INTERNATIONAL HYDROLOGICAL PROGRAMME

Variations of Snow and Ice in the past and at present on a Global and Regional Scale

Edited by

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FOREWORD

The role of snow and ice is particularly important in the formation of regional runoff and in its influence on global climate. Both glaciers and perennial snow core have not been sufficiently studied in this respect, and there is the necessity to employ new approaches in order to better understand glacial processes and phenomena.

To study the role of snow and ice in the global watercycle and in the world climatic system, research methods should be further standardized and information from different regions integrated. Representing these field of research, the International Commission on Snow and Ice (ICSI) has already made some guideline studies on the runoff prediction and evaluation. The first glacier mass balance bulletin has been recently issued and the World Atlas of Snow and Ice Resources is near completion now.

The present document summarizes the latest achievements in snow and ice research undertaken in view of the main objectives of the International Hydrological Programme. The document was prepared under the guidance of the ICSI Bureau, elected in 1987-1991. The staff includes: Vladimir Kotlyakov - President, Charles Bentley, Michael Kuhn and Xie Zichu - Vice Presidents, Bruno Salm - Secretary, Elizabeth Morris, Steve Ackley, Keiji Higuchi, and D.J. Klinger - Chairmen of divisions,

The structure of the document is as follows (the authors of the Chapters are mentioned in brackets):

The materials presented here are of special interest for hydrologists, climatologists, glaciologists, geographers, and geophysicists.

Professor V.M. Kotlyakov

Editor

Moscow, October 1991

1. New data on global snow and ice resources

The data on the world's glacio-nival resources are based on the materials of the recently compiled World Atlas of Snow and Ice Resources. These are subdivided into seasonal and perennial snow and ice. The seasonal resources comprise the moisture accumulated in seasonal snow covers on the land surface and on "marine" glaciers, as well as the seasonal sea ice. In the category of perennial resources are the perennial Ómarine" glaciers, as well as the inland and ground glaciation, i. e. the permanent land glaciers and the ice of permafrost. Icings occupy an intermediate position: Although a small portion of them maybe categorized as a perennial formation, they are a seasonal phenomenon by and large.

Aggregate Data

The total mass of ice, generated in the atmosphere (counting the snowflakes evaporating in it), is equal to 17 x 10^{12} tons per year. 11 x 102 tons of this snow is deposited on land, and 6 xl 0^{12} on the ice surface of bodies of water. The annual turnover of moisture involves about 3.4x1 0^{12} tons of snow, of which 2.6 x 10^{17} tons are discharged, together with icebergs, into the ocean where it melts, consuming nearly 9 x 10^{14} MJ/ year, while the **remaining 0.8 x 10^{12} tons** melt on the land surface.

The overall mass of sea, lake and river ice is equal to 30 x 10^{12} tons; 6 x 10^{12} tons of snow are added from the atmosphere. About 10 percent of the ice, i. e. approximatively 3 x 10^{12} tons are contained in river ice jams and in sea ice hummocks, an amount commensurate with the annual volume of glacier nourishment. Finally, the resources of ground ice are about the same as those of sea ice; however, the annual increment of ground ice can hardly be more than 3 x 10^{8} tons, i. e. 1/1000th of the total volume. This increment may be estimated as follows. The ground ice area makes up about 10^{6} km^{2}. The characteristic width of ground ice cracks is around 1 cm, with the distance between them being 15 m; consequently, the area of ice cracks is 0.66 x 10^{3} of that occupied by the ground ice. The layer of snowmelting above such cracks may be estimated to be 5 - 10 cm of water equivalent. Assuming that half of this water is utilized for the ground ice increment, we obtain 3 x 10^{6} tons. The melting of the ground ice at the present stage of its degradation may be an order of magnitude higher. Under steady-state conditions the complete turnover of the ground ice my take about 100.000 years; however, under present-day conditions ground ice may disappear in ca. 10.000 years.

The seasonal Snow Cover

In average years snow cover is formed over an area of 110 x 106 km^{2} (Kotlyakov, 1987; Shumsky, Krenke, 1964). The wide extent of the snow cover exerts a significant effect on the global climate; it increases the surface albedo values, and causes a great amount of energy to be expended on melting and on the seasonal delay of the water turnover.

Model experiments have shown the existence of a climatic mechanism stabilizing the snow cover limit under the conditions of present-day relief (Turkov, 1988). The limit of snow cover extent is determined by a balance between the positive (regenerative) feedback of the air temperature decrease due to the cooling effect of snow, on the one hand, and the negative (degenerative) feedback of the snow cover propagation equatorward to the increasing circulation of the atmosphere at the snow cover limit (which leads to a poleward flow of tropical air and, as a result, to snow melting), on the other. These limits must be sufficiently stable within a wide range of climatic changes, perhaps even during glacial periods.

The mean layer of global snow contains 20 cm of water equivalent (w. e.) varying from 10 cm to 40 cm, the continental snow cover (14.7 cm w. e. on the average) holds only half as much as that on sea ice (29.2 cm w. e.). The area of solid precipitation over the ocean is significantly larger, in terms of space and time, than that on sea ice: ca. 1/3 of the solid precipitation melts on the water surface. Since the continental snow cover is twice als large as that on sea ice (72 km^2 x 106 krn^2 and 36 km^2 x 106 km^2, respectively), the total volumes of snow reserve - 10.5 x 103 km^3 - are equal in either case.

The volumes of water held in snow cover in the southern and northern Hemisphere are about equal -10.3 x 103 and 10.8 x 103 km^3. The area of the seasonal snow cover in the southern hemisphere is 35 x 106 km^2 and that in the northern hemisphere 73 x 106 km^2; the snow cover area on sea ice is 19 x 106 km^2 and 17 x 106 krn^2, respectively, and that in continental regions 16 x 106 km^2 and 56 x 106 km^2 for the southern and the northern hemispheres, respectively. The layer of snow w. e. is twice as thick in the southern hemisphere (30 cm w. e.) as in the northern hemisphere (15 cm w. e.). The snow cover w. e. volumes on sea ice are distributed as follows: 34 percent in the Pacific Ocean; 31 percent in the Atlantic Ocean, 18 percent in the Arctic Ocean and 17 percent in the South Indian Ocean (Kotlyakov et al., 1991).

On continents the snow water equivalent decreases with an increase in the snow cover area (Table 1).

Table 1: *The seasonal snow water equivalent*

Continent	Area km’	Depth (cm w. e.)	Volume km^3
Eurasia	35,500,000	11.4	4,060
North America	18,200,000	20.5	3,730
South America	1,300,000	30.0	390
Antarctica	14,100,000	15.6	2,200
Africa	under 100	--	under 1
Australia	under 100	--	under 1
Total	69,100,000		10,380

Due to direct transfer of water vapor from ocean to ocean, 2,000 km^3 (or 25 percent) of snow water resources on sea ice in the Southern Ocean is formed by solid precipitation brought in form neighboring oceans.

The maximum redistribution of solid precipitation carried from the Atlantic Ocean to the continents is observed in Eurasia and in North America. Of the 4,880 km^3 of the Atlantic solid precipitation, only 1,260 km^3, or 26 percent, fall in the Atlantic Ocean basin; 3,000 km^3 or 61 percent, are a source of runoff to the Arctic Ocean; 500 km^3 or 10 percent, are channeled into inland basins; 20 km^3 go to the Pacific, and 1,000 km^3 to the Indian Ocean. The percentage of the Pacific-derived solid precipitation redistributed in Eurasia and North America is less -850 km^3, or 31 percent of 2,750 km^3; 660 km^3 are diverted to the Arctic Ocean and 200 km^3 to the Atlantic. Of the 200 km^3 volume of the solid precipitation derived form the Indian Ocean, 50 km^3 are channeled into inland basins, and 30 km are transferred to the Pacific Ocean. All told, a fraction of 58 percent of the 7.8 x 103 km^3 of solid precipitation transported to the continents of the northern Hemisphere does not return to its “home” ocean.

In the southern Hemisphere, the volumes of solid precipitation on the South American and Antarctic continents and their percentage redistribution are lower: of the total 2.6x10³km³, 74 % fall in the basin of a source ocean. For the Atlantic-derived moisture, the share of redistribution is equal to 27% of the total 410 km’; for the Pacific precipitation - 31% of 750 km^3. Altogether, 50% of the solid precipitation brought to the continents is transported via oceanic watersheds so as to become a source of runoff to other oceans.

Glaciers and Ice Sheets

At the present time, glaciers occupy an area of 16.2 x 10^6 km² or 10.9 percent of the land surface, and have a mass equal to 98.95 percent of the entire mass of ice on the earth; this is nearly 32 times as much as the mass of the surface (land) water. The total glacier-derived runoff (2.5 x 10^{18} g/year) constitutes only a *7* percent of the river runoff despite the fluvial water. The point is that ice flows at a rate hundreds of thousands of times slower than that of water. The turnover period for glacial ice averages 9,600 years, with the maximum period in Central Antarctica exceeding 200,000 years.

Table 2 lists data on the extent of glaciation in different parts of the globe, based on the "Encyclopedic Dictionary of Glaciology" (1984, in Russian). These data are based on glacier catalogs, maps of different scale and literary evidence for specific regions - this information is condensed in the World Atlas of Snow and Ice Resources.

According to the World Atlas of Snow and Ice Resources, the bulk of the perennial ice is concentrated in Antarctica, and its water equivalent has been estimated at 23,296,630 km³. For North America and Greenland, the figure is 2,431,773 km³. Quite unexpectedly, Europe is found to have a significant mass of ice 21,082 km³; but there is relatively little ice in Asia - 16,260 km³; evidently, the proximity of ocean nourishment sources must be playing an important role in the formation of large volumes of perennial ice. The South American Andes contain 12,690 km³ of ice, while Africa's mountains just under 0.4 km³.

More than 50 km³ of perennial ice has been detected on the largest islands of Oceania in the temperate and tropical belt, and some islands in the zone of icebergs in the Atlantic and Indian Oceans are found to have over 500 km³. 85 percent of the present-day ice mass exchange is provided by continental ice sheets - 2,900 km³ / year of the total 3,450 km³. An analysis of accumulation at the equilibrium line on the global and regional scales has revealed the following characteristics (Krenke et al., 1991):

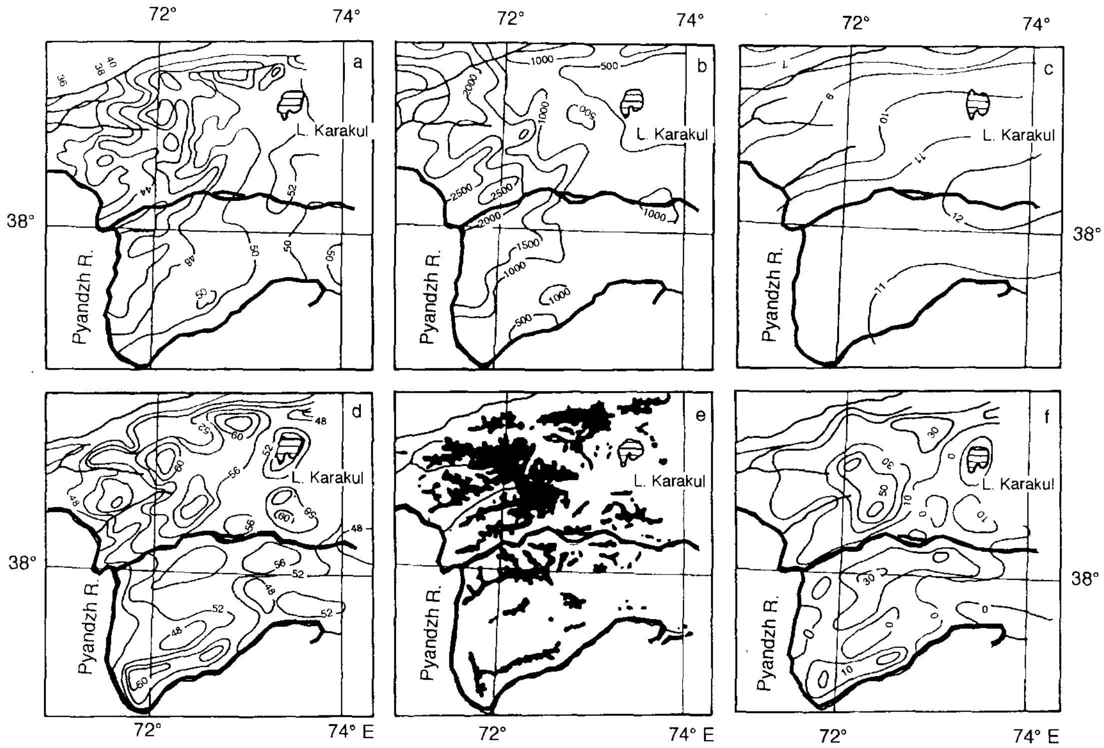
The glacial accumulation values increase from high to low latitudes concomitantly with an increase in the content of atmospheric moisture. These values rise from 25 g/ cm² in the subpolar zone to 100 g/cm² in the north of the Eurasian and American continents, and approximately to 200 g/ cm2 on the average in latitude 40° at the northern periphery of mountain systems, and to 250 g/cm² at 30° N, on the southern slopes of the Himalayas.

The accumulation field can be subdivided into structural parts depending on the distribution of continents and oceans. At the western windward periphery of the American continent, the ice accumulation value increases threefold from the interior toward the ocean. At the western periphery of the Eurasian continent it increases twofold - from 150 g/ cm² to 300 g/ cm² with the approach to the shore in the direction of the moisture flow; and on the eastern edge of the Eurasian continent, if we are to judge by the accumulation field, precipitation of Pacific origin is prevalent.

The accumulation field of the northen Hemisphere is marked by a Central Asian minimum. In the heart of Tibet, accumulation even from the leeward side of the Pamirs declines to 25 g / cm', i. e. a value characteristic of subpolar regions. A schematic representation of the northern Hemisphere's accumulation field shows a distinct connection between enhanced accumulation values (resulting in high intensity of glaciation) and climatological frontal zones.

An important feature of accumulation fields within mountainous countries is that they are distinguished by a "stream-like" pattern reflecting the hydrodynamics of moisture flows over the complex mountain relief. The intensity of atmospheric moisture streams in valleys is determined by the openness of the valley inlets and the straightness of the valleys. Streams of moisture get over large mountain passes, too.

