Ditrich *et al.*(DDR), I. G. Dorofeev (USSR) Glaciological expedition of Academy of Sciences of Uzbek, SSR, 1958 (IGG) Terrestrial photogrammetry	5 colour maps Nationalkomitee für Geodäsie und Geophysik der DDR 1964
3. Abramov Glacier 1:25,000; 1986 a) map of surface relief b) map of glacier bottom c) map of glacier thickness	V. A. Kuzmichenok Kirgizian Aerogeodetic enterprise, Bishkek Yu. Ya. Macheret Institute of Geography Russian Academy of Sciences Aerial photogrammetry and radiolocation	3 colour maps Kuz'michenok *et al.*, 1992
	TIEN SHAN	
1. 'Gletschergebiet Tujuksu' 1:10,000; 1958 Ts. Tuyuksu Glacier and 6 glaciers in its basin	H. Hartmann, M. Simon, J. Toppler (DDR) Soviet-German expedition in 1958 (IGG) Terrestrial photogrammetry	Colour map Petermanns Geographische Mitteilungen 105 1961
2. 'Fluctuations of the Akshyirak Range Glaciers from 1943 to 1977' 1:50,000; 1990 Changes of outlines and surface height of 178 glaciers. Tables of length, area and volume change of all glaciers	V. A. Kuzmichenok Kirgizian Aerogeodetic enterprise, Bishkek Aerial photogrammetry of aerophotosurveys from 1943 and 1977	Colour map Kuz'michenok 1990

TABLE 11.8 Long-term observation of glacier length change in China

Name of Glacier	*Location*	*Period*
Urumqihe S. No. 1	Tianshan Mts.	1962–now
Laohugou No. 12	Qilianshan Mts.	1960–1985
'1st July' Glacier	Qilianshan Mts.	1956–1985
Shuquahe No. 4	Qilianshan Mts.	1956–1984
Hailuoqou Glacier	Hengduan Shan Mts.	1930–now

TABLE 11.9 Glacier mass balance observations in China

Glacier	*Location*	*Period*
Urumqihe S. No. 1	Tianshan Mts.	1959–1967/1979–now
'1st July' Glacier	Qilian Shan Mts.	1974–77/1984–88
Yanlonghe No. 5.	Qilian Shan Mts.	1976–1979
Shiyanghe No. 4.	Qilian Shan Mts.	1962–63,1975–1978
Laohugou No. 12.	Qilian Shan Mts.	1975–1976
Dongkemadi	Tanggula Mts.	1988–now
Meikuang	Tanggula Mts.	1988–now
Kangwure	Tanggula Mts.	1990–now
Hailaoguo	Hengduanshan Mts.	1991–now
Guxian	Nyainqentangula Mts.	1964–1965
Qosiquo	Nyainqentangula Mts.	1975–1976
Qiagyun	Nyainqentangula Mts.	1975

map on a scale of 1:1,000,000, issued in 1972, and using some aerial photography from 1983.

11.4 EXISTING LONG-TERM OBSERVATIONS (LENGTH CHANGES, MASS BALANCE, MAPS)

Direct long-term glacier observations in Russia started in the middle of the 19th century and initially had a rather occasional character. The first attempt to coordinate long-term observations of glaciers in the Russian Empire dates back to 1892, when the first national programme for observing glaciers was published by I. V. Mushketov (1892). From that time on, comparatively systematic visitations of glaciers were made in the Caucasus, in Middle Asia and in the Altai. Many of these glaciers are still under observation today. As a result, length changes over a period of 100 and more years are documented for several tens of glaciers (Table 11.4), reflecting a general and strong secular tendency to retreat.

The regional distribution of glaciers with observations of length changes during the past 40 years is presented in Fig. 11.1a. The number of glaciers under observation accounts only for about 0.3% of the total number of glaciers. It has been reduced by almost a factor of two during the last period under review (1985–1990). Some 40% of all glaciers under observation are heavily debris-covered, have dead tongues, end in lakes, have long response times or show unstable behaviour (surging glaciers) and cannot, therefore, be used in a straightforward way for studies of the interrelation between glacier fluctuations and climatic change. Information about the availability of glaciers with long-term series of length-change measurements in the CIS over the past 40 years is given in Table 11.5.

Direct glaciological mass-balance measurements were widely developed in the former Soviet Union during the IGY (i.e., 1958) and especially during the International Hydrological Decade (1965–1974) when the processes and regularities determining the water, ice and heat balance of mountain glaciers were investigated in selected representative glacierized basins. The results of these investigations were published by Suslov *et al.* (1980) for Abramov Glacier, Galakhov *et al.* (1987) for the Aktru Glaciers, Golubev *et al.* (1978), for Djankuat Glacier, Vinogradov and Muraviev (1992) for Kozelskiy Glacier, Krenke *et al.* (1988) for Marukh Glacier, Gobedzhishvili *et al.* (1986) for Tbilisa (Tbilisi) Glacier, Makarevich *et al.* (1984) for the Tuyuksu Glaciers and by Bochin and Krenke (eds.) [1980, 1987]. The total of these mass balance measurements entered the *Glacier Mass Balance Bulletins* Nos 1 to 3 [IAHS(ICSI)/UNEP/UNESCO, 1991, 1993*b*, 1994], making up nearly half of all the glaciers under observation in the world. The distribution of glaciers with different length rows of mass balance observations is presented in Fig. 11.1 (cf. Table 11.6). It is evident that the north-east of Russia and Pamirs (Tajikistan) are poorly represented for mass balance observations. In the CIS, special attention has always been paid to individual components of the glacier mass balance: winter balance was measured on all glaciers and, in addition, summer accumulation and internal accumulation were determined on most of them. Ten glaciers have been under observation for more than 20 years; Tsentralniy Tuyuksuyskiy in Kazakhstan and Kara-Batkak in Kirghiszstan have the longest series covering almost 40 years. On nine of them, the observations continued up to 1993. The great value of such data is evident and a great effort should be made to continue taking the measurements without interruption.

The national programme of detailed observations of glacier fluctuations was developed by a group of Soviet glaciologists led by P. A. Shumsky (Voloshina *et al.*, 1973). The purpose of this programme was to investigate the causes and mechanisms of glacier fluctuations on the basis of integrated and synchronous observations with regard to individual characteristics of external and internal mass exchange including ice movement, ablation/accumulation, changes of form and dimensions of glaciers. The most valuable data from this programme were received on the Obruchev and Igan Glaciers in the Polar Urals (Shumsky *et al.*, 1972), on Tsentralniy

TABLE 11.10 Glacier maps in China compiled by the Lanzhou Institute of Glaciology and Geocryology

No	*Name of map*	*Scale*	*Method of survey*	*Date (survey)*	*Date (mapping)*
1	Map of Dunde Ice Cap in Qilianshan Mt.	1:50,000 1:25,000	aerial photogrammetry aerial photo and terrestrial photogrammetry	1956 1984	1988
2	The Map of No. 1. Glacier in the Yanlonghe (Qilianshan Mt.)	1:16,000	terrestrial photogrammetry	1977	1980
3	The Map of '1st July' Glacier (Qiyi) (Qilianshan Mt.)	1:12,000	terrestrial photogrammetry	1975	1980
4	The Map of No. 4 Glacier in Shiyanghe (Qilianshan Mt.)	1:7,500	terrestrial photogrammetry	1976	1980
5	The Map of No. 12 Glacier in the Laohugou (Qilianshan Mt.)	1:20,000 1:5,000	terrestrial and aerial photogrammetry terrestrial photogrammetry	1960 1976	1980 1980
6	The Map of Mt. Tomur Glaciers in Tien Shan Mt.	1:200,000	aerial and terrestrial photogrammetry	1979	1980
7	Map of Glaciers at the Source of Urumqi River (Tien Shan Mt.)	1:10,000	plane-table survey	1962	1965
8	Map of Glaciers at the Source of Urumqi River (Tien Shan Mt.)	1:10,000	terrestrial photogrammetry	1973	1983
9	Topographic Map of Glacier No. 1 Tien Shan Station	1:5,000	terrestrial photogrammetry	1980	1982
10	Map of Glacier No. 1 and No. 2 at the Source of Urumqi River, Tien Shan	1:5,000	terrestrial photogrammetry	1989	1989
11	Glacier Topographic Map of Bogda Rerion in Tien Shan Mt.	1:50,000	terrestrial and aerial photogrammetry	1966	1983
12	The map of Batura Glacier (in Pakistan)	1:60,000	terrestrial photogrammetry	1974	1978
13	Map of Glaciers in Mt. Xixiabangma (Himalayas)	1:50,000	plane-table survey	1964	1982
14	Mount Xixiabangma (8012m) (Himalaya Mt.)	1:100,000	aerial photogrammetry	1974	1994
15	Map of Qiag Yan Glacier (in South Tibet)	1:25,000	terrestrial photogrammetry	1976	1979
16	Glacier Topograhic Map of Mt. Qomolongma Regions.	1:25,000	terrestrial photogrammetry	1986	1972
17	Map of Glaciers of Mt. Qomolongma Regions	1:50,000	terrestrial photogrammetry	1977	1980
18	Mount Qomalangma (Sagarmatha) (8848m)	1:100,000	aerial photogrammetry	1974	1991
19	Map of Glaciers of Mt. Gongga	1:25,000	terrestrial and aerial photogrammetry	1982	1985
20	Map of Chongce Glacier in West Kunlun Mt.	1:50,000	terrestrial and aerial photogrammetry	1970 1987	1990
21	Map of Guliya Ice Cap in West Kunlun Mt.	1:30,000	aerial photogrammetry	1970	1992
22	Mount Qogori (K2) (8611m) (Karakorum Mt.)	1:100,000	aerial photogrammetry	1976	1994
23	Kongur Tagh-Mustag Ata (East Pamir)	1:100,000	aerial photogrammetry	1976	1994
24	Map of Snow, Ice and Frozen Ground	1:400,000			1988

(contd) TABLE 11.10

No	Name of Map	Scale	Method of survey	Date (survey)	Date (mapping)
25	Map of Peaks and Glaciers in Qilian Mountains	1:70,000	aerial photogrammetry	1956	1994
26	Map of Peaks and Glaciers in Daxue Mt. (Qilianshan)	1:250,000	aerial photogrammetry	1956	1994
27	Map of Peaks and Glaciers in Lenghong Ridge (Qilianshan Mt.)	1:250,000	aerial photogrammetry	1956	1994
28	Map of Peaks and Glaciers in Southern Sule Mountain	1:250,000	aerial photogrammetry	1956	1994
29	Map of Peaks and Glaciers in Southern Danghe (Qilianshan Mt.)	1:250,000	aerial photogrammetry	1956	1994

Tuyuksuyskiy Glacier in the Tien Shan, on Shumsky Glacier in the Dzhungarskiy Alatau and on surging Medvezhiy Glacier in the Western Pamirs.

Information about long-term changes in glaciers is often obtained by comparing different-time topographic maps. Unfortunately, the quality of glacier representation on the state topographic maps is unsatisfactory. On the initiative of the Soviet Working Group on Glacier Monitoring, some special educational workshops for improving the quality of standard topographic maps of glaciated regions were organized recently in order to foster the necessary collaboration between cartographers and specialists in glaciology.

The most important achievements with regard to large-scale maps compiled by cartographers and topographers with the help of glaciologists include the *World Atlas of Snow and Ice Resources* (in preparation for publication), the *Atlas of Elbrus Glaciers* (published in 1963) and some maps of separate glaciers in the Pamiro-Alai and Tien Shan. A summary of the main published topographic-glaciological maps is given in Table 11.7. An example of effective use of special topographic maps for studying glacier fluctuations of the whole region is the work of cartographer-glaciologist V. Kuzmichyenok (Kuzmichyenok, 1990; 1991). Using for the purposes of comparison two maps on a scale of 1:10,000, compiled in 1943 and 1977, he compounded the map of changes in the Akshyirak Range glaciers (Tien Shan) over 25 years and obtained the changes in 21 parameters for 178 glaciers. The common area of all glaciers shrank over this period from 426.7 km^2 to 411.7 km^2, and ice volume by 3.6 km^3. Numerous unpublished maps and survey materials provide key basic information on long-term glacier fluctuations. Today, they are stored in different organizations of the CIS. Thus, repeated terrestrial and aerial surveys of glaciers in the Polar Urals (6–8 surveys between 1953 and 1981), in Kamchatka (2–3 surveys), at surging Medvezhiy Glacier (some 20 repeated surveys from 1963 to 1990) and on other glaciers are kept in the Institute of Geography of the Russian Academy of Sciences in Moscow.

Long-term observations of length fluctuations and mass-balance measurements were carried out on several Chinese glaciers, as shown in Tables 11.8 and 11.9. The general trend is comparable to European observations and reflects a strong retreat, with a short interruption in the 1970s (Ding and Haeberli, in preparation). Table 11.10 lists about 30 special glacier maps prepared by the Lanzhou Institute of Glaciology and Geocryology.

Retreating termini of several Himalayan glaciers have occasionally been observed since the middle of the 19th century in the Punjab, Kumaun and Sikkim. In the Nepalese Himalayas, Higuchi *et al.* (1980) compared their inventory in 1974–1976 with Müller's inventory in 1955–1963 and found that most glaciers of the debris-free type in Khumbu Himal were in retreat during this period. The retreating speed of some of these glaciers has accelerated since the 1980s (Yamada *et al.*, 1992; Kadota *et al.*, 1993). Data on glacier length change (predominantly retreat) are now also regularly reported from Pakistan, India, Nepal and Indonesia (IAHS(ICSI)/UNEP/UNESCO, 1993*a*).

Most of the Mongolian mountain and valley glaciers seem to be retreating. Annual precipitation near the glaciers is characteristically around or below 1,000 mm (for instance, Turgen Glacier or Tsagaan Deglee: 600 and 900 mm). From south-east to north-west, decreasing precipitation causes equilibrium lines and median glacier elevations to increase from about 2,900 m to 4,100 m a.s.l. Accumulation area ratios also appear to markedly augment with increasing precipitation.

11.5 SPECIAL EVENTS

Glacier-related catastrophic phenomena, such as glacial mudflows, ice avalanches, outbursts of glacier lakes, glacier surges, etc., are widely distributed in the mountain regions of the CIS. In Kazakhstan, five enormous debris flows have taken place over the past 50 years. Three of them were of glacial origin. The debris flow of 1973, for example, was caused by the outburst of a moraine-dammed lake near Tsentralniy

TABLE 11.11 Long-term observations on surging glaciers in the territory of the former Soviet Union (CIS)

Region	*Glacier*	*Date of surge*	*Character of surge*	*Consequences*	*Type of observation*	*Source*
Caucasus	Devdoraki	1784	ice avalanche	destructions of the road, damming the river	observations of the changes of glacier terminus position	Kalesnik 1937
		1808	ice avalanche			Kalesnik 1937
		1817	ice avalanche			Kalesnik 1937
		1832	ice avalanche			Kalesnik 1937
		1842	adv. of 3.2 km			Glacier... 1977
		1885	ice avalanche			Kalesnik 1937
	Kolka	1902	advance of 12 km,water-ice -stone mudflow advance of 4.6 km	destructions of communications network	geomorphological investigations	Rototaev *et al.*, 1983
		1969			complex ground-based observations during the surge and after it	
Pamirs	Medvezhiy	1916	adv. of glacier	disastrous outbursts of dammed lake	–	Zabirov 1955
		1937	adv. of glacier		–	
		1951	advance of more than 1 km		–	
		1963	adv. of 1.7 km		complex ground-based observations, terrestrial stereo-photosurveys, aerophotograph. monitoring	Dolgoushin and Osipova 1982
		1973	adv. of 1.9 km			Osipova *et al.*, 1991
		1988–89	adv. of 1.1 km			IAHS(ICSI)/ UNEP/ UNESCO 1993*a*
	Didal	1897	ice avalanche	destruction	–	Uskov and Kvachev 1979
		1939	ice avalanche		–	
		1974	adv. of 1.3 km,		route and aerial inspections	
	Lenin	1945	adv. of glacier	–	–	Dolgoushin and Osipova 1982
		1969–70	adv. of 1.1 km		route inspections, analysis of aero and space images	
Hissaro-Alai	Abramov	1972–73	adv. of 0.6 km	–	complex ground-based observations (mass balance, repeated topo-surveys, measurments of ice movement, etc.)	Suslov *et al.*, 1980
Tien Shan	Bogatir	1985–90	adv. of 0.9 km	–	–	Kazanskiy and Fedulov 1990
	Shokalskiy	1939–40	inner surges of streams	possible surge of all glacier	periodical visitations, ice - movement measurements	Makarevich and Fedulov 1974
		1950				
Kamchatka	Bil'chenok	1959	adv. of glacier	–	–	Vinogradov *et al.*, 1982
		before 1982	inner surge, slow advance of glacier	–	repeated terrestrial stereo-photosurveys	

Tuyuksuyskiy Glacier. Disastrous mudflows and avalanches also occur in other regions of Middle Asia, in the Caucasus and in several Russian mountain ranges. State anti-mudflow and anti-avalanche institutions have been established in the Hydrometeorology Departments of Moscow University and universities of the Republics, as well as within the Academies of Sciences, to study the involved processes, observe potentially dangerous sites and organize anti-mudflow and anti-avalanche prevention measures. Outbursts of glacial lakes, advances of glaciers and ice avalanches had already been recognized and described in Tsarist Russia within the framework of geographical exploration. As a consequence, glaciers presenting a threat to the population and economy have been under observation for many years. The advances of Devdoraki Glacier in the Caucasus, for instance, which repeatedly obstructed the Voenno-Gruzinskaya Road in Georgia, have been observed since the middle of the 19th century (Kalesnik, 1937).

Surging glaciers have been the object of detailed monitoring since 1963, i.e. after the spectacular surge of Medvezhiy Glacier in the Pamirs. The Medvezhiy Glacier became a reference glacier in the study of peculiarities in dynamic regime, enabling the elaboration of procedures for surge forecasting. The published findings of this research (Dolgushin and Osipova, 1982) were used by many authors in comparable cases. In addition to the annual observations of Medvezhiy Glacier conducted from 1963 up to 1991, there were also studies made of some other surging glaciers in the Pamiro-Alai, Tien Shan, Caucasus and on Kamchatka (Table 11.11).

Recent technological advances have brought with them a switch to aerial and space-monitoring of surging glaciers (Desinov *et al.*, 1977; Osipova and Tsvetkov, 1991). As a collaborative effort between scientists from Moscow University and the Institute of Geography of the Russian Academy of Sciences, a method of aero-pseudoparallaxes was optimized to study the flow velocity of surging glaciers (Knuzhnikov *et al.*, in press). Thus, there is a wealth of experience and materials available on surging glaciers at different levels of research. Unfortunately, in 1991, these studies were suspended owing to the complex political situation, which placed some glaciers beyond the reach of scientific investigation. It is hoped, however, that decoding and stereophotogrammetric processing of large-scale space surveys on a scale of 1:50,000 currently undertaken in Russia will enable this important work to be continued.

The main glacier hazards in China are outburst floods from glacier lakes. One type of flooding results from the sudden drainage of lakes dammed by glacier tongues, another one from outbursts of moraine-dammed lakes. Most of the reported events from glacier-dammed lakes occurred in the Karakoram and Central Tien Shan mountains, while floods from moraine-dammed lakes appear to be more frequent in the Himalayas and the East Nianqintanggula mountains (Xu, 1987). Several surging glaciers have also been found in south-east Tibet. Outburst floods from glacier lakes can have disastrous effects on the local population and on construction for water resources control. In Nepal, 14 glacier floods have been reported since 1964 (Yamada, 1993). Comparable problems also exist in India and Pakistan (Tarar, 1985; Hewitt, 1982; 1985).

11.6 GAPS AND NEEDS

The primary problem with regular observations of glacier fluctuations in the CIS is the general reduction in funding in recent years. In the first place, this problem affected the Institutions of the Hydrometeorology Departments, which had submitted more than 75% of the data on annual changes in glacier termini to the WGMS and conducted annual mass-balance measurements on two glaciers (Abramov in the Hissaro-Alai and Golubin in the Tien Shan). After the collapse of the USSR, this problem also affected the scientific institutions (national universities and Academies of Sciences). Regular glacier observations could soon be ceasing in many places if nothing is done.