The data contained in the WorId Atlas of Snow and Ice Resources (including the maps of glacial systems) have made it possible to derive the general characteristics of glacier systems given in Table 2. They show considerable differences in glaciation areas - from 1.5 x 103 km² in the Caucasus to 90 x 103 km² in Alaska, while the range of glaciation altitudes is fairly stable. Only in one system, the Rocky Mountains, is this range equal to 300 meters, and in the Alps and in the Coast Range it is 800 m to 900 m, while in the four others it is between 1400 m and 1800 m. Despite the variety of climatic conditions, the glaciation region in each mountainous country, with the exception of Tien Shari, occupies a strictly defined climatic zone - the mean summer temperature is equal to 3 to 4.5+/- 1.5 to 2 °C.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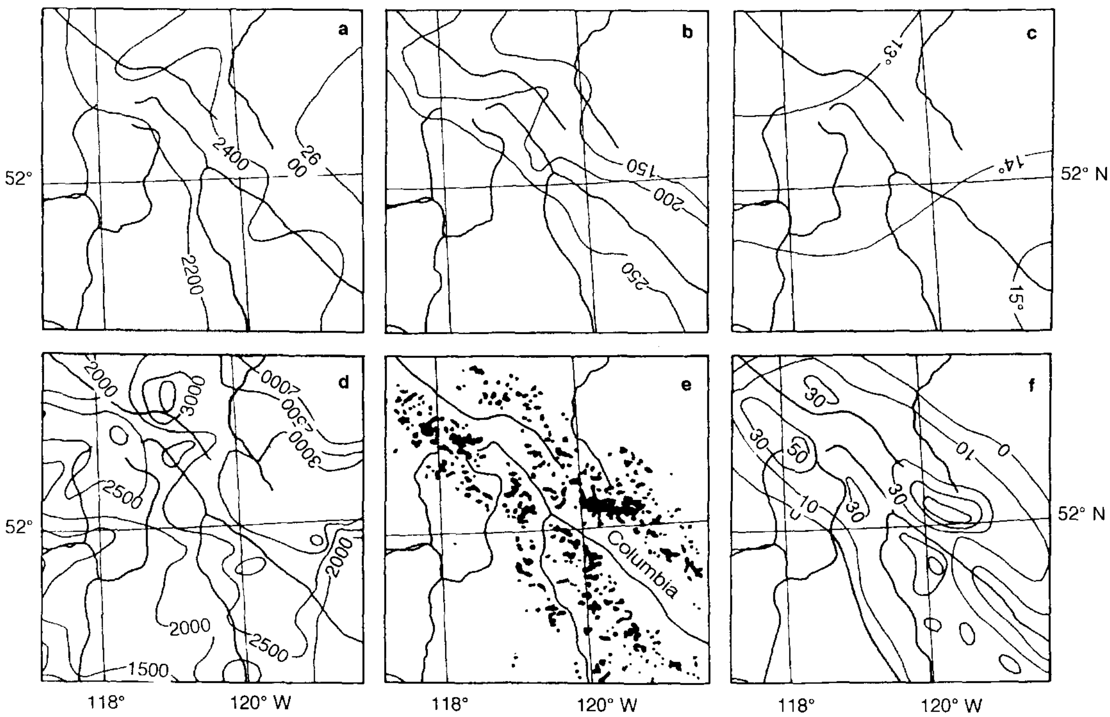
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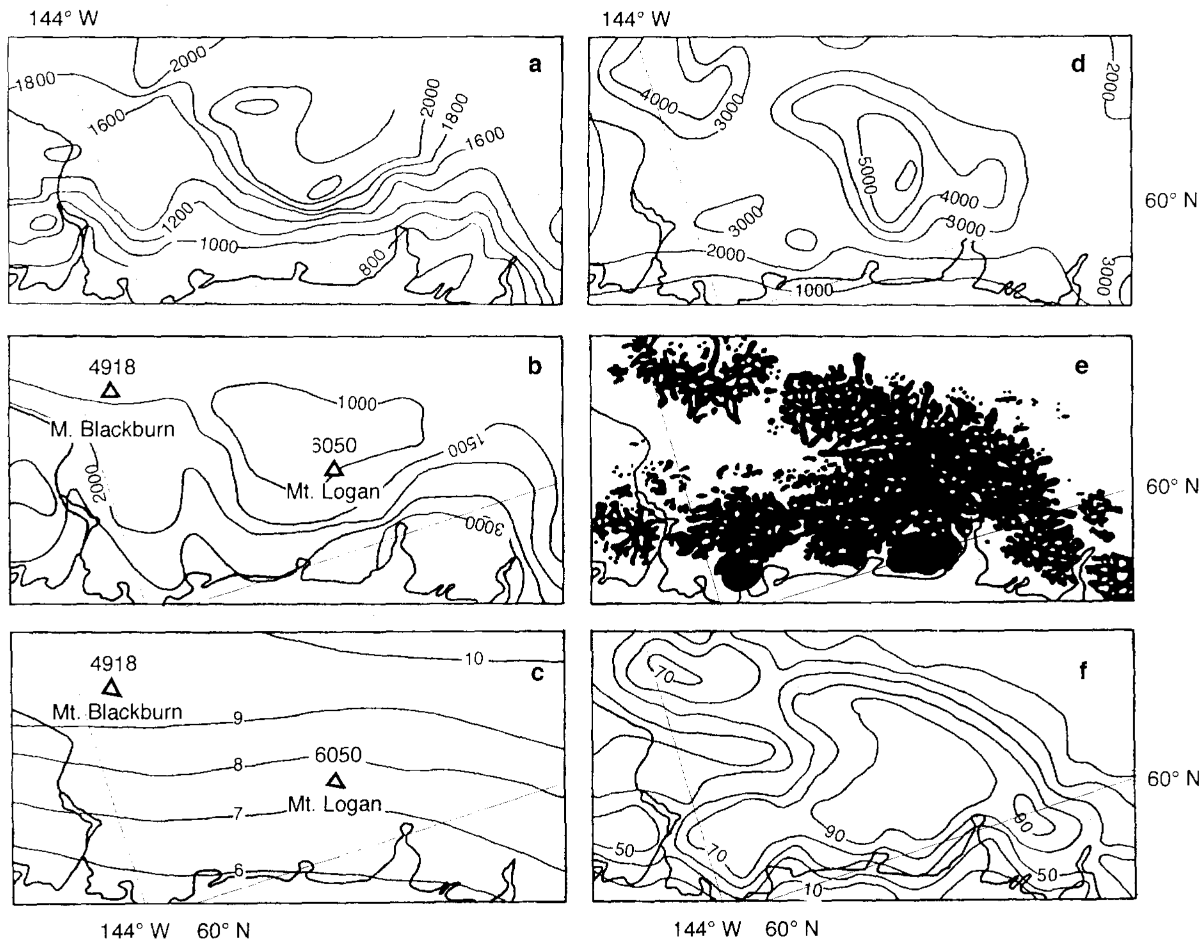


Fig 1: The Pamirs (A), Rocky Mountains (B) and South East A1aska (Wrangell Mountains and St. Elias Mountains)(C). a - the equilibrium line altitude in m, b - accumulation in mm, c - the mean summer temperatures of the air at the altitude of 3500 m (A) and 1000 m (B and C) in °C, d - maximum relief of summits in m, e - outlines of glacierization, f - degree of glacierization in %.

In mountain systems one differentiates among dispersed, semidispersed and semicompact glaciation (Khodakov, 1978). Examples corresponding to this effect are given for the glaciation of the Rocky Mountains, the Pamirs and southeastern Alaska, respectively (Fig. 1). Equilibrium line altitudes are a comprehensive indicator of climate (Krenke, 1982). Accumulation conditions are represented on the glacier regime charts by equilibrium line altitudes, while ablation conditions in a particular region are characterized, in general terms, by a mean summer air temperature at fixed altitudes. Finally, the generalized data on summit altitudes and glaciation extent reflect the role of orography in glacier formation.

Glacier-Derived Runoff and the Climatic Role of Glaciation

The total glacier-derived runoff, equal to 3,450 km^3 of water, accounts for about 8 percent of the total surface water runoff to the ocean (44,700 km^3) and for about 3 percent of the land precipitation (119,000 km^3). Of this amount, Antarctica and Arctic islands contribute 3,010 km^3, and mountain glaciers - 440 km^3, or 1.0 percent of the land surface runoff. Nearly another 1 percent (370 km^3) is contributed by the seasonal runoff of glacier snow in summer periods in the absence of melt-water runoff from ice-free areas; the latter factor is very important in enhancing the value of glacier-derived runoff as a water resource.

Snow cover, sea ice and ice sheets cool the surface of the earth by increasing the albedo value. Ice cover reflects about 5 x 1016 MJ of solar energy into outer space annually, or 3.5 percent of the solar radiation reflected from the earth. Owing to this additional reflection, the planetary albedo value increases by 0.02 in summer and by 0.01 annually, from 0.29 to 0.30 which, according to M. Budyko's energy budget model, results in a - l°C cooling of the surface air layer.

The aggregate radiation balance of glaciers constitutes -4.8 x 1015 MJ/year, or 0.2 percent of the solar energy absorbed by the terrestrial surface. The annual expenditure of energy on ice melting averages ca. 3 x 1014 MJ, i. e. considerably less than the amount of heat transfer to compensate the radiation balance. Overall, the transfer of turbulent heat from the atmosphere to glaciers is estimated at 51 x 1014 MJ/ year. This amount is sufficient for cooling as much as 5 x 1018 m^3 of air by 1.0°C at a pressure of 750 hPa, i. e. the volume of the entire troposphere of the earth. Consequently, glaciation cools the present-day troposphere by about 1°C through the heat transfer from turbulent heat flows, of which 6 percent is expended for glacier melting, with 94 percent being reflected by ice sheets, mainly in Antarctica.

In all, the global expenditure of heat for iceberg melting - a "runoff of cold to the ocean" - is above 8 x 1014 MJ/year. This value is comparable with the heat transfer by ocean currents. The glacier-derived „runoff of cold" affects both the atmospheric and the oceanic circulation.

So, the total volume of perennial land surface ice on continents and islands is estimated at 25.8 x 106 km^3. About 0.01 percent of this volume is renewed annually; annual accumulation and ablation, including iceberg calving, contribute 3.5 x 103 km^3; seasonal snow resources account for 20 x 103 km^3. Ground ice contributes 0.5 x 106 km^3. Snow cover, sea ice and glaciers thus exert a substantial effect on global climate.

2. Global changes over the last climatic cycle according to ice core data

The past of our natural environment is one of the keys to understanding present-day and future changes in the climate, and anthropogenic effects on it. New possibilities in this respect are offered by studies of ice cores extracted from deep boreholes drilled in ice sheets of the polar regions. A comprehensive analysis of such samples makes it possible to determine changes in the composition of the atmosphere an temperature over hundreds of thousands of years.

The drilling of one such borehole begun by a Soviet Antarctic expedition at Station Vostok early in the 1970s, has already brought significant results. Vostok Station is situated in the central part of East Antarctica at a height of 3,490 meters above sea level. The mean annual temperature there is -55.5° Celsius and the mean annual accumulation rate is 23 millimeters (water equivalent). The ice sheet thickness is close to 3,500 meters, which means that it is made up of ice deposited over hundreds of thousands of years, a factor of immense importance for paleontological reconstructions.

Thus far five deep boreholes have been drilled in the vicinity of the station Vostok, with two sunk to a depth of 2,500 meters. Russian and French researchers have joined hands in this undertaking: those from the Geography Institute of the Russian Academy of Sciences, the Arctic and Antarctic Institute, and form the Environmental Glaciology and Geophysics Laboratory of Grenoble University. Comprehensive studies of a 2,200-meter-deep core sample have now been carried out.

The dating of the ice is effected by mathematical modeling of ice flow (Kotlyakov and Gordienko, 1982). One proceeds from such initial data as the snow accumulation rate, the temperature and viscosity of the ice, the velocity of ice movement, and bed relief features. Our models assume steady-state ice flow and the absence of major gaps in columns, which conforms well to conditions in the neighborhood of Vostok Station.

The rather simple relation between the depth and the age of the ice in the vicinity of the Vostok Station is due to the fact that the bottom of the borhole is still far above the basal layers of ice in which all kinds of aberrations, caused by ice movement and deformation, are possible. The dating of ice at a depth of 2,200 meters (according to our computations, its age ought to be slightly over 160 thousand years) is determined with an accuracy of 10 to 15 thousand years. Isotope - geochemical investigations have been made through the entire column of ice; analyses by other techniques have been conducted on 1.5 to 2-meter-long cores collected every 25 meters.

The Vostok core samples thus encompass the Holocene (the last 10,000 years), the late Pleistocene Valdai or Würrn Glacial Epoch (10,000 to 120,000 years ago), the Mikulino or Riss-Würm Interglacial (120,000 to 130,000 years ago) and the final stages of the Dnieper or Riss Glaciation.

The main method of paleotemperature determination involves analysis of the stable isotopes ratios H/D and $^{16}O/^{18}O$ in ice. The isotopic composition of deposited snow depends on its formation temperature. It has been found experimentally that a decrease in the content ^{18}O in East Antarctic ice (expressed by the quantity $\delta^{18}O$ which denotes the relative deviation from the isotope composition of the standard mean of sea water) by 1 percent corresponds to a cooling 1.5° C whereas a 6 percent δ D shift will correspond to a temperature drop of 1°C. Using these correlations, we may transform an isotope curve into one of paleotemperature.

The isotope curves for ice cores obtained form deep boreholes indicate a dramatic shift at the boundary between the late Pleistocene and the Holocene. In Greenland the shift amounts to 12 percent at Camp Century, and 7 percent at Dye-3; the reason for this difference is that during the Pleistocene the height of the ice sheet at Camp Century station changed far more than in south Greenland. Applying a correction to the isotope data for changes in the glacier height across the transition to the Holocene, we find that the temperature rise in the polar regions 10 to 11 thousands year ago averaged about 10° C.

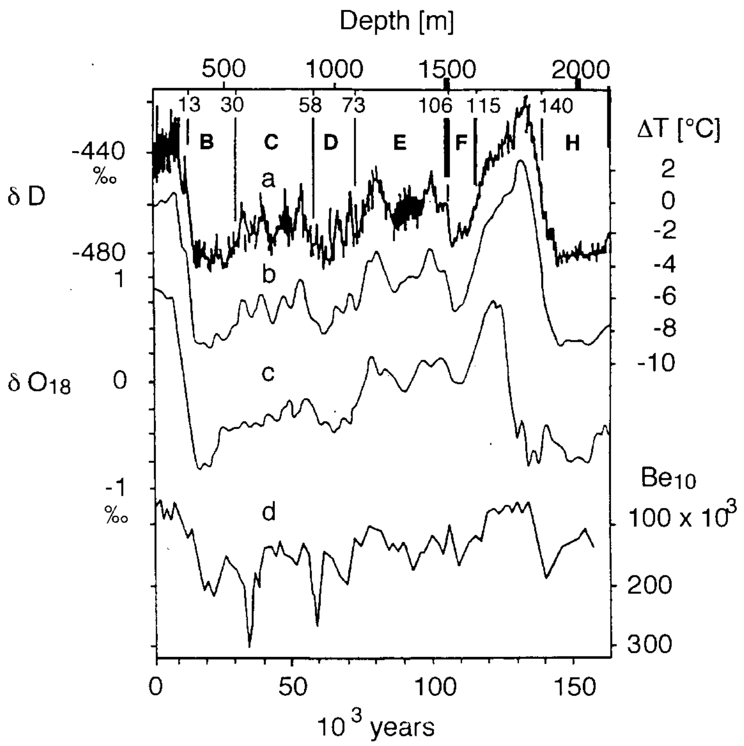


Fig. 2: Ice core data obtained from the Vostok Station (Jouzel et al., 1987; Lorius et al., 1985; Raisbeck et al., 1987):

a) Deuterium content;

b) Deviations from the present-day temperatures on the ice surface;

c) Oxygen-isotope from deep-sea sediments;

d) Be_{10} concentration in ice.