The only way to persuade the individual state administrations of the CIS to provide financial support for glacier monitoring is to design a modern programme of glacier monitoring with concrete and clear purposes for the observations, listing necessary characteristics and methods of measuring glaciers, and to make recommendations wih respect to the analysis of data, which are important not only to scientific but also to practical institutions. The outstanding leader in the field of dynamic glaciology and, in particular, of fluctuations of glaciers, P. A. Shumsky (1975), pointed out that the purpose of monitoring glaciers over more than 100 years had yet to be defined in a satisfactory way. The elaboration of a programme (or programmes) dealing with a combination of scientific and applied issues on the utilization of data from glacier monitoring is a task of highest priority.

With respect to glacier monitoring in China, overseas support (finance, instrumentation or technology) may be increasingly necessary in the future because long-term field measurements cannot be extended to include a greater number of glaciers. High-resolution satellite imagery is needed to monitor glaciers in remote mountainous regions. Studies of glaciers on the Tibetan Plateau are now open to international scientific research. Any kind of research on the 'third pole of the world' is welcome, especially if it includes cooperation with China.

Variations of the summer-accumulation type glaciers in the Himalayas are sensitive to summer air temperature, solar radiation and precipitation. Shrinkage of glaciers caused by global warming may lead to a decrease in water resources and to an increase in hazards from outbursts of glacier lakes.

Therefore, monitoring of climate, glaciers and glacier lakes has a high priority with a view to arriving at a rational use of water resources and the prevention of glacier hazards. Continuous international cooperation for such monitoring, related studies and training is necessary for the development of Himalayan countries.

11.7 SUGGESTED FUTURE DEVELOPMENT OF MONITORING ACTIVITY

A large range of data on glacier fluctuations, unique in duration, frequency of repetition and quality, has been accumulated in the states of the CIS. This information must be systematically checked and published. Only clearly-defined programmes will enable glacier monitoring to continue. Scientific substantiation of both observational systems and glacier characteristics must be reached, guided by concrete scientific and/or practical purposes. It is necessary to use and/or work out modern and efficient methods of observing glaciers, including well-founded criteria for selecting adequate observations in every country. In order to improve the quality of data on the fluctuations of glaciers, it is expedient to hold regular international workshops for field personal and scientists dealing with the interpretation of these data. It also appears to be necessary to raise the issue of purposeful financing of glacier monitoring at the level of interstate programmes and institutions, as is done in hydrology and meteorology. International support would be helpful to maintain simultaneous observations on selected glaciers in Asia.

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12 Local glaciers surrounding the continental ice sheets

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12.1 INTRODUCTION

The large continental ice sheets of Antarctica and Greenland are beyond the scope of conventional glacier monitoring because they involve completely different spatial and time scales. The methodology of observations differs accordingly. However, numerous rather independent local glaciers surround the continental ice sheets and form an important part of worldwide glacier occurrence and distribution. Long-term observation of these local glaciers surrounding the continental ice sheets is thus an essential contribution to monitoring the world's glaciers.

12.2 LOCAL GLACIERS IN GREENLAND: DEFINITION AND COVERAGE

Although the dominant glaciological feature of Greenland is its continental ice sheet (the Inland Ice), the island also hosts a large number of local glaciers. At places, there is a gradual transition between the ice sheet and local glaciers. The following is therefore an attempt to define the ice sheet proper and, hence, to estimate the extent of the residual ice cover – the local glaciers.

The term 'local glacier' is here applied to all types of ice cover across Greenland (i.e., to glaciers in the form of ice caps, valley glaciers, mountain glaciers etc.), with the exception of the continental ice sheet (the 'Inland Ice'). The local glaciers do not only occur on the coastal ice-free areas of Greenland but also on the nunataks of the Inland Ice.

The total area of local Greenland glaciers was first estimated by Holtzscherer and Bauer (1954) on the basis of the sheets of the World Aeronautical charts on a scale of 1:1,000,000, published by the USAF, Washington. New figures on the area of Greenland and its ice cover have now been published by KMS (Kort- og Matrikelstyrelsen, i.e., the National Survey and Cadastre, Copenhagen) on the basis of new maps by this institution and the DGGU (Geological Survey of Denmark and Greenland) related to a new 1:2,500,000 topographical map based on updating by satellite positioning and altimetry (Weng, 1995). The area figures by Holtzscherer and Bauer (1954) are compared with those of Weng (1995) in Table 12.1.

Table 12.1 Areas of ice sheets and local glaciers in Greenland

	Holtzscherer and Bauer (1954)	*Weng (1995)*
Ice cover – the Inland Ice	1,726,400 km²	1,707,038 km²
Ice cover – local glaciers	76,000 km²	48,599 km²
Ice-free land	386,000 km²	410,449 km²
Total	2,188,400 km²	2,166,086 km²

Figure 12.1 Inland Ice coverage of Greenland according to Holtzscherer and Bauer, 1954. 'Parties negligées' only refers to the largest local ice-covered areas measured by these authors.

Both figures are based on large-scale mapping, i.e., they essentially cover the larger ice-covered areas, whereas minor glaciers are omitted. Hence, the figure of about 50,000 km² is only a minimum for the area of local glaciers. This is about 5% of the total global coverage by glaciers and ice caps (e.g., Barry 1985, IAHS(ICSI)/UNEP/UNESCO, 1989) but nevertheless, in literature, is often labelled 'Greenland' and simply added to that of the ice sheet, if not entirely omitted.

The differences between the two published figures of local glaciers in Greenland must be ascribed to

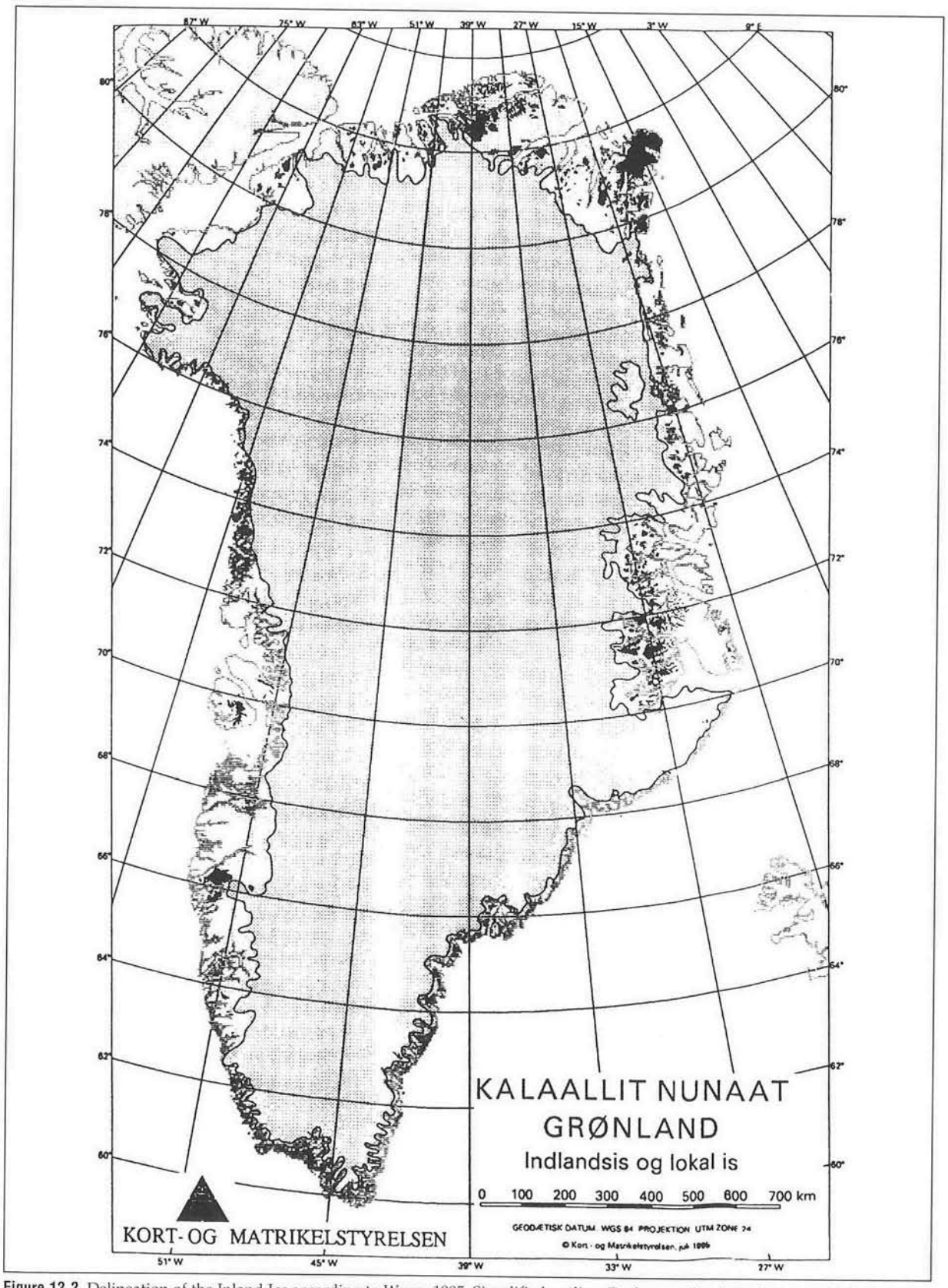

Figure 12.2 Delineation of the Inland Ice according to Weng, 1995. Simplified outline. Dark areas: local glaciers, central shaded area: the Inland Ice cover. Published 1995 with permission from Kort- og Matrikelstyrelsen, Danmark (A. 200/87).

differences in quality and scale of the cartographic material applied but also to the definition used for the ice sheet and, therefore, also that used for the remnant – the local glaciers. Outlines of the concepts of the Inland Ice are shown in Fig. 12.1 (Holtzscherer and Bauer, 1954) and Fig. 12.2 (Weng, 1995). Weng (personal communication, 1995) includes all glacierized areas connected to the main ice sheet as part of the Inland Ice, whereas Holtzscherer and Bauer seemingly omit part of this cover in northwest and east Greenland and refer to it as local glaciers ('surfaces englacées, à l'exception de l' Inlandsis').