Capital letters designate climatic stages characterizing warrn and cold periods: A and C are the Holocene and the Mikulino Interglacial, B - F the Valdai Glacial Epoch where C and E indicate relatively warm interstadials in glacial

The isotopic profile of ice from Vostok Station (Fig. 2) practically undistorted by ice flow conditions, provides a comprehensive picture of the temperature pattern in the polar region over 160 thousand years: firstly, the peak of the Ultimate (Last) Interglacial was approximately 2°C warmer than the Holocene; secondly, the final stages of the Dnieper (Riss) Glaciation was as cold as the maximum of the Valdai (Würm) Glaciation; thirdly, the transition from the Dnieper Glaciation to the Mikulino Interglacial saw a 12°C temperature increase; and fourthly, the Last Glacial Epoch reveals three clearly distinct temperature minima separated by two periods when the temperatures were 4°C and 6°C higher, respectively, than in the Late Pleistocene.

Atmospheric precipitation is another important climatic parameter. The rate of precipitation, being a function of the planet's temperature background, was not stable over the last 160,000 years. The present rate of snow accumulation on large ice sheets depends on the temperature above the inversion layer, i. e., above the ice-cooled air. The amount of water vapor there is strongly dependent on the temperature of the surrounding air; that is to say, the atmospheric moisture content shows a dramatic drop with a temperature decrease. A simple calculation suffices to show that, under the colder conditions of the Glacial Epochs, the amount of atmospheric precipitation and, correspondingly, of snow accumulation on the glacier surface would have been 50 percent lower than its present value (Lorius et al., 1985).

An independent method for estimating snow accumulation is to study changes in the concentration of Be, $_0$, a radioactive isotope of Beryllium (Raisbeck et al., 1987). The concentration of this long-living cosmic isotope was the same in both Interglacial, but it increased during the Glacial Epochs - approximately twofold at the peak of glaciation (see Fig. 2). Since the flux of the Be_{10} isotope deposited on earth is constant, we conclude that the concentration of radioactive Beryllium in an ice core is inversely proportional to the snow accumulation rate. Thus both methods of snow accumulation computation - according to isotope temperatures and by radioactive Beryllium data - are in good agreement. This means that half as much snow was being accumulated in East Antarctica during the Glacial Epochs as at present.

The Glacial Epochs were characterized not only by global cooling - also, sharper temperature gradients developed, interlatitudinally and between continents and oceans; this increased the energy driving oceanic and atmospheric processes. The Glacial Epochs were thus characterized by stronger oceanic currents and winds than the Interglacial; there was a considerable strengthening of cyclonic activity at the boundary between the ice sheets and the barometric depressions surrounding them.

Additional evidence for stronger atmospheric circulation in the Glacial epochs arose when the ice cores were analyzed for the concentration of continental and marine aerosols, i. e., microscopic particles of micron size, capable of “floating”, in the atmosphere for al long time. Al aerosols are a typical indicator of a continental source, whereas Na aerosols are generally of marine origin. The Pleistocene portion of a core obtained at Byrd Station in West Antarctica showed a concentration of continental dust 8 times as high as that in the Holocene portion and the corresponding ratio for the Vostok cores was 30. The increase in the marine aerosols concentration was 2-3 fold and 5 fold in the Byrd and Vostok cores, respectively. (The record at Vostok are shown in Figure 3.)

Such fluctuations may be explained partly by a decrease in the snow accumulation rate in the cold epochs, but the main reason must have been stronger winds as a result of the sharper interlatitudinal contrasts in the temperature and barometric fields. A general encroachment of deserts into periglacial regions and onto continental shelves exposed by the eustatic lowering of the sea level, also as water was transferred to the ice sheets had a significant part to play.

As polar snow is transformed to ice, the atmospheric air is trapped in bubbles. Therefore, by extracting the gases contained in ice cores, we may obtain data on the composition of the atmosphere in the past, specifically, on the concentration of greenhouse gases. In the absence of melting, the closure of ice pores proceeds at a slow pace: in central East Antarctica this process may take as much as 4000 years, during which some exchange of air between the pores and the free atmosphere takes place. Consequently the air extracted from polar ice cores is younger than the one existant at the time when the snow that formed the ice, was accumulated. Present-day analytical procedures enable us to extract some gases from the ice—carbon dioxide (CO_2) and methane (CH_4) are the most important and measure them with great accuracy. Analyses of CO_2 in ice cores from deep boreholes have shown that the content of carbon dioxide at the maximum of the Valdai glaciation, (190 - 200 ppmv (parts per million by volume), was 25% lower than in the Holocene, when it increased to 260-280 ppmv. These levels maybe typical of glacial and interglacial epochs, respectively, in general.

Measurements were made at 66 depths, which corresponds to a difference of 2 -4.5 thousand years between two neighboring levels. A scatter of data is shown for the curves representing the CO_2 and CH_4 dynamics; the age scale is obtained by taking account of the period of air bubbles closure with the conversion of firn to ice.

The data obtained from the Vostok borehole show an excellent correlation between the CO_2 concentration and the isotopic paleotemperature (Fig. 4) This is the first direct indication of the tight dependence between the level of atmospheric CO_2 and the changes of the climatic cycle, Yet there are some mismatches, It is noteworthy that, whereas the changes in CO_2 concentration and temperature are synchronous during the transition from a glacial epoch to an interglacial epoch, during the transition form relatively warm intervals to colder ones (e.g., about 115 and 75 thousand years ago) there is a lag in the carbon dioxide concentration decrease compared with the temperature drop.

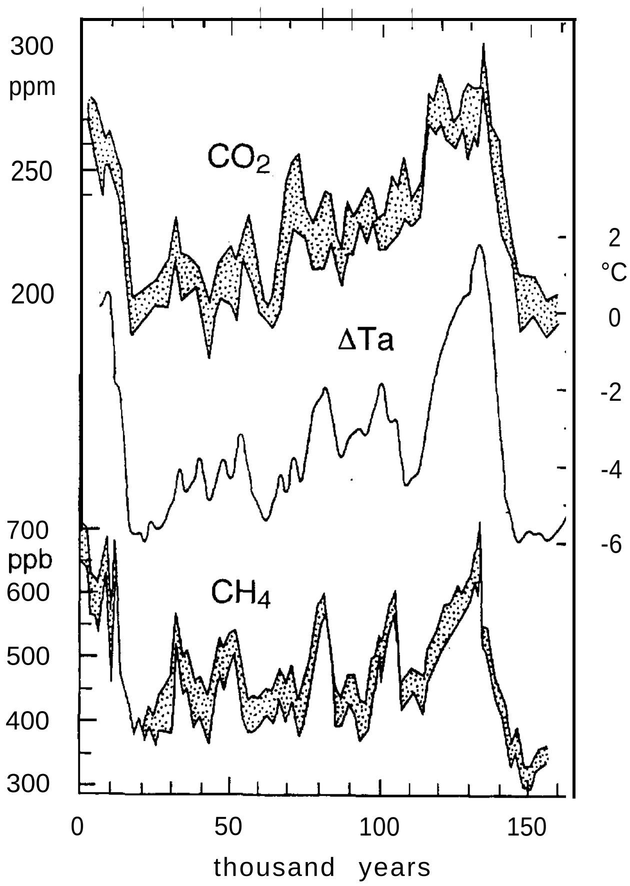


Fig. 4: Carbon Dioxide and Methane concentrations and paleotemperature deviations T_a from the present-day temperature values Vostok Station, Antarctica (Barnola et al., 1987; Chappelaz et al., 1990; Jouzel et al., 1987).

The changes in the CO, concentration and temperature though the glacial-interglacial cycle attest to some cause-and-effect linkages between the two phenomena. It is not clear, however, which is the cause and which the effect. Many researchers assign the role of causation to the carbon dioxide concentration; yet the above-noted lag of CO_2 changes at the end of interglacial points to the temperature changes as the cause of the CO_2 change which, in its turn, intensifies the process. The ice core data from Vostok Station suggest that different stages in the history of climatic changes must have involved different mechanisms of interaction between temperature and carbon dioxide cycles.

The concentration of the other carbon compound, methane (CH_4), in the paleoatmosphere is likewise coupled to temperature. Dramatic changes in methane concentration occurred at both glacial-to-interglacial transitions, that is, 150 - 135 and 18 - 9 thousand years ago. The concentration of methane displayed a dramatic increase from 0.35 ppmv at the peak of glaciation to 0.6 - 0,7 ppmv in the interglacial optima (Reynaud et al., 1988). Four distinct CH_4 maxima occur in the Valdai Glacial Epoch - these maxima, associated with relatively warm intervals, are less clear and somewhat displaced in time in the $C0_2$ record.

The cause of such discrepancies lies probably in the different sources of atmospheric carbon dioxide and methane. Whereas the CO_2 concentration in the atmosphere largely depends on the continents; biosphere and the oceans, the sources and sinks of methane are on the wetlands, deposits of hydrocarbons (including gas-hydrates), colonies of termites and so on. One of the causes of the drastic increase in the amount of atmospheric CH_4 at the close of the Glacial Epochs may have been that the gas, contained in giant gas-hydrate deposits buried on polar continental shelves has, over the past 100 years, become the target of modification by human activity. The results are already apparent in glacial records. The anthropogenic concentration of nitrates and sulphates in snow accumulating on glaciers is on the increase: in the last 100 years, the concentration of SO_4 anions in ice has gone up 3-4 fold and there has been a steady growth in the amount of NO_2 admixtures since 1950 -- the mass of this deleterious substance, contained in automobile exhaust fumes, has already doubled.

A comparison of the present-day concentration of greenhouse gases with the ice-core data for the pre-industrial period shows an increase in the past 200 years: 25% for CO_2, 100% for CH_4, and 8 - 10% for NO_2. This increase in the concentration of NO_2, is fully consistent with the data on the combustion of petroleum fuels.

Yet another important bit of evidence from the isotope profile of the Vostok ice core: it clearly shows temperature oscillations with periods of 100,000 and 40,000 years and, to a smaller extent, with a period of 20 - 21 thousand years. These temperature oscillations are compatible with fluctuations in summer insolation at the high latitudes of the Northern Hemisphere within the last 160 thousand years. The data from the Northern Hemisphere are particularly important, for they confirm the well-known hypothesis of Milankovic according to which the growth and decay of insolation at 65°N (Fig. 5) regulates the glacial cycles.

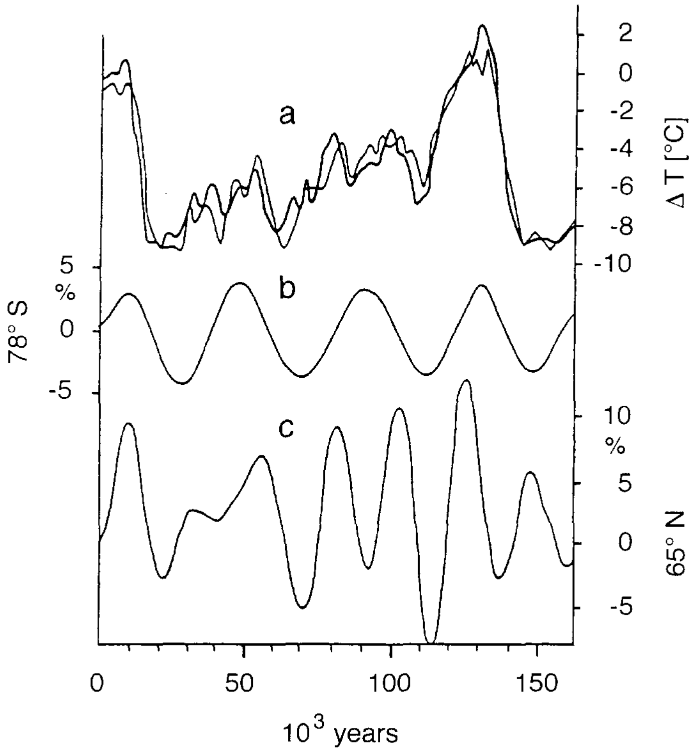


Fig. 5: a) A standard isotopic temperature course for an ice core from the Vostok Station and the temperature modelled from annual insolation at 78 °S, and July insolation at 65 °N, and the contribution of atmospheric CO_2; b) annual insolation at 78 °S; c) July insolation at 65 °S; both magnitudes are computed as standard deviation in percent, from the percent-day values (Berger, 1988; Genthon, 1987).

The paleotemperature oscillations in the Vostok ice cores must be a climatic reflection of the astronomic factors analyzed by Milankovic. Changes with a period of 20 thousand years are the effect of the precession of the equinoxes; those with a period of 40 thousand years, result from changes in the tilt of the Earth's axis of rotation; and those with a period of 100 thousand years sterm from the fluctuations of the eccentricity of the terrestrial orbit, a phenomenon enhanced in its effect on the Earth's climate by some unknown factors.

It should be taken into consideration that the glacial-interglacial fluctuations are coupled to fast feedbacks by the presence of water vapor in the atmosphere, by the properties of the cloud cover, by the characteristics of the seasonal snow cover and by the extent of sea ice. Long-term effects likewise have a role to play; these are related to slow changes of some boundary conditions and in atmospheric composition, factors transferring some of the glacial epoch cold to an interglacial. To gain more insight into the mechanism of these processes, it will be necessary to study the degree of sensitivity of the global climate to changes in the concentration of greenhouse gases. This can be accomplished with the aid of data on climatic and atmospheric changes - data such as those obtained from the study of the Antarctic ice core.

In order to determine the contribution of individual component factors to the cumulative radiation impact on the terrestrial climate, a multivariate analysis of the Vostok ice core data was carried out with respect to these five factors: greenhouse gases, the concentration of dust and sulphates of non-marine origin, the volume of continental ice, and the extent of local insolation (Genthon et al., 1987).

According to this study the contribution of greenhouse gases to the overall effects on temperature has never been less than 40% or more than 65% within reliable dating, i.e., reaching as far back as 110 thousand years. The total effect of greenhouse gases and natural processes in the Northern Hemisphere approaches 80%.

Consequently, the contribution of greenhouse gases to temperature changes in central East Antarctica over the latest climatic cycle may lie within the 40 - 60% range, or 50 - 10 %. This means that about 3° C out of 6° C - the amplitude of glacial-interglacial temperature changes - may be attributable greenhouse gases.

3. Mass balance of Antarctica and sea level change

The contribution contains a summary compilation of information about the net mass balance of the Antarctic ice sheet. The approach is to compare mass input values with corresponding mass output values where both are known, and then to extrapolate to the rest of the ice sheet in several ways.

The mass input is snowfall on the surface (precipitation minus evaporation), which is measured in snow pits and ice cores either by determination of the annual layering or by finding the depths to horizons of known age (particularly radioactive fallout horizons from hydrogen bomb tests in the atmosphere). Ice flows in the direction of the maximum surface slope from the interior to the coast. Output is determined as volume flux across a line at or near the margin of the ice sheet by measuring the speed of movement across the line and the ice thickness along the line. The difference between the output flux and the input upstream is the net mass balance.