With certain exceptions, the concepts given are conventional in the sense that the local glaciers are well-defined units on the coastland, usually separated from the main ice sheet by deglaciated land. The order of magnitude ranges from large ice caps, such as (Weidick *et al.*, 1992):

Flade Isblink (northeast Greenland), ca. 9,000 km^2; Julianehåb Ice Cap (south Greenland), ca. 6,500 km^2; Hans Tausen Ice Cap (north Greenland), ca. 3,200 km^2; Sukkertoppen Ice Cap (west Greenland), ca. 2,000 km^2, down to numerous glacierets smaller than 1 km^2.

Other area figures for major local ice caps (or groups of ice caps) are given by Cailleux and Lagarec (1977). However, a complete glacier inventory has only been attempted for south and west Greenland between 59°30′ and 71°00′ N (Weidick *et al.*, 1992). For this roughly 20% of the Greenland coastal area, the local glaciers cover 14,574 km^2 which could be taken as evidence for a total coverage of the Greenland coastland by some 50,000–80,000 km^2 if all local glaciers are included. An attempt at compiling all local glaciers with areas larger than 64 km^2 (126 units) also indicates a figure of ca. 70,000 km^2 based on the data of Weidick (1985) from an earlier 1:2,500,000 map. Although still a cautious minimum guess, a figure of ca. 70,000 km^2 for the conventional coastal local glaciers must at present be adopted for these locally glacierized areas.

12.3 DISTINCTION BETWEEN THE INLAND ICE AND LOCAL GLACIERS IN GREENLAND

Interest in local glaciers and their present response to climatic change is connected to the greenhouse effect, where thinning and recession of the global ice cover strongly relate to the minor local glacier areas, as witnessed by the generally strong recession of these glaciers during the 20th century (Meier, 1984; Haeberli *et al.*, 1989). The same trend can be observed at most local glaciers in Greenland, as exemplified for west Greenland by Weidick *et al.* (1992), partly for north Greenland by Davies and Krinsley (1962) and for east Greenland by Ahlmann (1948) and by later observations listed in Weidick (1995).

Response patterns of local glaciers deviate from those of the Inland Ice where there is current discussion on the total gain or loss (Zwally *et al.*, 1989; Douglas *et al.*, 1990) and where direct observations on the activity of the ice sheet margin show local behaviour of both advance and retreat (Weidick, 1991). Such complex fluctuation patterns seem to be connected with the superposition of short-term impacts from temperature variations on the ablation at the ice margin onto dynamic response of the ice margin to the long-term evolution of mass balance in the interior of the ice sheet (Huybrechts, 1994). These local deviations could imply that another useful definition for local glaciers probably is their sensitivity and frequency response to effects of climatic change. In general, a distinction between the Inland Ice sheet proper and the larger local ice caps relates to the size of the ice body, as stressed in the definition by Armstrong *et al.* (1973): not only the surface area but also the characteristic average thickness of kilometers for ice sheets differs from the typically encountered hectometers of glaciers and ice caps. Some marginal areas of the Inland Ice illuminated by low sun angles on Landsat images and documented by more recent detailed information on surface contours clearly distinguish themselves from the central ice sheet. Although these features are largely merged with the ice sheet proper along some distances of their margins, they must have their own life and response to climatic change and, therefore, exhibit the same fluctuation patterns as other glaciers elsewhere around the globe. A simplified map of the corresponding Greenland areas is shown in Fig. 12.3 (Weidick *et al.*, 1992).

12.4 CLASSIFICATION OF GREENLAND LOCAL GLACIERS

Besides the usual classification of glaciers (IAHS (ICSI)/UNEP/UNESCO, 1977), the discussion above points to a distinction of local glaciers in Greenland according to their connection with the ice sheet. Three groups appear in this context with gradual transitions:

1. The conventional local coastal glaciers, situated on the 'ice-free' coastland or on the islands, which (possibly with the exception of northeasternmost Greenland, cf. Davies and Krinsley (1962)) have essentially behaved in the same way as most other glaciers in the world during the 20th century, i.e., with marked recession. A few of these glaciers merge with the ice sheet for shorter distances (up to ca. 50 km) but, by their morphology, are discernible as units independent of the ice sheet but dependent by their subsurface (examples: Inglefield Ice Cap, Pituffik Ice Cap and Julianehåb Ice Cap, Fig. 12.3).
2. Local 'fringing' ice caps or ice fields (cf. IAHS(ICSI)/UNEP/UNESCO, 1977) which, by the location of ice divides and by the surrounding and subglacial topography, can be classified as local glacier units and which, in a dynamic sense, must also act as such. Their mergence to the ice sheet, however, is so extensive that inland parts of these units largely contribute to the nearby ice sheet outlets, whereas the coastal parts must more or less be regarded as local glaciers (examples in Fig. 12.3: Blosseville and Ammassalik). In conventional area determinations, this transitional type is included in the determination of the Inland Ice area and, so far, little is known of their geometric changes over time. Surging behaviour is often exhibited for the Blosseville Ice Cap out-

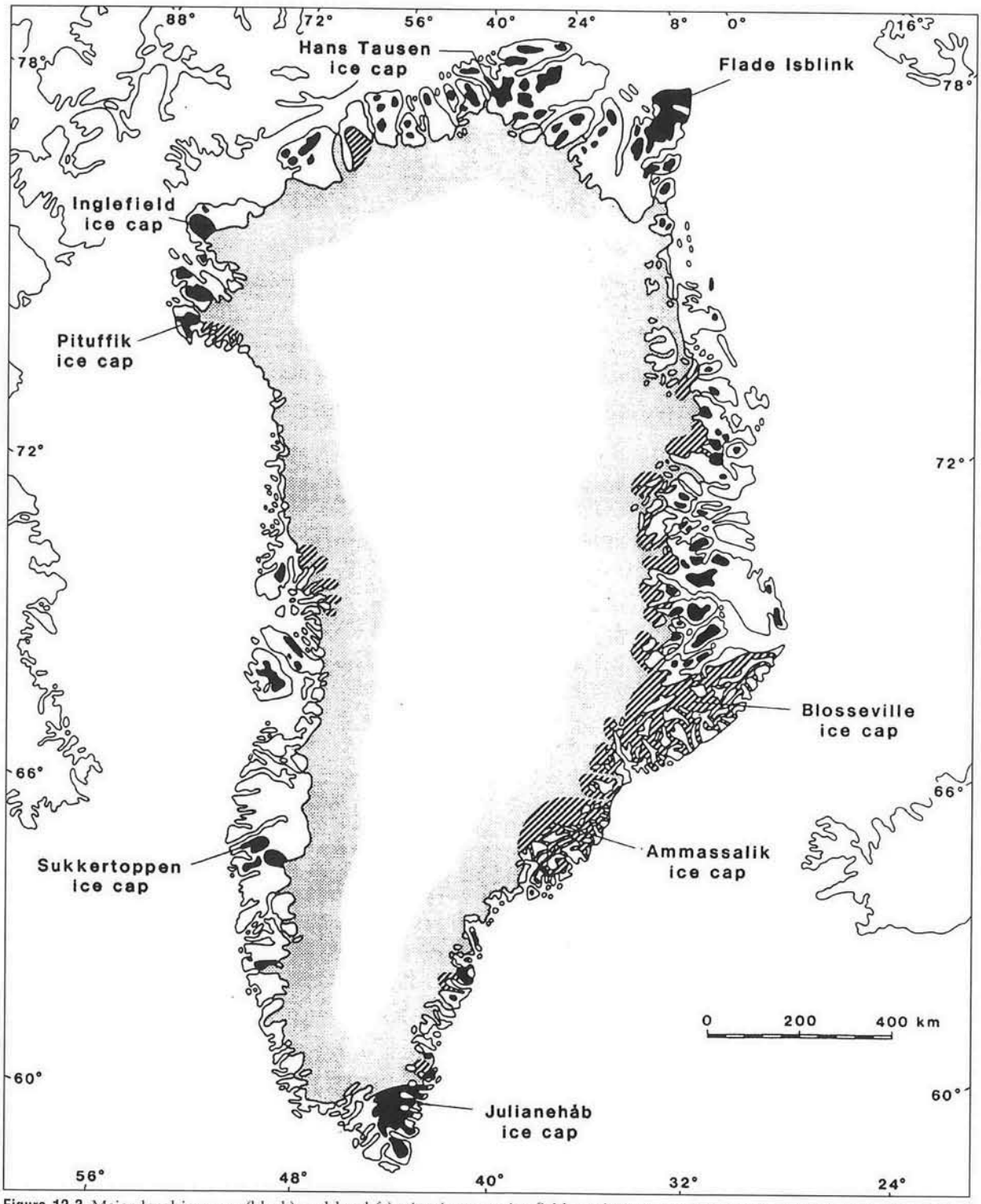

Figure 12.3 Major local ice caps (black) and local fringing ice caps, ice fields and ice domes (oblique hatching). Inland Ice proper shaded grey. Simplified outline from Weidick *et al.*, 1992.

lets (Weidick, 1988). Area determinations of the individual basins draining to the coast have here been determined on the basis of Landsat data (Dwyer, 1993).

3. Ice domes and other local glacier units within the ice-marginal zone or on the nunataks of the ice sheet were formerly considered a third group of local glaciers but, to a restricted degree, are independent ablation areas and are – as compared to the areas of the local glaciers of groups 1 and 2 – of negligible importance.

At present, only the coastal local glaciers described under 1 are considered as 'true' local glaciers. The question of definition and investigation of groups 2 and 3 is only mentioned here in order to draw attention to the anomalies.

Table 12.2 Locations of mass-balance series covering more than a budget year

Glacier	*Period*	*Reference*	*Lat.*
Narssaq Bræ	1970–71, 1980–83	Clement, 1984	61°00′
Valhaltindegletscher	1978–83	Clement, 1984	61°26′
Gletscher 33	1982–89	Braithwaite, 1989	64°02′
Qapiarfiup sermia	1980–89	Olesen, 1986	65°34′
Amitsulooq Ice Cap	1981–90	Olesen, 1986	66°08′

Source: Weidick, 1995

12.5 MEASUREMENT OF MASS BALANCE AND GLACIAL CHANGE IN GREENLAND

Series of mass-balance measurements covering several years are only available on a very restricted number of glaciers, essentially concentrated in south and southwest Greenland. The locations are given in Table 12.2.

Supplementary mass-balance measurements have been made in northwest Greenland (Redrock: Goldthwait, 1971), east Greenland (Frøya Bræ: Ahlmann, 1948) and southeast Greenland (Hasholt, 1986). In north Greenland (Hans Tausen Ice Cap), mass-balance investigations were initiated in 1993 by the Geological Survey of Greenland (GGU)[1]. Location of this ice cap is shown in Fig. 12.3. Thus, mass-balance measurements are still sparse, as are measurements of glacier-length changes. Nowhere in Greenland are the measurements up to European standards for annual registration but there is now historical documentation on the changes in several local glaciers from all parts of Greenland, often going back 200 years in west and south Greenland and less than 100 years in north Greenland. The glaciers nearly all show a general trend towards recession throughout the 20th century. Dominant expansion has been reported from the Red Rock Ice Cap between Inglefield and Pituffik Ice Caps (Fig. 12.3: Goldthwait, 1971) and repeated photogrammetry on Mitdluarkat Glacier near Ammassalik on the southeast coast has shown an increase in glacier volume over recent decades (Hasholt, 1986). A slow-down in recession and a reversal of frontal retreat trends to the point of discernible advance has also been noticed for other glaciers on the west coast (Weidick *et al.*, 1992). Gordon (1981) concluded, on the basis of investigations on mountain and valley glaciers in the Sukkertoppen area, that changes in the 19th century were in harmony with temperature changes but showed a lag of 9 to 30 years in response to precipitation change, with the greatest lag being observed for the largest glaciers. In south Greenland, most outlets from the Julianehåb Ice Cap retreated considerably during the 1980s (Weidick, 1994).

The two local ice caps of Reenland (71°10′ N; 26°43′ W) in east Greenland and Hans Tausen (82° 30′ N; 38° 20′ W) in northeast Greenland are both penetrated by deep ice core drillings. The ice core climatic record of Reenland Ice Cap extended to Eemian interglacial (Johnson *et al.*, 1992) and that of Hans Tausen Ice Cap only to the Holocene climatic optimum, implying that this ice cap reformed at the onset of the Late Holocene climatic cooling (Reeh, 1995).

12.6 FUTURE WORK IN GREENLAND

The evidence sketched above points to the local glaciers of Greenland constituting an important part of the total global coverage by glaciers and ice caps. They should therefore not be forgotten in spite of the fact that a concise total value and listing of glacier units in accordance with WGMS standards has not been made so far. The estimate from the west Greenland inventory indicates that such an undertaking would presumably concern 20,000 to 30,000 glacier units.

The presently predominating preference for short-term scientific projects leaves scarce resources for systematic glacier inventory work, which otherwise might be considered a potential national contribution to international scientific collaboration. Whereas detailed information on glaciers was useful for hydropower-planning in west Greenland, this impetus for systematic work is absent in other regions of Greenland. Besides renewed recommendations for such work by international organizations concerned with the development of glacier monitoring, efforts should possibly concentrate on upscaling the work from the present 1:2,500,000 scale maps to the current total coverage of Greenland on a scale of 1:500,000. Uniform coverage of Greenland's coastland by aerial photographs on a scale of 1:150,000 (west Greenland in 1985, north Greenland in 1978 and east Greenland in 1981 and 1987) at KMS makes this scale feasible and attractive. This is especially true with respect to a potential inventory of local coastal glaciers classed in group 1, where the arguments for supporting glacier inventory work would be the same as for glacier inventory work elsewhere on the globe.

For ice cover labelled under points 2 and 3 (that connected to the ice sheet), this work so far offers only a tentative classification as local glaciers but, at

1 The Greenland Geological Survey (GGU) and Denmark Geological Survey (DGU) have since merged to form the Geological Survey of Denmark and Greenland (DGGU).

present, without much substantiation for their importance with respect to trends in sea level. This work, as well as the continued monitoring of Greenland glaciers by the use of remote-sensing techniques, as outlined by Williams and Hall (present report), will certainly be preferable in Greenland where its application has been well demonstrated by Dwyer (1993).

12.7 LOCAL GLACIERS IN THE ANTARCTIC

Many of the ideas developed in studying local glaciers in Greenland are applicable to the Antarctic, although research in this field is not as well developed as in the north and estimates of the total area of local glaciers must be very tentative. Coastal local glaciers are most obvious on the Antarctic Peninsula, which is a mountain range stretching over 1,500 km nothwards from Ellsworth Land (latitude 75°S), with peaks rising to over 2,500 m a.s.l. The Antarctic Peninsula Plateau is covered by a long and relatively narrow ice sheet. The ice sheet acts as a source for valley glaciers, which cut through the coastal mountains and terminate in ice cliffs at sea level. In some cases, mostly in the north, the glaciers end on land but generally the equilibrium line is below sea level in this region. Some 2% of the total area of Antarctic ice could be classified as essentially coastal local glaciers with rapid dynamic responses to the order of tens of years.

There is a continuous transition to the second class of local glacier defined above, that is to say, a discrete dynamic unit attached to a larger ice sheet. In the southern part of the Antarctic Peninsula, for example, the Plateau Ice Sheet broadens out and this large reservoir of ice ensures that the reponse of the valley glaciers it feeds is on a longer time scale, to the order of 100 years. The extreme case is demonstrated by the great ice streams, which flow from the central Antarctic Plateau down to the ice shelves girdling the continent. These include, for example, Ice Streams A, B and C flowing into the Ross Ice Shelf and the Evans, Carlson and Rutford Ice Streams flowing into the Ronne Ice Shelf. These are clearly defined dynamic units with rapid flow in one direction but are so intimately linked to the ice sheet that their reponse could not be regarded as local.

Isolated domes of ice could be regarded as a third type of local glacier, for example the small ice rises on the Larsen Ice Shelf or even Berkner Island on the Filchner-Ronne Ice Shelf. Again, there is a continuous transition from completely isolated domes to those like Law Dome which, although there is a local radial pattern of flow, will still have a response controlled by the main continental ice sheet.

Choosing a rather arbitrary definition of a local glacier as a body of ice having a typical response time of up to 100 years, Weidick and Morris would judge that local glaciers do not form more than about 5% of the area of Antarctic ice. Using the figure of 13,918,000 km^2 quoted by Drewry *et al.* (1982) for the total conterminus area of Antarctica (which includes 2.4% ice free terrain), they estimate the area of local glaciers to be no more than about 680,000 km^2.

It seems clear that the area of local glaciers in Antarctica is several times greater than the area they cover in Greenland but this, of course, does not imply that their contribution to sea-level rise over the last century has also been several times larger. What is significant from that point of view is the area of ice experiencing ablation. Drewry and Morris (1992) estimated that an area of some 20,000 km^2 of ice on the Antarctic Peninsula has a mean annual surface temperature warmer than −11°C and will experience some melting in summer months. It is the local glaciers in this region (and the very few coastal areas around east Antarctica which experience summer melting) which need to be considered separately from the main mass of the ice sheet. Zwally and Fiegles (1994) have derived the extent and duration of surface melting on the Antarctic ice shelves and margins of the ice sheet from satellite passive-microwave data for 1978–87. They find that the daily average area with surface melting for December and January ranges from 2% to 4% of the total continental area. However, much of this area is made up of ice shelves or ice sheet marginal areas which cannot be classified as local glaciers. Weidick and Morris estimate that the area of ablating local glaciers is of the same order of magnitude in Greenland and Antarctica.

12.8 MEASUREMENTS OF THE MASS BALANCE OF LOCAL GLACIERS IN THE ANTARCTIC

Measurements on local glaciers in the Antarctic are few and far between and there has been no long-term mass-balance monitoring comparable to that undertaken in less remote regions. Estimates of mass balance are based either on measurement of changes in glacier geometry or on calculations using an energy balance model. In the Antarctic Peninsula region, it is possible to begin to get an overview of the regional response to the recent climatic warming because measurements have been made over a wide range of latitudes, from 62–71°S. However, it should be noted that there are major problems in estimating calving when glaciers do not terminate on land. In this case, it is more useful to consider the surface mass balance rather than to attempt to calculate the glacier mass balance.

12.8.1 South Shetland Islands

This archipelago north of the Antarctic Peninsula has been the site of much of the mass balance work undertaken so far. There are several Antarctic bases collecting meteorological data on King George

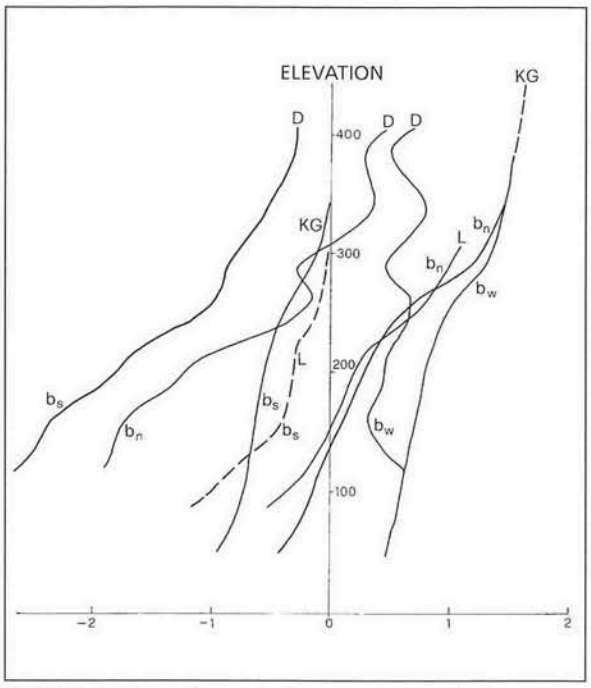

Figure 12.4 'Normal' mass-balance curves for the three locations: 'G1' on Deception Island (D), Rotch Ice Dome on Livingston Island (L) and King George Island (KG). b_s is summer balance and b_w is winter balance (after Orheim and Govorukha, 1982).