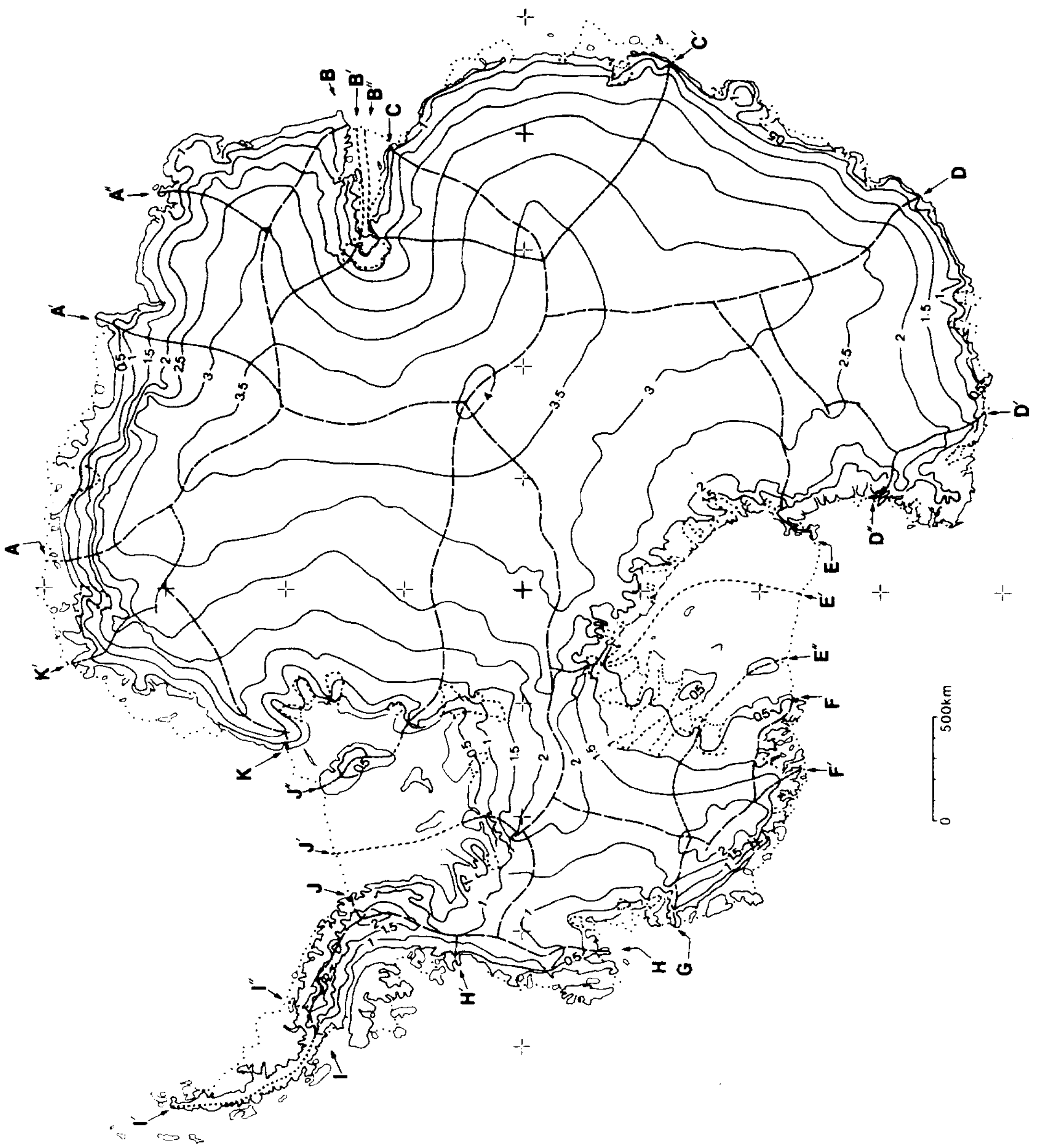


Fig. 6: Map of Antarctic surface elevation and drainage systems. Contour interval is 0.5 km. Drainage systems are referred to in the text by their coastal limits, marked by capital letters. From Giovinetto and Bentley (1985).

The surface elevation contours on the ice sheet (Fig. 6) are complex in form. Consequently, the normals to the contours define a series of drainage basins or systems - in places the ice flow converges strongly into a glacier or ice stream, in other regions the flow is divergent. Ice divides separate the different systems, of which there are about a score according to the division of Giovinetto and Bentley (1985).

It is important to recognize that different systems maybe behaving very differently. Some are strongly out of balance within the limits of our measurements. It is thus important to collect data from as many of the drainage systems as possible in the hope of getting an average that is representative of the ice sheet as a whole.

The surface elevations of systems that are out of steady state must be changing with time. Such changes are particularly noticeable near the coast where the ice sheet margin may be advancing or retreating, or the height of the ice against a protruding nunatak growing or dropping. However, because of the very long response times of the ice sheet to changes in input, times measured in thousands of years, such indicators of non-steady state reveal little about the net mass balance at the present time. Marginal changes maybe transient responses to events that took place far form the coast thousands of years ago; conversely, changes in input in the interior in recent centuries will not yet have had any effect on the outflow. Thus these indicators of non-steady state cannot be used as gauges of net mass imbalance in the sense in which we wish to know it today - is the Antarctic ice sheet right now contributing to sea level change?

This matter was recently studied by Bentley and Giovinetto (1991) using values for mass inputs updated from Giovinetto and Bentley (1985). The compilation by Giovinetto and Bentley (1985) yields the smallest total input of any published study and thus tends to minimize algebraically the overall net mass balance. However, difference between compilations are not great in most places. Outputs were compiled from published reports (Bentley and Giovinetto, 1991).

Balance assessments. Bentley and Giovinetto (1991) summarize the observations in three groups: systems with input and outflow on the (grounded) ice; systems entirely on an ice shelf and combined systems that include both inland ice and ice shelves but without measurements at the grounding line. The most desirable measurement in regard to sea level change would include output flux precisely across a grounding line. In actuality, output is usually measured either some distance inland of the grounding line or at or near the front of an ice shelf. In the latter case, some interpretation of the regimes of the ice shelf, particularly the rate of bottom melting, is necessary in order to estimate the net mass balance inland of the grounding line. The ice-shelf systems are, of course, irrelevant for the direct determination of sea level change since they are already in the ocean, but they were included because measurements on ice shelves are an aid to interpreting the third type of system wherein the output is known only at the front of an ice shelf. Results (Bentley and Giovinetto, 1991) are summarized in Tables 4-6 and Figures 7-9.

Current Antarctic net mass balance. There are many ways that one could proceed from the data presented to assess the state of the ice sheet as a whole. Bentley and Giovinetto (1991) chose three. First, they assumed that all the systems, both with and without measurements, are in balance except those, indicated in Tables 4-6 and Figures 7 - 9, that they considered to be significantly out of balance. The sum of the net balances for those four systems, Lambert Glacier, East Antarctica into the Ross Ice Shelf, Pine Island Glacier, and Brunt/Riiser-Larsen Ice Shelves, is 78 Gt/yr, which is equivalent to a sea level lowering of 0.22 mm/yr.

As a second approach, they took the sum of all the inputs and outputs for the systems that comprise inland ice only and then assumed that these regions are, on the average, typical of the ice sheet as a whole. Extrapolations were weighted in two ways; by mass input, because the regions without measurements are predominantly in the coastal zones of heavier snowfall, and by area.

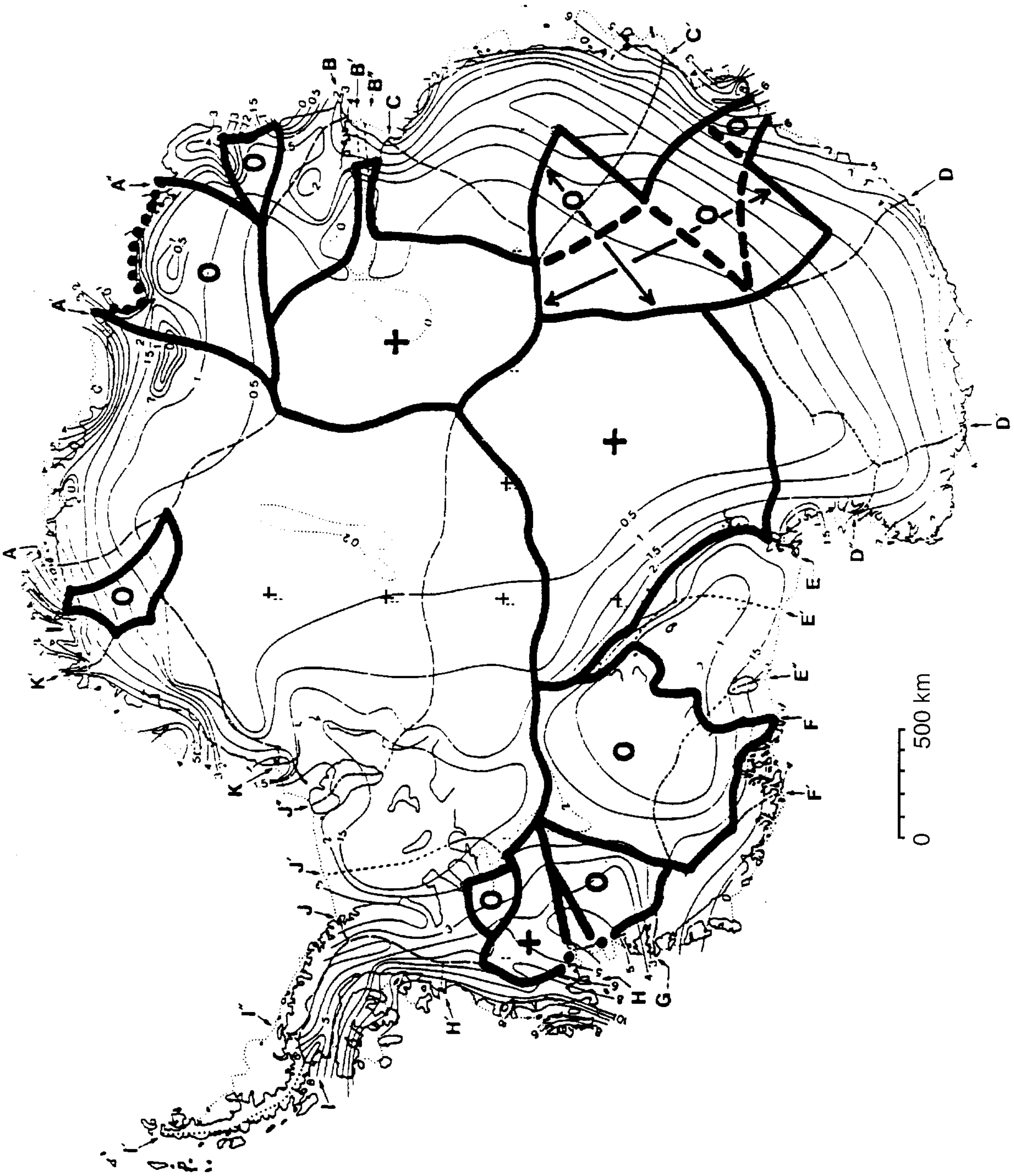


Fig. 7: Map of surface mass balance rates (from Giovinetto and Bentley, 1985), upon which the inland-ice systems containing net balance determination have been delineated. O means no significant imbalance and + means a significant positive net balance, according to Bentley and Giovinetto, 1991.

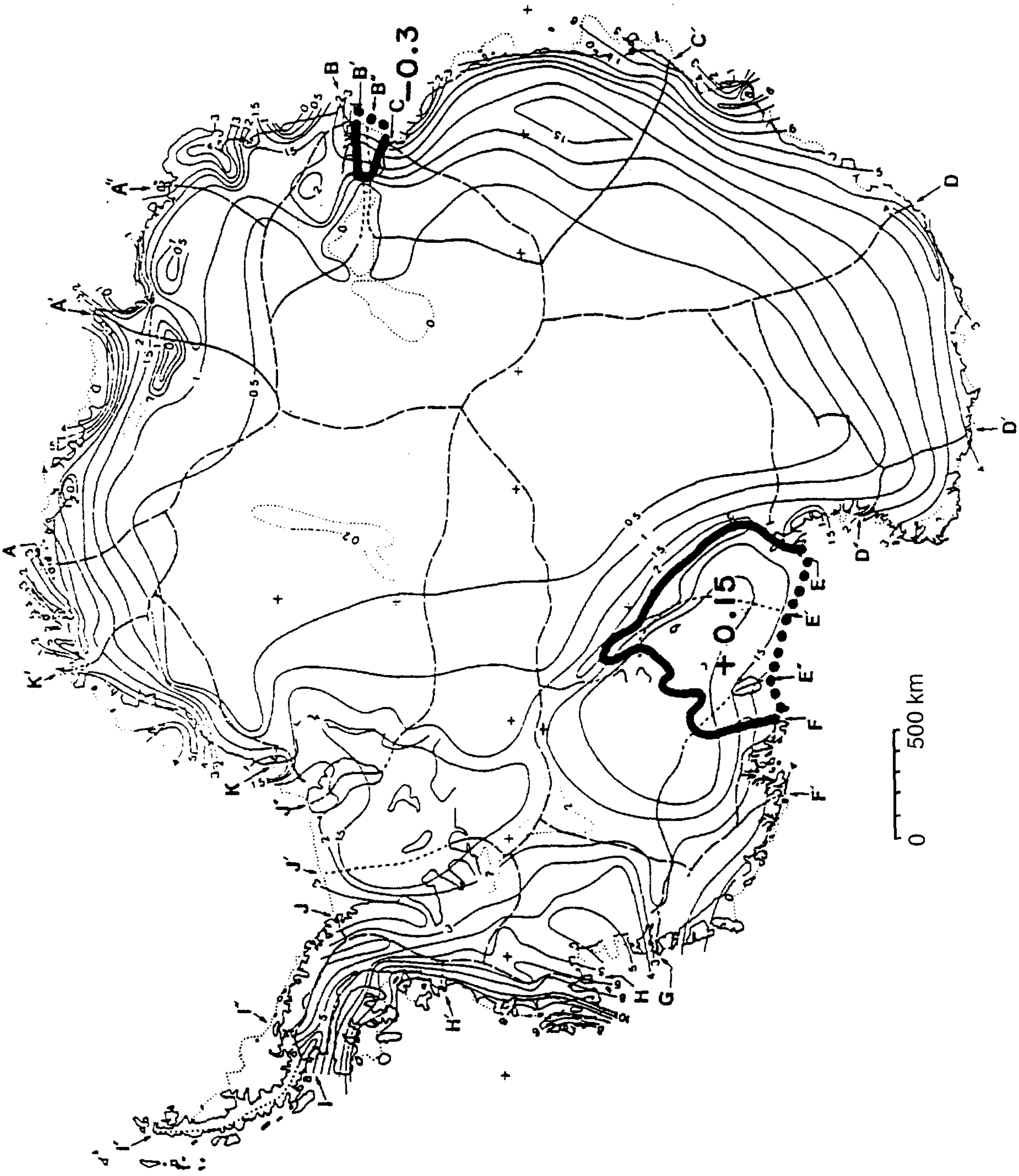


Fig. 8: Map of surface mass balance rates (from Giovinetto and Bentley, 1985) upon which the ice-shelf systems containing net balance determination have been delineated. Numbers are the bottom melt rates calculated on the assumption of steady state (see Table 5). From Bentley and Giovinetto (1991).