Island, the largest central island. Early papers include a study (Noble, 1965) of Flagstaff Glacier (0.094 km²) on King George Island during the International Geophysical Year (IGY, 1957/58). The specific mass balance of the glacier was calculated as –0.53 m water for the IGY i.e., it was undernourished. Orheim and Govorukha (1982) summarized the results of surface mass balance studies on Deception Island (D) from 1968/69 to 1973/74, on King George Island (KG) in 1969/70 and 1970/71 and from Livingstone Island (L) from 1971/72 to 1973/74. The measurements were made along stake lines running from the ice edge (on land in all cases) to the watershed. Fig. 12.4 shows the average variation of summer (b_s) and winter (b_w) surface mass balance for the three locations. Further work was undertaken on the Ecology Glacier (King George Island) in the 1990/91 season (Bintanja, 1995). The surface mass and energy balance were measured for one month at a site 100 m a.s.l, during which time there was 0.735 m water ablation, which is close to the summer ablation of 0.75 m water at this altitude on KG shown in Fig. 12.4. Both sets of measurements were made during periods of a warmer than average mean annual temperature in the region. Knap *et al.* (in press) used an energy balance model driven by climatological data from nearby stations to predict annual surface mass balance curves for the KG Ice Cap and demonstrate their sensitivity to changes in air temperature and precipitation. In sum, the various studies demonstrate that South Shetland ice masses experience considerable ablation in summer and will lose mass as air temperatures increase.

12.8.2 Argentine Islands (65°S, 64°W)

Two small ice caps, Galindez (0.151 km²) and Skua (0.209 km²) lie close to the UK station Faraday (where meteorological measurements have been taken since 1944). Measurements in 1963, 1965 and 1966 (Sadler, 1968) suggested an accumulation rate of 0.58 m water/a and ablation of 0.39 m water/a, including calving of 0.09 m water/a on Galindez Ice Cap. The ice cap mass balance is therefore positive (0.19 m water/a). The surface mass balance is 0.28 m water/a. Later measurements by Shanklin (pers. comm.) show a significant reduction in height between profile surveys in 1987 and 1995. This amounts to around 3.4 m snow at the top of the Galindez Ice Cap. Before 1987, there was only a small variation in the height of the ice cap, which was between 55 and 56 m high. The mean annual air temperature at Faraday has increased from around –5.8°C in the late 1940's to –3.0°C in the mid–1990s, although warmer and cooler periods have occurred during this time (Fig. 12.5). The increased ablation in the last decade can be linked to warmer air temperatures.

12.8.3 East Antarctic Peninsula ice rises

Measurements on three small ice rises on the Larsen Ice Shelf (Martin and Sanderson, 1980) were reassessed by Doake (1982). He calculated a rate of surface lowering of 0.42 m ice/a for Butler Island (72° 13′ S, 60° 50′ W), 0.59 m ice/a for Dolleman Island (70° 35′ S, 60° 50′ W) and 0.52 m ice/a for Gipps Ice Rise (68° 45′ S, 60° 05′ W), indicating that all three ice rises were undernourished by the measured accumulation rates of 0.23 m, 0.41 m and 0.38 m ice/a respectively. Doake postulates a period of higher accumulation rates some hundreds or thousands of years ago to which the ice rises are still responding. Ablation by surface melting need not be considered as the mean annual temperature in the region is ca. –17°C.

12.8.4 Spartan Glacier (71°03′ S, 68°20′ W)

During the International Hydrological Decade (IHD, 1965–74), which was a warmer than normal period,

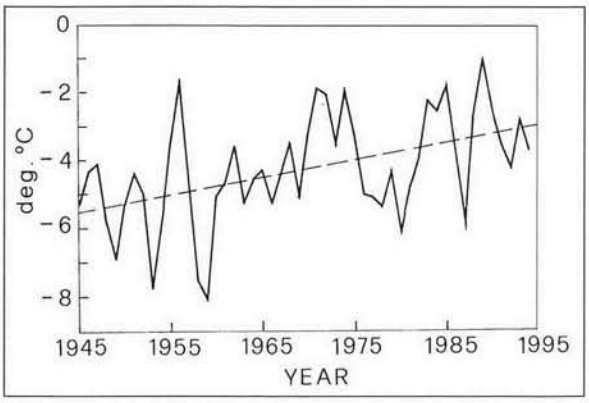

Figure 12.5 Annual mean air temperature at Faraday.

measurements were taken on this small (6.3 km²) corrie glacier on Alexander Island, which lies just west of the Antarctic Peninsula. Mass balance was measured from 1969 to 1974, surface movement from 1969 to 1973, energy balance from 1972 to 1974 and ice depth was surveyed in 1972 (Jamieson and Wager, 1983; Wager and Jamieson, 1983). A survey of the glacier volume showed a decrease of 0.2% per year. The mass budget is shown in Table 12.3.

Table 12.3 Mass budget of Spartan Glacier

Year	*m water/ a*
1970	−0.084
1971	−0.123
1972	−0.186
1973	−0.030

12.8.5 Moraine Corrie (71°20′ S; 68°16′ W)

This small glacier on Alexander Island lies close to the UK summer station of Fossil Bluff, where meteorological measurements have been taken intermittently since 1961. The elevation of the glacier surface has been measured along a roughly longitudinal transect in 1972, 1986 and 1993 (Morris and Mulvaney, in press). Between 1986 and 1993, there was significant melting of 3.64 m water. The corrie is very sheltered and a considerable amount of black moraine lies on the surface of the glacier. It seems likely that the increased melt rate is related to a marked decrease in albedo produced by the loss of snow cover over the summer months. The ablation sensitivity over the period 1972 to 1993 is 1.0 m water/a K, much greater than that measured nearby on the ice sheet proper (0.17 m water/a K) where, although the mean annual temperature is the same, the surface is white, unsheltered from the wind and there is no backscattered radiation from rock walls.

12.9 FUTURE WORK IN ANTARCTICA

Results from the Antarctic Peninsula region, sparse though they are, demonstrate a clear response to the warming of some 2°C that has occurred over the past 40 years. Increased summer melting in areas with a mean annual temperature greater than about −11°C has contributed to sea-level rise (Drewry and Morris, 1992), although by a small amount compared to the fall in sea level from increased accumulation over the continent as a whole. As in Greenland, further work is needed to establish mass-balance curves for the region but the primary need is for mapping on a sufficiently detailed scale to resolve the form of local glaciers and enable more precise estimates of their area to be made.

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13 Monitoring ice sheets, ice caps and large glaciers

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13.1 INTRODUCTION

Most of the ice on Earth (more than 99 per cent) is to be found in the two large ice sheets in Antarctica and Greenland. Much of the remaining ice occurs in individual glaciers larger than 100 km^2 in area. Glaciers smaller than this, however, are more numerous, generally better known, simpler in shape, often more accessible and more amenable to simple hydrometeorological and dynamic modelling. Thus, much of what we know about glacier/climate interrelations stems from observations on small glaciers. However, the larger glaciers, including ice caps, ice fields and ice sheets, cannot be ignored, especially as global warming causes the removal or reduction of much of the small glacier ice mass. Areal measurement of surface mass balance and/or energy balance for these glaciers is difficult owing to spatial variability and logistics costs; the use of satellite imagery or altimetry to detect volume change is promising but still uncertain. New geophysical techniques may make it possible to estimate, over large areas, the net changes in ice mass over periods of 1 to 100 years. These include 1) repeated laser altimeter soundings from aircraft equipped with the dynamic Global Positioning System (GPS) or from Earth-orbiting satellites 2) changes over time in gravity measured from bedrock stations and 3) changes over time in the positions of bedrock stations. Perhaps oceanographic measurements of freshwater outflow and geodetic measurements of changes in Earth rotation will help to confirm or constrain these estimates. Monitoring of large glaciers is needed but cannot be done in accordance with the conventions currently applied to small glacier studies.

13.2 INVENTORIES OF THE ICE SHEETS

Recent publications (Swithinbank, 1988; Weidick, 1995) have used satellite imagery to update earlier atlases (Tolstikov, 1966; Drewry, 1983) of the Antarctic and Greenland Ice Sheets respectively. As a result, the broad features of these ice sheets are well known. The surface configuration, however, is not yet mapped in sufficient detail for some kinds of analyses, such as defining drainage-basin boundaries, and many areas still lack information on the bed configuration. Radar (and, in the future, laser) altimetry from satellite platforms is producing much-needed detail on surface topography, but unfortunately, northern Greenland and much of Antarctica are not yet covered by altimeters in polar-orbiting satellites.

Another problem concerns changes in the margins of these ice masses. Most of the ablation in Antarctica and more than half of that in Greenland is

by iceberg calving. Calving can be very episodic, so rapid changes in the margins do occur. Recent breakups of the Larsen and Wordie Ice Shelves and the release of enormous icebergs (several exceeding 10,000 km^2 in area) from the Ross and Filchner Ice Shelves in Antarctica have focused public attention and concern about a possible 'collapse' of the West Antarctic Ice Sheet. Also, observable changes in the ice streams feeding these floating ice shelves are occurring, lending further credence to the notion that these huge ice sheets do change on relatively short time scales and must be continually monitored in order to be fully understood.

Detailed inventories of Greenland and Antarctica cannot be easily incorporated into the formats used by the WGMS for small-glacier inventories primarily because of the difficulty in defining individual 'glaciers', which can only be separated by obscure and changeable drainage divides, and because of their huge size. Hence the practical difficulties in measuring them.

13.3 ICE SHEET DYNAMICS

Ice flow observations are needed in order to model the response of glaciers to climatic change and for other purposes. Fortunately, new technology now allows surface velocity measurements to be made on the huge ice sheets where this was previously very difficult. Global Positioning System (GPS) receivers are now readily available, relatively inexpensive and sufficiently precise to obtain velocity data with reasonable re-survey intervals (weeks, months, up to a year or so). Repeated satellite images (Lucchitta and Ferguson, 1986; Bindschadler and Scambos, 1991) can be used to measure surface velocity and strain-rate fields for active, crevassed ice streams. It may also be possible to apply satellite synthetic-aperture radar (SAR) for this purpose (Fahnestock *et al.*, 1993); although data reduction is problematical, this is an all-weather system which will be important in some areas.

Satellite radar interferometry is an exciting new possibility for obtaining surface velocity fields with spatial detail never before possible (Goldstein *et al.*, 1993). If this method can be combined with the high resolution and all-weather capability of satellite synthetic-aperture radar (SAR), the result will be a powerful tool for measuring large glacier dynamics.

13.4 ICE SHEET MASS BALANCES

No study of the relation of glaciers to climate can be considered complete without due attention to the two huge ice sheets dominating the Earth's land ice system. Only the ice sheets significantly affect global climate, as well as being affected by the climate. Thus, determining the mass balance of the Greenland and Antarctic Ice Sheets and its sensitivity to climate is an important scientific task.