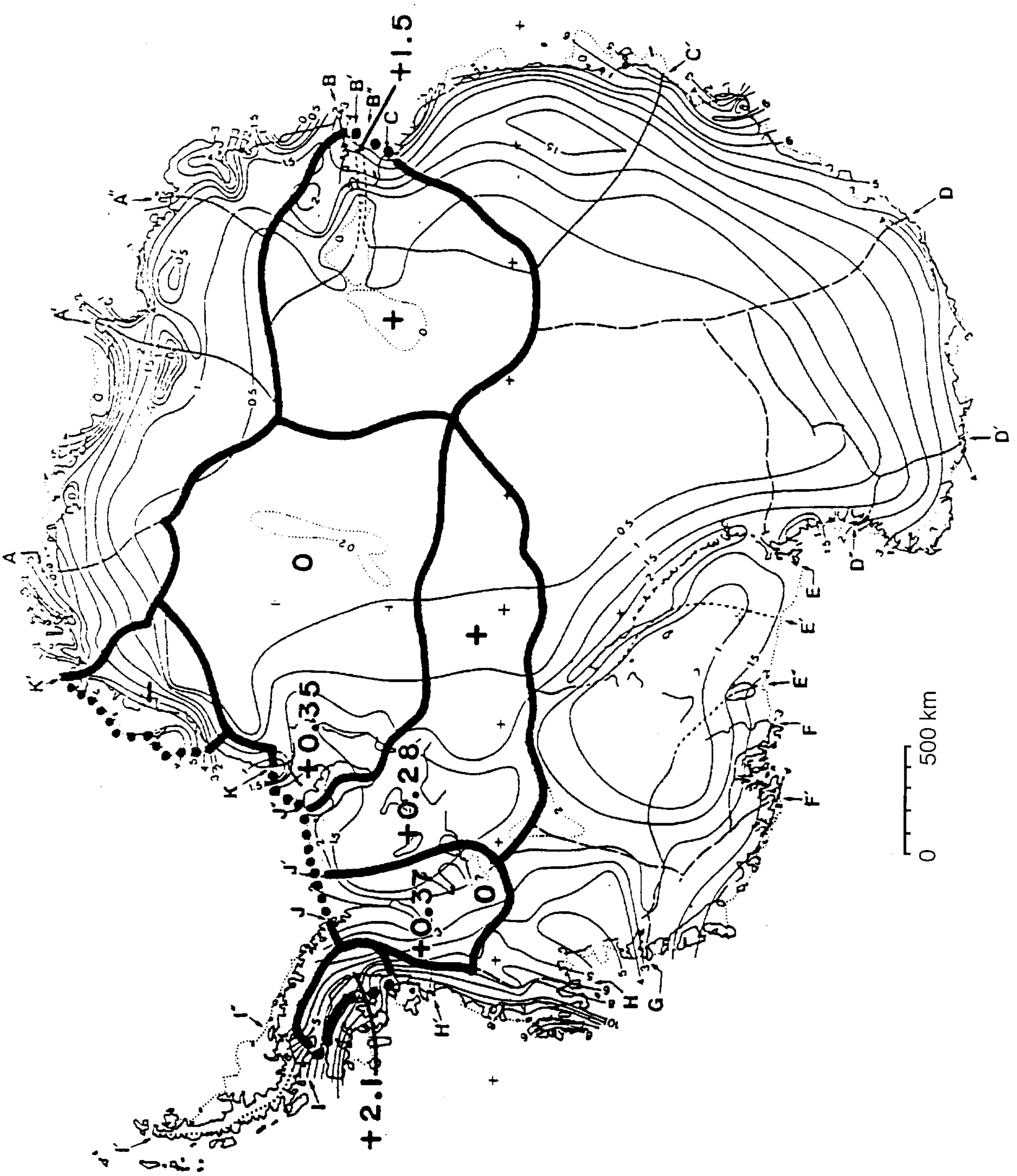


Fig. 9: Map of surface mass balance rates (from Giovinetto and Bentley, 1985) upon which the combined systems containing net balance determination have been delineated. 0 means no significant imbalance, + and - mean significant positive and negative net balance, respectively, according to Bentley and Giovinetto (1991). Numbers are bottom melt rates calculated on the assumption of steady state (see Table 6). From Bentley and Giovinetto (1991).

Table 4

Measured Mass Balances, Inland Ice Alone

System	Mass			Imbalance
	Accumulatio n Gt/yr	Outflow Gt/yr	Net Gt/yr	%
Jutulstraumen	16	11	+5	+31
Eastern Queen Maud Land	35	35	0	0
Eastern Underby Land	13	10	+3	+23
Lambert Glaciers	50	11	+39	+78
Western Wilkes Land:				
interior	27	21	+6	+22
flank	64	65	-1	-2
Totten Glacier	44	40	+4	+9
Combined	79	75	+4	+5
East Antarctica				
into Ross Ice Shelf	77	51	+26	+34
West Antarctica				
into Ross Ice Shelf	91	99	-8	-9
Thwaites Glacier	49	44	+5	+10
Pine Island Glacier	76	26	+50	+66
Rutford Ice Stream	12	18	-6	-50
Totals:	498	381	+118	+24 %

Table 5

System	Mass				Implied
	Inflow Gt/yr	Accumulation Gt/yr	Outflow Gt/yr	Net Gt/yr	Basal Melt m/yr
Amery I.S., Lambert flow band	11	4	20	-5	-0.3
Grid eastern Ross Ice Shelf*	51	41	55	+37	+0.15
Grid western Ross Ice Shelf*	99	34	97	+36	+0.15
TOTALS:	161	79	172	+68	

N. B. The grid eastern Ross Ice Shelf is fed from East Antarctica, the gridwestern Ross Ice Shelf from West Antarctica.

Table 6

System	Mass				Basal Melt	
	Accumu–lation Gt/yr	Basal melt Gt/yr	Outflow Gt/yr	Net Gt/yr	Needed for Steady State m/yr	Reason–able
Amery Ice Shelf	97/65	-11 (-0.7m/yr)	25	+83/+51	+1.5/+0.9	no
George VI Ice Shelf	52		4	+48	+2.1	yes
Western Ronne I. S.	91		47	+44	+0.37	yes
Eastern Ronne I. S.	103		44	+79	+0.28	no
Filchner I. S.	107		60	+47	+0.35	yes
Brunt-Riiser-Larsen I.S.	80	45 (1m/yr)	72	-37	+0.16	no
TOTALS:	530	34	252	+244		

Thirdly, they extrapolated from all systems that include inland ice. The two sets of extrapolated results are shown in Table 4. These three approaches together suggest an overall positive mass balance in the range 80 - 400 Gt/yr, i. e. a contribution to sea level lowering of 0.2 - 1.1 mm/yr.

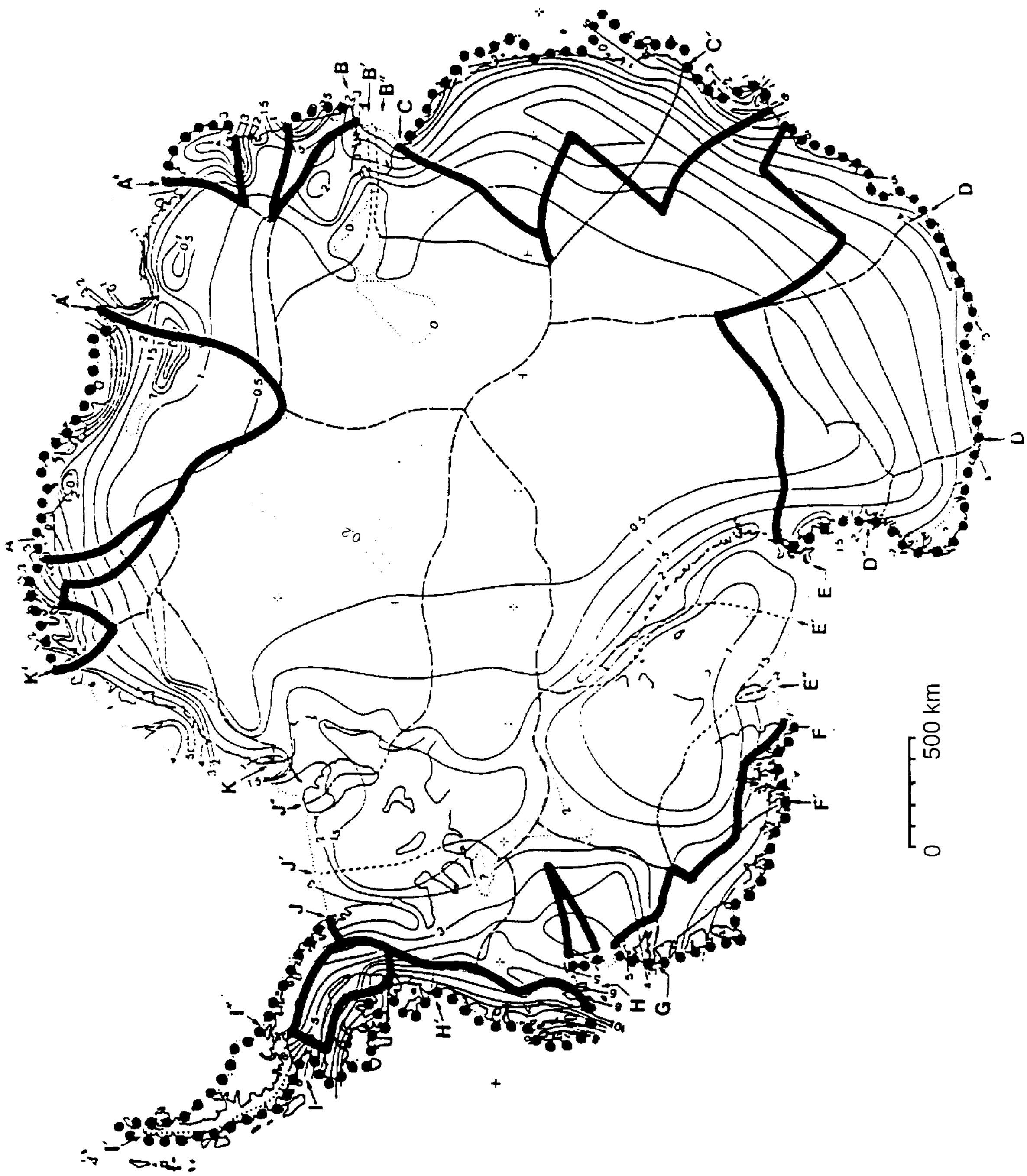


Fig. 10: Map of surface mass balance rates (from Giovinetto and Bentley, 1985) upon which the systems without net balance measurement have been delineated. Dotted lines denote the outflow perimeters.

A different type of approach they applied was to consider the implication for the regions without measurements (Fig. 10) of assuming either that Antarctic ice sheet is making no contribution to sea level change at all, or that it actually contributes to sea level rise. If the true overall net balance were zero, the negative net mass balance required for the unmeasured regions, into which the input would be 870 Gt/yr, is 80-400 Gt/yr, or11 - 5 % of the input. For the Antarctic to be contributing 0.5 mm/ yr to sea level rise, as has sometimes been proposed, an additional 180 Gt/ yr would have to be supplied, which would imply a negative net balance of 36 - 79 %.

Is there any reason to suspect that the regions without measurements are characterized by strongly negative net balances? In fact, what evidence there is suggests the contrary. For example, system AA' along the Queen Maud Land coast (see Fig. 6) has an input of 77 Gt/yr and a coastline that is1,150 km long. The mass flux for balance would be 0.067 Gt/km yr, essentially the same as for neighboring system K'A, which contains Jutulstraumen. But satellite images (Swithinbank, 1988) show no outflow systems in section AA' nearly comparable to Jutulstraumen. Extensive mountain ranges in the interior sub-parallel to the coast block ice flow through them - the flux through the entire Sor Rondane is less than 2 Gt/yr (Van Autenboer and Decleir, 1978). The problem in this sector seems to be how the outflow can even equal the input, let alone exceed it.

In Wilkes Land, early work by Lorius (1962) in a small sector along the Adelie Coast (section DD') (not included in our compilation because there were no measurements of ice thickness) suggested a positive net mass balance. The only system whose outflow actually has been measured at the coast, that through Totten Glacier, is in balance. Farther to the west, the small coastal sector in Enderby Land in balance or slightly positive.

In Graham and eastern Palmer Lands (section IJ) the evidence is mixed and there is no clear trend, as extrapolation of sparse data in this region of rugged terrain is difficult (C. S. M. Doake, personal communication, 1989).

Thus what evidence there is suggests that the coastal areas are no more likely to show large negative balance than the interior. We find nothing to contradict the conclusion of a net positive balance of a few hundred Gt/yr, i.e. a negative contribution to observed sea level rise of several tenths of a millimeter per year. Note that this conclusion has not changed from that of Budd an Smith (1985) who estimated 0 - 1 mm a sea level lowering based on fewer data and a different approach.

Iceberg flux. Data from a systematic program of iceberg observation in the waters around Antarctica, in which nearly all ships traveling to and from Antarctica participate, have been analyzed carefully by Orheim (1985, 1988, personal communication, 1989), taking into account multiple sightings, mean residence times, and size distributions. His conclusion is that the total calving rate from the continent, ignoring the recent giant icebergs, is at least as large as the total mass input onto the continent. If melting under the ice shelves is added as an output term, the implication is for a negative net mass balance.

Implicit in this conclusion is the assumption that there is no net break-back of the ice-shelf fronts. The indicated imbalance is equivalent to an average break-back rate of 1 km/yr along all the ice shelf fronts around the continent. There is at least qualitative evidence to indicate that such a break-back is occurring. Zakharov (1988) has compiled information from maps and satellite images on frontal positions of the ice shelves and floating glacier tongues around Antarctica over the last century. He found that through the 1970's a larger proportion of the fronts were advancing than at any other time in his coverage. Unfortunately, his data end in 1980, but it would be reasonable to suppose that after a period of unusual advance a net break-back would occur. In the Antarctic Peninsula region the Wordie (section HI), George VI (section HI), and Larsen Ice Shelves (section I'I) have all been decaying in recent years (even aside from the 11,000 sq km calving event from the Larsen Ice Shelves in 1986) (C.S.M Doake, personal communication, 1989; Doake, 1982; Doake and Vaughan, 1991). Shirase Glacier tongue has broken back in the 1980's (Nishio, 1991). Williams and Ferrigno (1988) and Ferrigno et al. (1991) document large-scale retreat or recession in several areas, although their study is dominated by the super-giant icebergs that are not included in the iceberg flux calculations. Overall, a general retreat of the floating ice margins of the ice sheet is indicated, but whether that retreat is large enough to account for the excess iceberg flux remains problematical.

Recent glacial changes

Although marginal changes in the ice sheet cannot be interpreted directly in terms of net mass balance, they are still of interest as indicators of change of some sort, so they are briefly summarized here.

In some places the ice sheet has been shrinking. At Gaussberg, a nunatak on the Queen Mary Coast, the ice level dropped 8 m in the 55 yrs between 1902 and 1957 (Dolgushin et al., 1962). The ice cap on Drygalski Island, which lies off the Queen Mary Coast, was lowering at 0.4 m/yr in 1957 - 58 (Shumsky, 1963). Brooks et al. (1983) interpreted satellite radar altimetry from Seasat in 1978 as revealing surface lowering of about 4 m on the Amery Ice Shelf, a 40-km retreat in the grounding line at the junction of the ice shelf with Lambert Glacier, and lowering up to 2.5 m/yr near that junction, all relative to an optical leveling survey in 1968 - 70 (Brooks et al., 1983). I. Allison (personal communication, 1991) doubts these results based on preliminary results of a recent remeasuring survey by the Australian National Antarctic Research Expedition. Shirase Glacier shows a surface lowering that varies irregularly from 0.3 m/yr 450 km inland from the coast to as much as 2.5 m/yr 150 km inland (Nishio et al., 1989).

In other places, however, changes are very slow or non-existent. The ice margin against the Schirmacher Hills on the Queen Maud Land coast was shown by the evidence of lichen and repeated photography to have moved by less than 2 or 3 meters in 20 years (Dolgushin et al., 1962). In the ice drainage basin around station Molodezhnaya the surface lowering is less than a few centimeters per year, and the retreat of the ice-sheet margin is no more than 1 km in a thousand years (Meier, 1979). And on LOW Dome direct measurements of elevation and gravity change show that over 20 years the lowering has been less than 0.2 m/yr, and is probably close to zero (Young et al., 1989).

Interestingly enough there have been no reports of marginal advances or near-marginal thickening of the ice sheet, such as one might expect for a positive mass balance. In light of the long response time of the ice sheet, this may simply reflect the fact that the positive mass balance is a thing of relatively recent times, whereas the marginal retreat or stability reflects conditions in the respective drainage systems centuries ago. For example, the extreme imbalance in the Amery/Lambert system combined with marginal retreat (if it is real) is suggestive of recovery from a glacial surge (Budd and McInnes, 1979).

Recent temperature and accumulation rate changes

There is increasing evidence from around the continent that both temperature and accumulation rates are increasing. In the Antarctic Peninsula there has been an increase of more than 25 % in accumulation rate and a corresponding1.8 K rise in temperature over the last 30 years (Peel and Mulvaney, 1988). If we associate these two phenomena together, they imply that there is a 15 % increase in accumulation rate per degree increase in temperature. At six widely scattered locations on the main part of the continent, Pourchet et al. (1983) found an increase in accumulation rate between the periods 1957 - 56 and 1965-c. 1980 of amounts varying from 10 % to 42 % with an average of 30 %. At these sites the temperature increase over the same time span was only 0.1 K, not nearly enough to cause the increase in accumulation rate. Four cores recovered from Law Dome and the slope of the ice sheet above Casey Station show an average accumulation rate increase of about 20 % above the long-term mean in the last 30 years (Morgan et al., 1991). Although these accumulation-rate trends appear to be consistently positive, one should note that tenor twenty percent variability is commonly observed (Giovinetto, 1964; Barkov and Petrov, 1966; Jouzel et al., 1983). But Jacka and Budd (1991) present data that do statistically support an increase in both temperature and accumulation rate compatible in magnitude to that expected from greenhouse warming.

Predictions

If mean air temperature increases in future years, it is virtually inevitable that snowfall will also increase over the Antarctic interior. One way to estimate the amount of the increase per degree of temperature change is to compare the accumulation rates at interior stations with the corresponding atmospheric temperatures. This was done by Robin (1977) who found that the variation was just about what one would expect from the basic dependence of saturation water-vapor pressure on temperature, which would imply a 10 % increase in accumulation rate per degree of temperature rise. Various other analyses lead to coefficients that range from 5 %, to 15 % per degree (Lorius et al., 1984; Muszynski and Birchfield, 1985; Giovinetto et al., 1990; S. G. Warren, personal communication, 1989; Budd and Simmonds, 1991, Antarctica Peninsula data cited above). We do not try to choose between these models, but will instead adopt a range of 5 - 15% per degree.

The actual temperature change to be expected is also uncertain. We simply assume an average increase of 2 degrees to 2100 AD. The result is a total predicted contribution to sea level change that amounts to a drop of 0.1 to 0.2 m by the end of the next century. It is unlikely that there can be any significant contribution to sea level rise from accelerated flow of the inland ice on that time scale (MacAyeal, 1989; Lingle et al., in press; Oerlemans, 1989), so no contribution from this process has been included in this estimate. On the longer time scale, however, the potential effect of accelerated flow far outweighs that of precipitation increases. Furthermore, ice-dynamic effects may be irreversible set in motion in the coming century even if their effects become serious only later. Thus it remains important to gain a better understanding of just what those effects will be, and to include consideration of them in planning for global change.

At the present time the Antarctic contribution to sea level change is a drop of 0.2 - 1.1 mm/yr, probably between 0.3 and 0.8 mm/yr (Table 7). That amounts to 100 - 300 Gt/yr total excess mass input. Excess iceberg flux probably reflects a secular break-back of ice-shelf fronts.

In the next century that contribution will probably increase on an average of 1 -2 mm/W, i.e., a total of 0.1 - 0.2 m by the year 2100.

Table 7

Contribution to Sea Level

	From Grounded Areas Alone		From All Measurements	
	By Mass	By Area	By Mass	By Area
Total mass input M	1660 Gt/a		1660 Gt/a	
Total area (A), km^2		$12.1 \cdot 10^6$		$12.1 \cdot 10^6$
Mass input to measured systems (m)	500 Gt/a		980 Gt/a	
Fraction of total mass input ($f_m = m/M$)	30 %		59 %	
Areas of measured systems (a), km^2		$5.1 \cdot 10^6$		$8.6 \cdot 10^6$
Fraction of total area ($f_a = a/A$)		42 %		71 %
Mass excess in measured systems (δ_m)	120 Gt/a	120 Gt/a	80 Gt/a	80 Gt/a
Mass excess extrapolated to whole ice sheet				
(δ_m/f_m)	400 Gt/a		140 Gt/a	
(δ_m/f_a)		290 Gt/a		110 Gt/a
Fraction of total input	24 %	17 %	8 %	7 %
Equivalent sea level lowering	1.1 mm /a	0.8 mm/a	0.4 mm/a	0.3 mm /a

4. The problem of surge-type glaciers

A glacier surge seems to be the best example of natural instability in a number of diverse snow and ice variations on the Earth. Surge-type glaciers exhibit the quasi-periodic oscillations between normal and fast motion. The unusual behavior during surge motions is one of the most interesting scientific problems in glacier dynamics. The surge problem is not only purely academic. Glacier surges may really create a problem, when advancing ice dams the side valleys, forming large lake basins, from which sudden outburst may result in downstream flooding. Surging glaciers may block or destroy roads and mountain settlements, or make alpine recreation routes inaccessible for long periods. Floods of very turbid water associated with the end of a surge may affect downstream culture and wildlife habitat (see Meier and Post, 1969; Dolgushin and Osipova, 1982; Raymond, 1987; Sharp, 1988 for reviews). Sometimes, the rapidly advancing lower part of a surging glacier could transform into a disintegrated water-ice mixture, as happened during the 1974 Didal surge in the Pamir Mountains and the 1902 Kolka surge in the Caucasus. At these occasions ice mixture flows of 2. 10^6(Didal), $10^8 m^3$ (Kolka) volume rushed down the valleys for 3 and 13 km with average velocity of 60-70 km/h, destroying everything and eroding the valley slopes up to 50-140 m above the valley bottom (Rototayev, Khodakov and Krenke, 1983).

Virtually all glaciers exhibit flow variations over time and generally the glacier length changes can be good indicators of climate. But in some cases they result from nonclimatic processes such as volcanic heating, tide-water influence, seismic events, shifts of ice divides, rockfalls on the surface, hydrothermal and hydraulic phenomena, and even human impact. All of them may produce glacier termini changes which are often larger than those produced by climate fluctuations. Thus glaciers in one region may react in an nonhomogeneous manner to a given climate change.

The same is true for the surge-type glaciers, but in contrast to above cases, they seem to oscillate in a self-sustained regime without any external forcing. Nevertheless, resemblance of rapid glacier advances, which can be triggered by different mechanisms, and lack of experimental data are the main cause of arguments, terminology confusion and misinterpretation.

Terminology problem

Surge-type glaciers are characterized by a multiyear, quasi-periodic oscillation between extended periods of normal motion and brief periods of comparatively fast motion (Meier and Post, 1969). Surge behavior is described by Raymond (1987) as follows. (1). Surges occur repeatedly. (2) in a glacier the quiescent (inactive) interval between surges is fairly constant (10 to 100 years). (3) The surge phase is relatively short (several years). (4) During the surge phase, ice speed is 10 or more times the speed during quiescence; accumulated ice displacement may be one tenth or more of the glacier length; ice is drained rapidly from an upper reservoir area to a lower receiving area; large elevation drops and rises (10 - 100 m) occur in reservoir and receiving areas respectively. (5) During the inactive phase, ice speed is low total displacement is smaller than during surge; ice surplus in the reservoir area is lost in the receiving area, which gradually returns the glacier to near presurge state.

Some purists prefer the term surge-type glacier to surging glacier because the latter might indicate that the glacier is just in the active phase. Those glaciers that respond slowly to climatic changes and other external factors, and respond with predictable perturbation of ice flow propagating down glacier are generally known as normal glaciers or stable glaciers. There are different opinions about whether there is a continuos spectrum of behavior from surging glaciers to normal ones, or whether surging glaciers are a class apart. in this connection the descriptive term pulsating glacier is suggested by L. Mayo (1978) for those glaciers which show evidence of periodical instability but to a lesser degree than surging glaciers. However, for many years in Russian terminology the term “pulsating glacier” has been used in the meaning of surge-type glacier and the author of the term emphasized the self-sustained nature of oscillations (Kazanskiy, 1965), while the English term “pulsating glacier” into Russian as impelling glacier. The category of dynamically unstable glaciers including surge-type glaciers was proposed in an attempt to cover all anomalous phenomena irrespective to mechanisms, origin of oscillations and degree of manifestation (Rototayev, 1978a). Sometimes, the glaciers exhibiting rapid fluctuations of any origin are referred to as flooding glaciers (Kazanskiy, 1978).

Shumskiy (1978) described four physically different types of glacier oscillations: (1) the forced oscillation due to varying external load; (2) two types of parametric oscillations: (2a) the high-frequency parametric oscillation of the sliding velocity caused by the fluctuations of the effective roughness, which are in turn affected by variations in the surface melting rate, the subglacial runoff and the hydraulic head of subglacial water, (2b) the low-frequency parametric oscillations of the ice flow velocity caused by temperature waves into the ice; and finally, (3) the self excited oscillations of the sliding velocity due to oscillations of dry friction (debris matter against the bed) depending on debris concentration in ice, which is controlled by the movement itself. If the external conditions remain constant long enough a glacier would either become stationary or undergo self-sustained oscillations.

It seems obvious that the answer to the terminological problems could be found in the data of comprehensive field studies.

Observations

First reports of disastrous glacier surges were documented in the seventeenth and eighteenth centuries. Surges of the Vernagtferner in the Ötztal Alps have been known since 1599 (Hoinkes, 1969). Glacier surges in Iceland were determined from the beginning of eighteenth century (Thorarinsson, 1969). Information on surges of the Devdorak Glacier in the Caucasus dated back to 1776 (Statkovskiy, 1879). The surging Devdorak Glacier, as well as Chach and Abano Glaciers in the same Kazbek massif, created well known “Kazbek Blockages” that were a serious problem in the beginning of nineteenth century when Russian engineers started to build the Georgian military road. Systematic observations of the surges of Devdorak Glacier were conducted from 1862 to 1887 (Khatasyan, 1888). R. Tarr and L. Martin (1914) described the surges of some glaciers in the Yakutat area, Alaska, as “glacier floods”.

Nevertheless, the surges were previously believed to be rare events. The situation has changed in the last three decades when a large number of surge-type glaciers were studied and described in Alaska, Canada, Greenland, Spitsbergen, Island, the Alps, the Caucasus, Central Asia, Kamchatka, South America, and New Zealand (Dolgushin and Osipova, 1982). In Svalbard, for example, 90 % of the glaciers are surging (Lefauconnier and Hagen, 1991). But the most extensive field works covering nearly complete surging cycles were only on Medvezhiy Glacier, Pamirs (Osipova, Tsvetkov, 1991) and Variegated Glacier, St. Elias Mountains (Kamb et al., 1985).

Measurements of surface elevation and velocity over 25 years from 1963 on Medvezhiy Glacier are now providing data on glacier surface kinematics during three surge phases (1963, 1973 and 1988- 1989) and two quiescent phases (Dolgushin and Osipova, 1975; Buynitskiy and Sorotkin, 1986; Dolgushin and Osipova, 1978, 1982; Osipova et al., Tsvetkov and Sorotkin, 1982; Osipova et al., 1991).

Studies on Variegated Glacier began in 1973, in the middle of an inactive phase, and have continued through the 1982 -1983 surge event. The y include geometry and velocity investiga - tions, water pressure measurements in the bore holes, dyetracing and water flow discharge measurements, seismic experiments (Bindschadler et al. 1977; Harrison et al., 1986; Humphrey et al., 1986, Kamb, 1987; Richards, 1988).

Comparison of the Medvezhiy and Variegated surges

There is information of different degree on the Medvezhiy surges in following years: 1913 -1915, 1937, 1947-1949, 1951, 1963, 1973, 1988 -1989. There were no observations in the period from 1913-1915 to 1937 and possibly, one surge was missed. The recurrence period is thus about 10 to 15 years.

Surges of Variegated Glacier are known to have occurred in 1906, the late 1920s or early 1930s, about 1947, 1964-65 and 1982-1983. The recurrence period is thus about 17 to 20 years. The following description is based on the observations during the last surge cycles of both glaciers.

Quiescent phase

The build-up stages of both glaciers are characterized by the filling of a reservoir area and depletion in a receiving area. The reservoir area of Medvezhiy Glacier is in the upper part of the tongue. It fills by the ice flowing from a firn basin through an icefall, because the ice inflow exceeds the ablation on the tongue. The reservoir area of Variegated Glacier coincides with an accumulation area and builds up due to excess of accumulation over ablation and ice outflow. The boundary between thickening and thinning has been referred as the dynamic balance line (DBL) or the active front (Dolgushin and Osipova, 1978). The DBL on Medvezhiy Glacier, which appeared as a bulge on the surface, moved down the glacier at a speed of 0.2 - 1.2 km a^{-1} while on Variegated Glacier it was morphologically less pronounced and moved down at 0.3 km a^{-1} (Osipova and Tsvetkov, 1991).