Current changes in the ice sheets can be measured using surface mass-balance observations or by geodetic (volume-change) methods. However, the ice sheets respond to processes on all time scales ranging from the last glacial-interglacial transition to climatic fluctuations of the last decade or year. As most observations extend over a few decades at most, how can one decide whether a current change is in response to the current climate or to a slow adjustment to changes that happened a long time ago? This question requires numerical modelling (e.g., Huybrechts, 1990; 1994). Data from automatic weather stations are needed to calibrate satellite data and to provide input data for mass-balance modelling. Currently, most automatic weather stations on ice sheets are in the accumulation zone. It is, therefore, important to extend the use of automatic weather stations to the ablation zone of Greenland.

The size of the ice sheets is the major obstacle in mass-balance studies. In principle, the net mass balance can be determined by summing balance observations on the surface (with due attention to infiltration/refreezing) and subtracting the loss of ice by calving and sub-ice shelf melting. However, this procedure leads to large errors because of incomplete coverage. When the interest is in sea-level changes, the mass flux across grounding lines is more appropriate than the iceberg-calving discharge.

Geodetic methods can also be used to determine mass balance changes by measuring the rate of change of thickness over an ice sheet. This can be done using ground-based traverses, including precise levelling, or through the use of altimetry from satellite platforms. The first method is labour intensive and thus limited in scope; the second method has produced results which are still somewhat controversial and is limited to the current coverage of satellite orbits (not more than 82° north or south latitude).

The current net balances for the whole Antarctic or Greenland Ice Sheets are still somewhat in doubt. Even the sign of the net balance is uncertain, as the net value is thought to be close to zero yet an imbalance of ± 25 per cent cannot be detected with existing data (Bentley and Giovinetto, 1991; Jacobs *et al.*, 1992). On the other hand, mass-balance histories of specific points on these ice sheets can be determined by ice coring. These techniques provide seasonal or annual resolution in some areas of Greenland as far back as 40,000 years or more and with coarser resolution in Antarctica back at least 200,000 years. These chronologies are very important as they can be used to place the current observations in a longer-term context. In addition, they show abrupt climatic events unlike anything seen in historic times, which suggest that the mechanism for change in the Earth's climate system is not yet fully understood.

Field (surface-based) observations of mass balances of the big ice sheets need to be continued, extended and reported in an efficient data-exchange system. Attention also needs to be paid to detecting

changes in ice dynamics, including the flow of ice streams, the movement of grounding lines and the rates of iceberg calving. Additional observations of water-vapour flux divergence would be useful to better define the processes leading to changes in snow accumulation. New high-resolution ice cores from Greenland and Antarctica will be needed to better characterize changes in precipitation associated with changes in air temperature. Dynamic interactions between ice streams, ice shelves and iceberg discharge should be observed and reported. Much of this work will require observations from orbiting satellites; radar/laser altimeters will be critical for synoptic monitoring, but synthetic-aperture radar and high resolution images will also be important. Until these instruments are installed in polar-orbiting satellites, their potential to solve the difficult problem of monitoring large ice sheets will not be realized.

13.5 INVENTORIES OF LARGE GLACIERS EXCLUSIVE OF THE ICE SHEETS

Large glaciers are clearly visible on high-resolution and even medium-resolution satellite imagery such as SPOT, Landsat-MS and TM, and NOAA-VHRR (e.g., Dowdeswell and McIntyre, 1987; Krimmel and Meier, 1983; Williams and Hall this volume). This means that their areal extent can be clearly delineated and measured. Measuring surface topography for some very large ice caps may be possible using satellite radar altimetry, or perhaps through some of the new and developing techniques using satellite stereoscopic image analyses. One important difficulty in measuring large glaciers using remote sensing from satellites is that some of them exist at very high latitudes, beyond the latitude of most current satellite orbits.

Ice thickness measurement for these large glaciers generally requires airborne radio-echo sounding, which has limitations when liquid water exists in the ice. On some relatively uncrevassed ice caps and glaciers, surface radio-echo sounding traverses using over-snow vehicles are possible but on large, very active glaciers, such ground traverses are usually impractical.

Large glaciers also create a data-management problem for a global inventory. Many have multiple termini, such as numerous outlet glaciers from an ice cap (e.g., Axel Heiberg Island, Canada) or the several termini of inter-mountain ice fields (e.g., Juneau Ice field, Alaska). This creates difficulties in categorizing glaciers by drainage basin because there is no precise method for delineating drainage basin boundaries and several major river systems may be involved. A similar difficulty ensues because these glaciers may cross political boundaries (e.g., Patagonian Ice Fields in Argentina/Chile) or span mountain ranges (e.g. Bering Glacier in Chugach and Saint Elias Mountains, U. S./Canada).

13.6 DYNAMICS OF LARGE GLACIERS

Repeated satellite (Lucchitta and Ferguson, 1986; Bindschadler and Scambos, 1991) or aircraft (e.g., Meier *et al.*, 1985) images can be used to measure surface velocity and strain-rate fields for active, crevassed glaciers. It may also be possible to apply satellite SAR for this purpose (Fahnestock *et al.*, 1993). Satellite radar interferometry (Goldstein *et al.*, 1993) is a possibility for obtaining surface velocity fields for large valley glaciers in the future.

13.7 LARGE GLACIER MASS BALANCE

Conventional techniques for measuring surface mass balance (IAHS/UNESCO, 1970; 1973) were developed for use on small glaciers; these are frequently impractical in terms of cost and manpower for use on large glaciers and ice caps. This is especially true for large inter-mountain glacier systems in maritime climatic environments, such as in the Ice Field Ranges or the Patagonian Ice Fields, where the combination of a great distance from a logistics base, pervasive crevassing, heavy accumulation and poor weather combine to make surface observations extremely difficult, dangerous and expensive. Compounding the problem is the fact that a single glacier may span climatic zones from maritime to continental and snow facies from temperate/ablation to the percolation or dry-snow facies. It may not be possible in the near future to produce area-integrated net balances for large glaciers such as these and certainly not of a quality similar to that routinely obtained from many small glaciers. This brings up the question, currently unanswered, of whether conventional mass-balance measurements on a nearby small glacier can be used to infer regional mass balance in areas where most of the ice is in a few very much larger masses. Can a small glacier in one such area be considered truly 'representative' of the region?

Index techniques using a limited number of stakes and pits (Koerner, 1986) can be used for determining changes from year to year, assuming that annual access is possible using aircraft or over-snow vehicles. The position of the equilibrium line may be determined on an annual basis using aircraft photography or satellite imagery for large temperate glaciers where superimposed ice is not a major component of the mass balance. Multispectral satellite imagery may allow the delineation of superimposed ice and other snow facies (Williams *et al.*, 1991) and it is possible that snow accumulation rates and patterns may be obtained from analysis of passive microwave imagery (Zwally, 1977). These techniques will be useful in defining mass-balance trends and their relation to climatic fluctuation. Yet, how are these data to be reported in a global mass-balance data exchange scheme?

Iceberg calving is another component of the mass-balance equation which can be important in many

large glaciers in the Arctic, the Antarctic and in several temperate regions, such as Alaska and Patagonia. Iceberg discharge speeds can be determined readily from the difference between ice velocity and terminus advance/retreat, both obtainable from repeated satellite or aircraft imagery. Iceberg discharge, however, requires knowledge of ice thickness, which is more difficult to obtain. As iceberg calving and ice-flow dynamics are coupled through various feedback mechanisms (Meier, 1994), it is important to combine calving and flow studies over common time periods.

13.8 LARGE GLACIER VOLUME CHANGES

For large glaciers, it may be more feasible to measure changes in volume than mass balance. This has be done for moderately large glacier systems in Central Asia (e.g., Akshiyak Range, Fedchenko Glacier) using aerial photography and this work needs to be continued, where possible. For large, relatively flat surfaces, such as some ice caps, repeated satellite altimetry (e.g., Zwally *et al.*, 1989; Herzfeld *et al.*, 1994) may be a useful new tool. Another promising new technique involves using repeated laser altimetry from an aircraft positioned with a dynamic GPS system (Echelmeyer *et al.*, submitted, Thomas *et al.*, 1995). This later technique promises to be especially useful for measuring glaciers, large and small, in mountainous, rough terrain. In the future, it may be possible to use satellite laser altimetry to determine thickness changes on ice caps.

Solid-earth geophysics offers additional ways of monitoring change in large ice masses. Elastic deformation (horizontal and vertical) of the Earth can be measured with centimeter accuracy using modern GPS techniques. This may be sufficient to detect changes in nearby large glaciers. This method has been used successfully to measure mass changes due to the surge of Bering Glacier (Sauber *et al.*, 1995) and has been proposed to measure thinning of the West Antarctic Ice Sheet (James and Ivins, 1995). These methods, although still under development, have the great advantage of responding to an integrated change over a large area and avoid the need to place stations on the moving ice. By combining these deformation surveys with precise gravity measurements, it may be possible to add valuable information.

Other geophysical techniques may help solve the problem of detecting or measuring change in the large glaciers of the world. Oceanographic surveys can measure the meltwater and iceberg discharge from areas such as the glacierized mountains around the Gulf of Alaska (Royer, 1982); when combined with weather or other data, this may yield new information on change integrated over very large areas. Another source of large-area information stems from modern studies of the rotation of the Earth. Both polar motion excitation and length-of-day are sensitive to changes in mass, such as that produced by regional glacier thinning (Trupin, 1993; Wahr *et al.*, 1995).

13.9 LARGE GLACIERS AND THE WORLD GLACIER MONITORING SERVICE

Most of the data currently reported by the WGMS pertains to glaciers less than 100 km^2 in area, yet the larger ice masses have appreciable effects on the Earth's system. Will data from these larger ice masses be reported and exchanged routinely in the future? This is problematical, for three reasons:

1. Although some of the possible observational techniques can be considered routine, such as mapping glacier margins from satellite images or measuring ice flow by GPS receivers on the surface, most of the methods listed above are still experimental. A considerable research effort will be needed if they are to reach their full potential and be incorporated in any 'standardized' data collection plan. This also means that it will take time to gain confidence in the method and to be able to assign realistic error limits.
2. Working on large, less-accessible glaciers will always be more expensive than working on the small glaciers that are the heart of the current glacier monitoring programme. The resource base for basic science is not growing, so funding is becoming increasingly competitive; priorities will have to be shifted to redress the small glacier/large glacier research imbalance of the present. There may be strong objections to any attempt to redress this imbalance for obvious, and sometimes parochial, reasons.
3. Monitoring of a global system is only useful if an efficient data exchange programme, such as WGMS, is in place. Nevertheless, monitoring of large glaciers will of necessity involve research projects and unconventional data collection programmes. The results of these investigations will be very important but difficult to incorporate in data exchange procedures and protocols. This will require a high level of innovation.

Large glaciers are an integral part of the Earth's interactive ice-ocean-atmosphere-land system. They occur in areas which may be able to provide valuable insights into the cause of changes in the Earth's climate system. They are not likely to disappear in the 21st century. Perhaps it is time to devote more effort and resources to measuring these components of the Earth's system.

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Conclusions and recommendations

Systematic and internationally coordinated monitoring during the past hundred years clearly reveals that the observed glacier shrinkage at decadal to secular time scales

- is a worldwide phenomenon;
- contributes to the observed rise in eustatic sea level;
- influences the water cycle and natural hazards in cold mountain areas;
- reflects an additional energy flux roughly comparable to the estimated radiative forcing, and, hence,
- constitutes a key indicator of ongoing climatic change.

Moreover, glacier mass-balance measurements provided the basis for the present understanding of the upper boundary condition of small and large ice masses. The mass balance represents a principal forcing function for glacier, and especially ice sheet, models. It is obvious that the systematic measurement and combined analysis of

- glacier mass balance as a direct climatic change signal;
- glacier length change, which may be a delayed but enhanced and perhaps a more easily determined climatic signal; and
- glacier spatial distribution patterns (glacier inventories) for assessing regional effects

should continue into the future. The value of the easily measured cumulative glacier length variation as a true and complete reflection of glacier mass change should not be underestimated, even though the glacier dynamics connecting mass and length change are still not fully understood. Experiments with glacier flow models, thus, need further refinement and testing against long-term observations. Advanced technologies have been found to record glacier length changes either directly or by remote-sensing methods. However, traditional methods, including repeated mapping, may, in some cases, continue to be useful and cost effective.

Complete data sets, and long-term records of glacier variations in space and time especially, exist in Scandinavia, in the European Alps and in a some other regions of the world, such as the former Soviet Union or the American Cordillera. The main problems, therefore, relate to

- continuing long time series;
- reaching global coverage; and
- monitoring large glaciers, especially with respect to sea-level change.

The continuation of long-time series in the coming years when global circulation models predict a warmer climate will be a special challenge. Reactions of glaciers existing in different regions, including those in maritime and continental climates as well as those at different latitudes, should be investigated. It is important to improve the existing knowledge of polar ice masses, perhaps through a network of key areas. In addition, calving losses are significant in some regions and will need to be included in mass-balance

programmes. For practical reasons, surging glaciers, glaciers on volcanoes and glaciers with marginal lakes will need special attention. Water-balance investigations, along with glacier mass-balance programmes, will provide a basis for assessing the effects of glacier changes on the water cycle under changing climatic conditions. Such assessments are especially important in arid inland environments such as the Tibetan Plateau where water resources are significantly influenced by high-mountain glaciers. Similarly, glacier mass balance and fluctuation programmes will assist in assessing the effects of glacier variations on other natural systems downstream. Geomorphic processes and changes to channel stability, for example, are important to determine the ecological state of riverine systems. The freshwater-saline water balance of tidal estuaries below glacierized catchments will, too, be affected by changes in glacierization. It is essential that glaciers be viewed by glaciologists and those working in related fields as part of the 'environmental system'.

Recent developments in glacier modelling now make it possible to compare the response of different glaciers to climatic change in a straightforward way. This will greatly facilitate the climatic interpretation of long-term records of glacier length. An attempt should be made to extend such series back in time.

For global coverage, remote-sensing techniques must become more widely applied to phenomena, with an appropriate signal-to-noise ratio such as cumulative length change. Degree-day (or temperature/precipitation), energy-balance and flow models should be used in combination with remote sensing data and digital terrain information in order to extrapolate the results from the small number of measured glaciers to larger glacier networks and to investigate feedback mechanisms. Monitoring of large glaciers in Patagonia, Alaska or in the Arctic etc., requires the development and operational use of new technologies such as airborne and/or satellite laser altimetry combined with kinematic GPS, gravimetry or earth-rotation analysis. In order to achieve this, collaboration with sister organizations in the fields of geophysics, geodesy, remote sensing, etc. must be intensified. Fast distribution of data by electronic media has developed quickly over the past years. Glacier inventory and recent fluctuation data are available in digital form via Internet, the World Data Centers and UNEP's Global Resources Information Database. Further work needs to be done to facilitate the exchange of digital data. However, speed of communication is only one criterion. Aspects of authorship, data safety, quality, compatibility, continuity etc. are important and must be considered seriously.

The large ice sheets (Antarctica and Greenland) and the largest mountain glaciers of the world play an especially significant role in the global climate system. However, the mass balance and fluctuations of these huge ice masses must be measured and the results reported according to standards and formats which will necessarily differ from those developed for smaller glaciers. Efforts must be made to explore how ice-sheet data could be compiled. This task could begin with a workshop, which may produce recommendations as well as a survey of existing observational programmes, and should be undertaken in cooperation with the Scientific Committee on Antarctic Research (SCAR), the International Arctic Science Committee (IASC) and other relevant organizations.

The specific characteristics of surging glacier fluctuations and the practical significance of this phenomenon, as well as of other special events such as calving instabilities, rockfalls onto glacier surfaces, eruptions of glacier-covered volcanoes, outbursts of ice-dammed lakes, ice avalanches or periglacial debris flows demand the working out of particular observational programmes and forecasting methods.

Experience shows that the programme of worldwide glacier monitoring needs:

- active coordination with respect to internationally collecting standardized high-quality data;
- systematic development of monitoring strategies for reaching clearly defined goals; and
- open communication between all people and organizations involved.

This is a fascinating and important but also increasingly difficult task, especially as concerns the growing discrepancy between the importance of glacier signals as a reflection of global-climatic change and the funds available for long-term monitoring programmes. In this situation, the need arises for a more widespread distribution of responsibilities and further development of advisory services linked with the World Glacier Monitoring Service (WGMS) and the programme as a whole. The advisory function is fundamentally important in that it must guarantee the quality of the products and protect the prestige of all organizations involved. Clear interrelations should exist, for instance, between the International Commission on Snow and Ice (ICSI/IAHS) as the parent organization, the scientific consultants to WGMS and external reviewers. External review should take place regularly, perhaps every three to five years or so. The WGMS – and indeed the entire programme – needs partners who can help develop and bring to fruition a vision for coming years and decades of the project as a whole in a worldwide and multidisciplinary context. Thereby, the interests, responsibilities, and merits of the various national and international organizations and funding agencies involved must be carefully considered. The WGMS objectives should be more closely-related to those of the Arctic and Antarctic science programmes.

The 100-year old programme of worldwide glacier monitoring continues to be active because of the feeling of professional and scientific responsibility and the solidarity and openness among those who

collect and compile the information under sometimes quite difficult conditions. This outstanding international collaboration, together with the quality of the products and the importance of the observed natural signal, provide the real strength of the programme. The scientific community, in conjunction with international organizations, should take care to preserve this precious treasure.

These conclusions and recommendations were discussed and agreed upon during an expert meeting at the ETH in Zurich, 12-13 October 1995 (cf. appendices for programme and list of participants).

Appendices

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II Consultants of WGMS

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IV Programme of Expert Meeting on World Glacier Monitoring 12–13 October 1995, ETH Zurich, Switzerland

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ETH Zürich, VAW Glaciology, 8092 Zürich
12/13 October 1995

IV. Expert meeting on world glacier monitoring

12-13 October 1995, ETH Zurich, Switzerland

PROGRAMME

Wednesday, 11 October

Arrival at Zurich, informal evening meeting (18.00 at the Hotel Arc Royal, breakfast room)

Thursday, 12 October

VAW/ETH - E32: UNESCO publication - overview chapters: discussion/conclusions

0900 - 1045	Regional chapters (continents)
coffee break	
1115 - 1300	Thematic chapters
lunch break	

ETH (Main Building - HG G5): Public presentation and reception

1430 - 1500	Addresses ETH, UNEP, UNESCO, FAGS, ICSI
1500 - 1530	The World Glacier Monitoring Service (W. Haeberli)
1530 - 1600	Global overview (V. Popovnin)
1600 - 1630	Modelling and statistics (H. Oerlemans)
1630 - 1700	Perspectives for the future (M. F. Meier)
1700	Cocktail reception at ETH

Dinner on Uetliberg for invited participants

Friday, 13 October

ETZ/ETH E81: UNESCO publication - recommendations chapter: discussion, editing

0900 - 1045	discussion recommendations chapter: science
coffee break	
1115 - 1300	discussion recommendations chapter: funding and administration
lunch break	
1430 - 1700	final editing of recommendations chapter: decisions on further immediate actions

Saturday, 14 October

Full-day excursion to Jungfraujoch for invited participants:
Alpine glaciers; high-mountain construction in snow, ice and permafrost

Sunday, 15 October

Departure

International co-ordination of long-term glacier observations is a century-long tradition that began back in 1894 with the establishment of the International Glacier Commission in Zurich, Switzerland. Over the past century, the goals of internationally co-ordinated glacier monitoring have changed somewhat and multiplied. Today, the evolution of glaciers and ice caps is recognized as being one of the key variables for the early detection of possible man-induced climatic change. The general shrinkage of mountain glaciers during the 20th century is a major reflection of the fact that rapid secular change in the energy balance of the Earth's surface is taking place on a global scale.

This volume opens with the facsimile of an article written in 1894 by F.-A. Forel, President of the International Glacier Commission, followed by thematic chapters dealing with glacier monitoring, data handling, modelling and remote-sensing techniques, as well as a selection of regional accounts. Characteristic examples are given from all continents, including the special cases of the continental ice sheets.

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Cover photo: Vatnajökull Glacier (Eric Frogé)