Strong seasonal variation of velocity are quite common for both glaciers. On Medvezhiy Glacier the longitudinal velocity abruptly increased every year at the beginning of the warm season in the whole active part of the glacier. The increase of velocity was accompanied by a rise of the surface in the lower part of the activation zone and by an advance of the DBL. In 1967 and 1968, as well in 1978 and 1982 the seasonal velocity increase and the DBL advances were considerable. Surface and elevation changes were similar to those during the surge phase, but the surge died out not reaching the glacier's terminus. Such event were described as a "microadvance" by Dolgushin and Osipova (1982). On Variegated Glacier summer velocity was over twice as high as the winter velocity. Faster motion in summer than winter indicated a seasonal sliding contribution (Raymond, 1987). The same is true for the surge-type glaciers of the Susitna River Basin in Central Alaska, where the early melt season velocity peak is two times higher than late melt season minimum (Clarke, 1991a).

Minisurges anomalous events were observed on Variegated Glacier during a quiescent phase in the reservoir area in the early melt season in a sequence of four to six, spaced at several days to two weeks. A minisurge is an abrupt increase in speed over a few hours from about 55 cm/d to a peak of 100 -300 cm/d followed by a slower decay, over about one day, to near background speed. The flow-velocity peak propagates down-glacier at a speed of about 0.3 km/h. It is accompanied by a peak of high micro-seismic activity, as well as rapid uplift of the glacier surface by 6 - 11 cm and propagating pressure wave in the basal water system of the glacier, in which, after a preliminary drop, the pressure rises rapidly to a level greater than the ice-overburden pressure at the glacier bed and then drops over a period of 1 - 2 days. Minisurges seems to be due to enhanced basal sliding caused by high basal water pressure and the consequent reduction of bed friction. Similar short-lived events of rapid motion occur on the surge-type glacier of the Susitna River Basin in Central Alaska, as well an on Medvezhiy Glacier (Clarke, 1991a; Dolgushin and Osipova, 1982). Whether or not these phenomena have the same origin cannot be decided because of the limited evidence.

The closest relation, it appears, is between the minisurges and the peaks of main surge velocity. Both of these phenomena can be accounted for by the propagation of a water pressure wave. Both minisurge and main generate abnormally high turbidity of the outflow-stream water (Kam and Engelhardt, 1987). Although the fast motion of rninisurges and surge motion may have similar mechanisms, and minisurges may be premonitory to a surge, similar short-therm movement pulses occur on Alpine glaciers that are not known to surge (Raymond, 1987). Probably, the initial build-up of basal water pressure in the surge took place by a different mechanism, because it occurred in mid-winter, when there was no surface melting water (Kamb and Engelhardt, 1987).

Surge phase.

The Medvezhiy Glacier surge phase in 1988-1989 can be divided into three stages. The first stage began before June 21, 1988, but not earlier than August 1987, when the ice velocity was less than 1.5 m day^{-1} at 4.5 km from the ice fall. During the first stage considerable increase of ice velocity was accompanied by surface lowering and formation of crevasses in the upper part of the glacier tongue, while the morphological terminus of the glacier was stable. During the second stage (from autumn 1988 to the middle of May 1989) the glacier terminus slowly advanced, in the upper glacier the ice thinned and in the lower glacier thickening occurred. Up to the middle of April the glacier velocity increased gradually from 1 to 3 m day^{-1}, to the end of May the velocity increased to 15 m day^{-1}. The third stage (from the middle of May to the beginning of July 1989) is a period of rapid advance of the glacier terminus and the surge termination. On the 20 June the velocity peaked at 50 m day^{-1}. After that the velocity decreased sharply to practically zero. The total advance of the glacier tongue was 1100 m, the maximum uplift in the lower part was 180 m, and the upper part of the tongue dropped by 50 m. In 1988 the situation on Medvezhiy Glacier differed from the previous presurge periods: (1) the ice volume in the reservoir area was $40 \cdot 10^6$ m³ lower than before the 1963 and 1973 surges, and (2) all ice from the previous surge was melted away, while before the 1963 and 1973 surges there were $20 \cdot 10^6$ m³ of dead ice below the active part of the glacier, which later on was involved in the surges (Osipova and Tsvetkov, 1991).

The 1982-1983 surge of Variegated Glacier went on in two distinct stages. The first stage began in January 1982. The upper glacier was moving at about 2 m day^{-1} and from May to late June the velocity rose gradually from 2.6 to 9.2 m day^{-1} (maximum 10.4 m day^{-1} on 26. June). By mid-August the velocity had declined to about 1 m day^{-1} and remained low through August to September. The surging seemed to have stopped. During the first stage the flow velocity was high on the glacier part from 4 to 7 km but decreased both up and downstream from this reach. The second stage (October 1982-4 July 1983) began with a gradual increase in micro-seismic activity and flow velocity in the upper glacier. In the beginning of January 1983 the velocity reached the level of 5-7 m day^{-1}, in April it began a further increase and reached a maximum of 15 m day^{-1} in mid-June. In the afternoon and evening of 4 July the surge abruptly terminated. By 26 July the velocity in the upper glacier was less than 0.2 m day^{-1} (Kamb et al., 1985). As a result of the surge the glacier surface was strongly broken by crevasses; in the upper glacier the ice thinned by as much as 50 m, below 8 km thickening occurred up to a maximum of 100 m. The total advance of the surge front was some 6.5 km, but it did not reach the glacier snout. The forward propagation of the surge front from 17 May to the surge termination occurred at a nearly constant speed of 80 m day^{-1} as a consequence of conservation of ice volume.

The problem of surge mechanisms

Surging in cold glaciers has been explained by creep instability warming the basal ice to the melting point (Robin, 1955). The model results show that due to the viscous dissipation of energy inside a cold glacier, a steady state may exist within certain limits of mechanical and thermal properties of ice and bedrock shape. Beyond these values rapid warming of ice occurs. The bedrock temperature reaches the pressure melting point, the friction on bedrock abruptly decreases, and the cold glacier surges (Bozhinskiy and Grigoryan, 1978). This mechanism may be responsible for the advance of Milne Glacier (Northern Ellesmere Island) at a mean annual rate of 250 m/a. As Milne Glacier is a cold glacier, surges may possibly be thermally regulated (Jeffries, 1984).

Another possible mechanism of glacier self-oscillations is that the dry friction forces drop after they are exceeded by critical shear due to upward movements of debris from the bottom along shear planes (Shumskiy, 1974).

Following observations of the 1982-83 surge of Variegated Glacier form the basis of Kamb's surge model (Kamb, 1987): (1) the fast flow motion during, the surge is due to rapid basal sliding (2) during surge, the pressure of water in the basal conduit is high, within 2 - 5 bars of the ice overburden pressure, and occasionally reaching overburden; in the nonsurging state it is generally 4- 16 bars below overburden. Peaks in pressure correspond to peaks in sliding motion, indicating that the high sliding speed in surge is directly caused by high basal water pressure; (3) major slowdowns in surge motion and particularly surge termination are accompanied by large flood peaks in the terminus outflow streams and by a drop of the glacier surface by 0.1 - 0.17 m. The high sliding speeds and high basal water pressure seems to be coupled with extensive basal cavitation; (4) dye-tracing experiments show that the water flow through the basal water conduit system is much slower (0.025m/s) in surge than in the nonsurging state (0.7m/s); (5)during the surge the dye appeared in all outflow streams, whereas after surge the dye appeared in one stream only; (6) the outflow during the surge was much more turbid (ea. 100kg/m^3 for particles sizes ≤ 10μm) than after surge (ca.1-10kg/m^3).

What is the key element of the surge mechanism that enabled the high basal water pressure to be built up and maintained? The pressure in the basal conduit system is a result of a balance between the buildup of pressure from water input and drawdown of pressure by outflow. Since increased pressure tended to increase the cavitation and hydraulic conductivity of the basal water system, it seems at first inexplicable why the higher basal water pressure in surge is accompanied by increased retention of water in the glacier, as necessary for the transition from the normal tunnel configuration of the basal water conduit system to a linked cavity configuration that tends to restrict the flow of water, resulting in increased water pressures that cause rapid basal sliding. The linked cavity system consists of basal cavities formed by ice-bedrock separation, ca. 1 m high and ca. 10 m in horizontal dimensions, and hydraulically linked by orifices, where separation is small (s 0.1 m). Orifice shapes and the amount of roof melting are determined by Ξ

$$\Xi = \frac{2^{1/3}(\alpha\Lambda/\omega)^{3/2}}{\pi^{1/2}\;DM}\left(\frac{\eta}{\nu\sigma}\right)^{1/2} h^{7/6},$$

were Ξ is the orifice melting-stability parameter, (α is the surface slope of the glacier; ω is an average tortuosity of the cavity system; A is the head gradient) $\alpha\Lambda/\omega$ is the local hydraulic gradient in the orifice, M is the Manning roughness, D is a constant with dimensions of length ($D = \rho_{ice}H / P_{water}\, g$; D = 31 km), η is the ice viscosity, v is the ice sliding velocity, σ is the excess of ice overburden pressure over basal water pressure, (h) is the step height in cavity (Kamb, 1987), The melting-stability parameter Ξ provides a measure of the importance of roof melting by dissipation in the linked cavity system. When $\Xi \geq 1$, so that the system is “viscous-heating-dominated”, the orifices are unstable against rapid growth in response to modest increase in water pressure or in orifices size over their steady state values. When $\Xi \geq 1$, the orifices are stable against perturbations of modest to even large size. Glaciers for which $\Xi \geq 1$ can go into surge, while those for which $\Xi \geq 1$ cannot.

The simplest mechanism of surge initiation is that the basal water conduit system first becomes a linked cavity system without tunnels, which happens in winter, when the water flux is low, and then, when the flux increases in spring, happens to be low enough that the linked cavity system does not degenerate to a tunnel system. The linked cavity system can be stabilized in the reservoir area by the increase in basal shear stress there as the ice thickness builds up prior to surge. However, a surge speed of 2 m/d in midwinter still needs further explanations.

Kamb’s model builds on the premise that the subglacial bed is rigid and impermeable. But seismic experiments on Variegated Glacier allow to detect a phase reversion of compressional wave reflections from the deepest part of the glacier during the 1982 surge event. The result might be interpreted in terms of a very weak, highly attenuating, fluidized sediment of debris layer, presumably in addition to basal cavitation, resulting from high water pressure during surge. Seismic phase reversal due to water alone at the glacier bed is unlikely (Richards, 1988). Thus another theory was developed by Clarke, that emphasizes flow in a porous substrate undergoing deformation. The hydraulic transmissivity of soft bed depends on the balance of consolidation and shear-induced dilatation or piping. Five variables, such as Darcian water transport, fine-grained sediment transport, consolidation, shear deforma - tion, dilatancy and comminution, uniquely determine the compressibility, permeability, shear strength, and other physical properties of subglacial till (Clarke, 1987). They may interact in a complex way that could lead to surge initiation.

The problem of identification of surging glaciers

There is a group of indirect signs which might be used to distinguish surging glaciers from others. Most of these indicators can be determined from aerophoto survey data and maps, and some of them from space images. Other parameters can be obtained only in field studies. The most reliable results are based on the combination of different indicators. There are specific evidences indicating the surging behavior with different degree of reliability. To clas - sify a glacier as a surge-type, one highly reliable indicator (shown in the following list by ✸) is necessary, or at least three less reliable indicators (shown in the list by ●) (instructions .... 1982). The brief list of the indicators looks as follows:

Variations of glacier outline.

• (a) rapid and relatively short-term advance of the glacier tongue;
• (b) slow and long-term advance of the glacier tongue;
• (c) spreading of the glacier tongue in a “paw-like” shape;
• (d) short-term variation of glacier width;
• (e) sharp glacier retreat;
• (f) overflow and collapsing of ice over the lateral moraines;
• (g) collapse of some part of the glacier;
• (h) sudden calving in a lake;
• (i) breaking away and transportation of tributary by the main glacier.

Variations of glacier surface elevation.

• (a) considerable drop of the glacier surface;
• (b) considerable rise of the glacier surface;
• (c) simultaneous thickening up-glacier and thinning down-glacier;
• (d) formation of precipices in the mouths of tributary glaciers.

Variations of ice speed.

• (a) variation of ice speed 10 - 100 times and more;
• (b) sharp advance of the glacier part inside its outline;
• (c) damming of tributary glaciers by swollen main glacier;
•(d) the large-scale, irregular intrusion of tributaries;
•(e) irregular ice feed from the tributary to the main glacier;
•(f) essential advancing (or retreating) of dynamic balance line on degrading passive tongue;
• (g)passing of swelling ridges down the glacier.

Variations of glacier surface morphology.

* (a) transformation from smooth to rough or block like appearance of the glacier surface;
* (b) relatively rapid (1 -2 years) transformation of a light block-like glacier surface to the dark one because of closing of crevasses and regeneration of debris cover;
• (c) very uneven glacier surface, quickly smoothed down by ablation;
• (d) surface with depressions resembling the rupture pattern;
• (e) steep glacier front;
• (f) rapid degradation of large areas of glacier;
• (g) formation of erosion gaps in passive ice;
• (h) rampart arc swells occur on tongue surface,
• (i) large irregularities in configuration of superimposed and imbedded tributaries in the main glacier.

Glaciotectonics variations.

•(a) appearance and fast development of crevasses;
* (b) active rupture deformations in the glacier body, formation of prismatic blocks and ice pulp;
* (c) for–mation of large longitudinal ice ruptures along the glacier margins;
•(d) formation of marginal crushed zones;
•(e) appearance of large thrusts in terminus area; (f) large transverse ruptures and faults up glacier; (g) large collapsing systems in passive tongue, and (h) push deformation of loose glacier deposits.

Glociohydrologicd variations.

* (a) fast simultaneous disappearance of surface water flows;
•(b) delay, blocking of glacier river runoff;
•(c) outburst floods from beneath the glacier;
•(d) water-ice and mixed mud flows;
•(e) formation of temporary dammed lakes;
•(f) outbursts of glacier dams.

Variations of moraines and their pattern.

•(a) loop-like pattern of medial and lateral moraines;
•(b) fresh erosional gaps through the lateral moraine ridges;
* (c) rejuvenated surfaces of old lateral moraines;
•(d) formation of push moraines;
* (e) fresh lateral moraines on high levels;
* (f) dispersal and collapsing of debris cover in crevasses;
* (g) fast melting out and closing up of debris cover on a large area of the glacier;
• (h) appearance of transverse moraine arcs on the glacier;
•(i) debris saturation of the glacier tongue.

Variations in periglacial zone.

•(a) formation in short time of large stagnant and buried ice massifs;
* (b) cut avalanche cones and suspended ice remnants on valley slopes;
* (c) fresh moraine flutes on the valley slopes above the glacier;
•(d) fresh glacier trimlines on high levels;
•(e) dammed lake shore lines and lake terraces;
•(f) relatively fresh evidence of glacial mudflows and outbursts;
•(g) disturbance of normal vegetation age sequence on slopes, moraine and glacier forefield;
•(h) formation of fresh step-like series of settling terraces on the glacier valley slopes.

Besides the specific evidences indicating the surging behavior at a certain degree of reliability, there are general morphological parameters which might be used for the surging glacier recognition:

Geometry and mass balance. In Central Asia the ratio of accumulation area to glacier tongue, width, the average slope of glacier and the ratio of accumulation area to total glacier area seems to be the best parameters to distinguish the surging from non-surging glaciers. The first parameter is considered as a criterion of damming in the accumulation area. In total 12 morphometric parameters have been applied to 64 glaciers with known behavior (43 surging and 19 non-surging) (Glazyrin, 1978).

Statistical analysis of 1754 normal and surge-type glaciers of the Yukon Territory (Clarke, 1991b) shows that the two glacier types have significantly different average geometries. Surge-type glaciers tend to be longer, wider and to have lower overall slope than normal glaciers. Correlation between length and surge tendency is the fundamental one. Apparent direct correlation between surge tendency and width and the inverse correlation between surge tendency and slope are entirely a result of strong length-width-slope correlations. The observed correlations lend no support to the predictions of the Kamb and Fowler theories, that small slopes (Kamb, 1987) or the small value of product of slope into width squared (Fowler, 1989) favor surging.

In case of compound glaciers consisting of many surge-type and normal tributaries, the observation, interpretation and forecast of main trunk surging behavior becomes very complicated even under quasistable climatic conditions. A random juxtaposition of many different surge periods of tributary glaciers may result in apparently irregular behavior of trunk glacier (Rototayev, 1978b). Clarke and others (1986) statistically showed that tributary glaciers have a higher probability of being surge-type than trunk glaciers. The likely explanation is that a surge occurring in a trunk glacier can induce surges of its tributary glaciers whereas the reverse situation is unusual. Meanwhile, observations on East Svalbard show that 55 surging independent ice-streams are represented by 42 main ice-streams and 13 tributaries (Lefauconnier and Hagen 1991). From the other hand, many of the modern large main trunks, which seem too large for the present climate. (Fedchenko and Garmo Glaciers, Pamirs) presumably have been formed and are kept stable by the surges of their tributaries (Rototayev), 1978b).

A difference between balance flux and actual down-glacier transport, which is taken as an indicator of surge behavior, shows two tributaries of Susitna Glacier in Central Alaska to be surge-type and one tributary to be non-surging. The main trunk of Susitna Glacier and its two unstable tributaries surge simultaneously with a period estimated to be 50-60 years. Susitna Glacier last surged in 1951-1952. The next surge is expected in the first decade of the next century (Clarke, 1991). Nevertheless, of the four parameters form, balance flux, flux rate and slope of 150 Alaskan glaciers, the hypsometric form is the best discriminator to distinguish the surging distribution, while three other parameters are poor discriminators (Wilbur, 1986), which is contrary to the proposal of using the balance flux and flux rate as an indicator of surge tendency. Nevertheless, it seems very interesting to find the relation of morphometric and mass balance parameters with surge behavior in different mountain systems for indicating and forecasting surge-type glaciers.

5. Fluctuations of mountain glaciers

Due to the retarding effect of latent heat exchange and the slowness of heat diffusion, secular climatic trends are clearly reflected in mass and temperature changes of 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). The following paragraphs concentrate on glaciers; they attempt to mention briefly the available information, outline a few ideas for their interpretation and suggest concepts of monitoring strategies for the coming century in view of possible anthropogenic warming.

The worldwide collection of information about ongoing glacier changes was initiated in 1894 with the foundation of the International Glacier Commission at the 6th International Geological Congress in Zürich, Switzerland. It was hoped that long-term glacier observations would give insight into processes of climatic change such as the formation of ice ages. Since then, the goals of international glacier monitoring have evolved and multiplied. Today, the World Glacier Monitoring Service (WGMS) collects standardized observations on changes in mass, volume, area, and length of glaciers with time (glacier fluctuations), as well as statistical information on the distribution of perennial ice on the Earth's surface (glacier inventories). Such glacier fluctuation and inventory data form a basis for hydrological modelling with respect to possible effects of atmospheric warming, and provide fundamental information in glaciology, glacial geomorphology, and quaternary geology. The highest information density is found for the Alps and Scandinavia, where long and uninterrupted records are available. The tasks of the WGMS are

(1) to continue collecting and publishing standardized data on glacier fluctuations at 5-year intervals.
(2) to complete and continuous y upgrade an inventory of the world's glaciers,
(3) to publish results of mass balance measurements from selected reference glaciers at 2-year intervals.
(4) to include satellite observations of remote glaciers in order to reach global coverage. and
(5) periodically to assess ongoing changes

This work is being carried out at the Laboratory of Hydraulics, Hydrology and Glaciology (VAW) of ETH Zurich under the auspices of the International Commission on Snow and Ice (ICSI/IAHS), the Global Environment Monitoring System (GEMS/UNEP), the Federation of Astronomical and Geophysical Data Analysis Services (FAGS/ICSU) and the Division of Water Sciences of UNESCO. Data from WGMS flow into the World Data Center (WDC-A) for Glaciology (Boulder/ Colorado) and the Global Resources Information Database (GRID of GEMS/ Geneva).

Observation Programme

With the internationally coordinated collection of information about ongoing glacier changes initiated in 1894 it was hoped that long-term observations of glacier fluctuation would show whether modern variations of glaciers and climate are globally synchronous or regionally variable, and how the dramatic processes of the geologically most recent past, the Ice Age, could be understood. Until the middle of the 20th century, information mainly concerned changes in length of glaciers (Fig. 11). Especially detailed data became available from the Alps, Scandinavia and Iceland. Some of the observations were started or carried out in connection with glacier hazards such as ice avalanches and outbursts of ice-dammed lakes. Signs of shrinkage and glacier retreat clearly dominated, with the exception of a short but marked readvanced of glaciers taking place in the Alps around 1920.

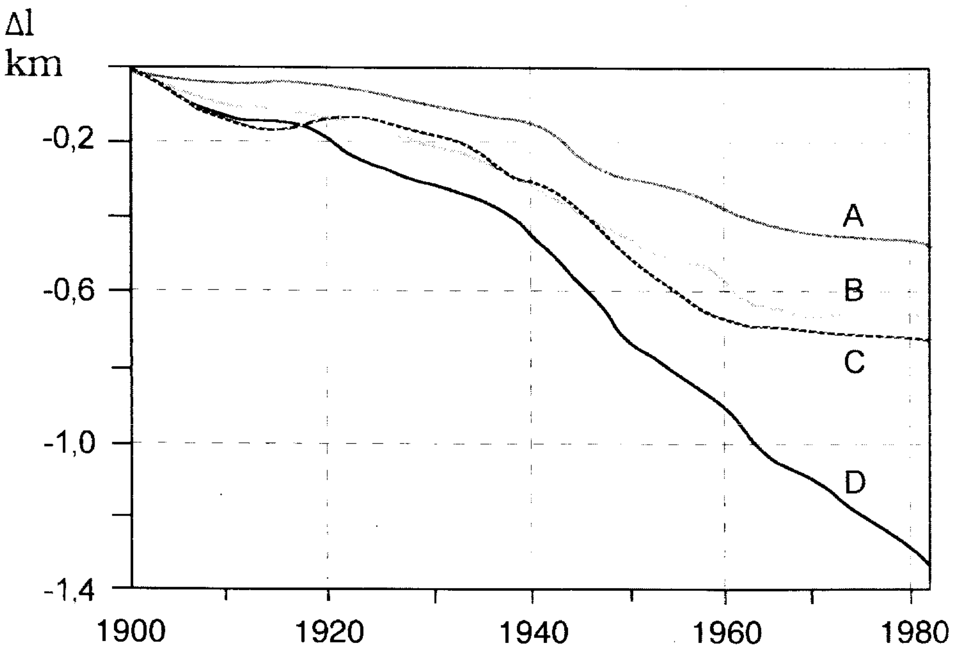


Fig. 11: *Length changes of Alpine glaciers since the beginning of the 20th century: A Pizol, Sardona, Punteglias, Basodino; B - Grand Desert, Lavaz, Kehlen, Valsorey; C - Trient, Saleina, Morteratsch, Zinal; D - Rhone, Gorner, Aletsch. Three-year running means are averaged from the indicated glaciers in each of the following size categories: (A) small cirque and mountain glaciers shorter than 2 km, (B) large cirque and mountain glaciers 2 to 5 km long, (C) short valley glaciers 5 to 10 km long, and (D) long valley glaciers more than 10 km long. The four groups reflect balance with increasing signal strength, time delay and smoothing: High-frequency (annual) signals show up in category A, decadal trends are clearly visible in B and C, whereas D mainly gives secular trends. Data basis: Kasser et al. (1986).*

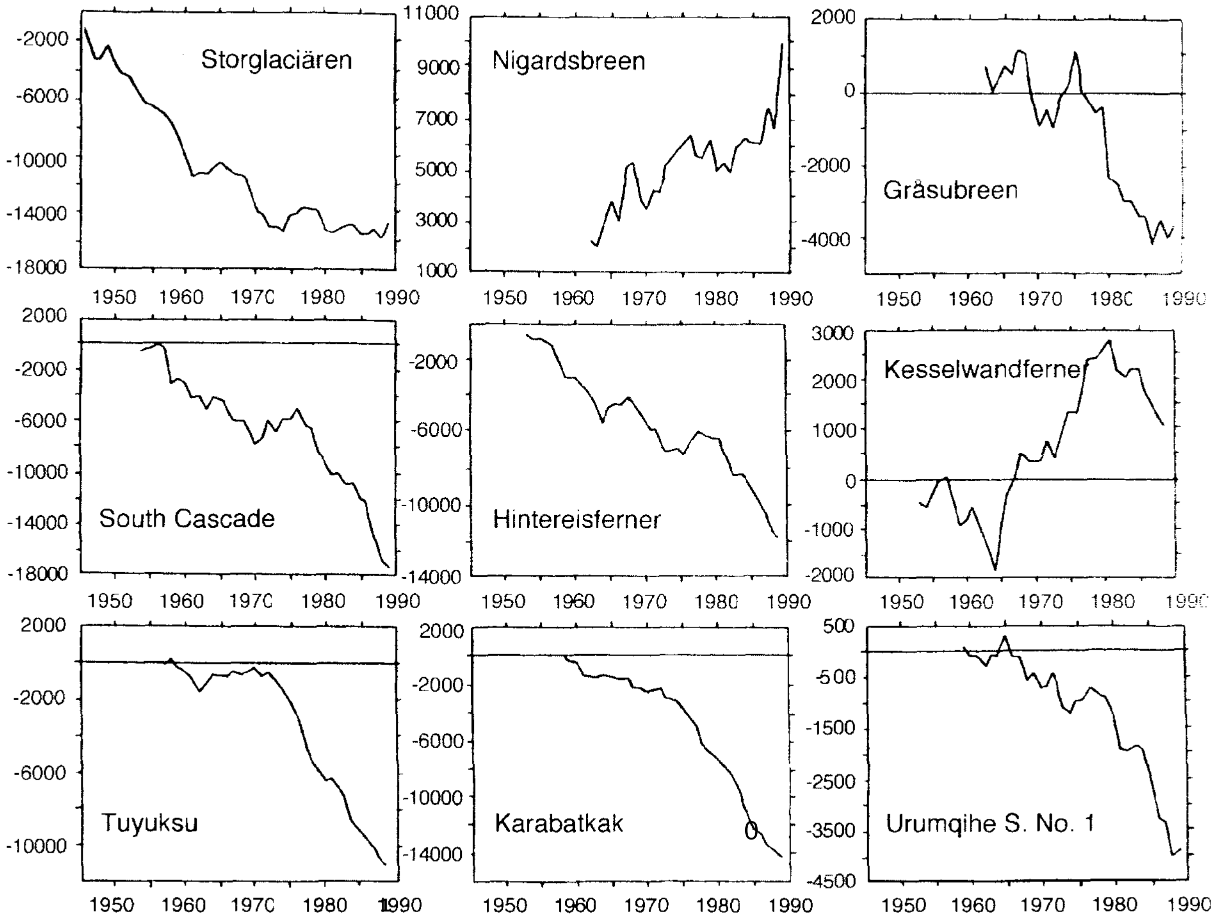


Fig. 12: Examples of cumulative glacier mass balances measured with extensive stake networks (from IAHS (ICSI)/UNEP/UNESC(), 1991).

Shortly after the Second World War, mass balance measurements started in Scandinavia and were soon followed by similar programmes in several other countries; a considerable number of these mass balance programmes were part of the design and operation of hydropower schemes. The essential link between climatic change and glacier length changes was in this way established. An increasing number of glacier advances were now reported from various parts of the World, especially from the Alps where mass balances were predominantly positive between the mid 1960's and the late 1970's. Despite the continued overall tendency for strong ice melt and tongue retreat, the worldwide signal of glacier changes had become much more heterogeneous (Fig. 12) than during the first half of the century. At the same time and for the first time in history, empirical information started to become available about glacier reactions to well documented changes in mass balance. Hence, the basis has been 1 aid for an improved quantitative analysis of secular changes in glacier length.

Collection of standardized glacier fluctuation data today follows recommendations published by UNESCO (1969, 1970, 1973) and regularly updated instructions for submission of data for the publication series Fluctuations of Glaciers (vol. I: 1959-65, vol. 11:1965-70, vol. 1111970-75, V. IV: 1975 - 80, v. V: 1980 - 85). The third and fourth volumes of this series saw a major step towards sumptuary-based processing of data, and in the fifth volume an effort was made to collect internationally and publish short abstracts on special events such as glacier surges, ice avalanches, glacier floods or debris flows, drastic retreats of tidal glaciers, and glacier-volcano interactions. The first biennial Glacier Mass Balance Bulletin was published in 1991 and reports 1989/90 and 1990/91 is presently being prepared. This new publication series is designed to speed up and facilitate access to information concerning mass balance of selected reference glaciers. The results are made more easily understandable through the use of graphic presentation rather than purely numerical data. Thus, 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.

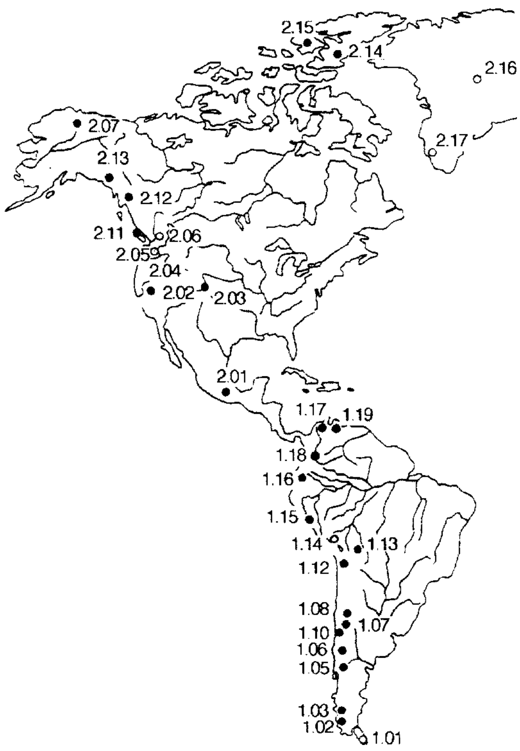
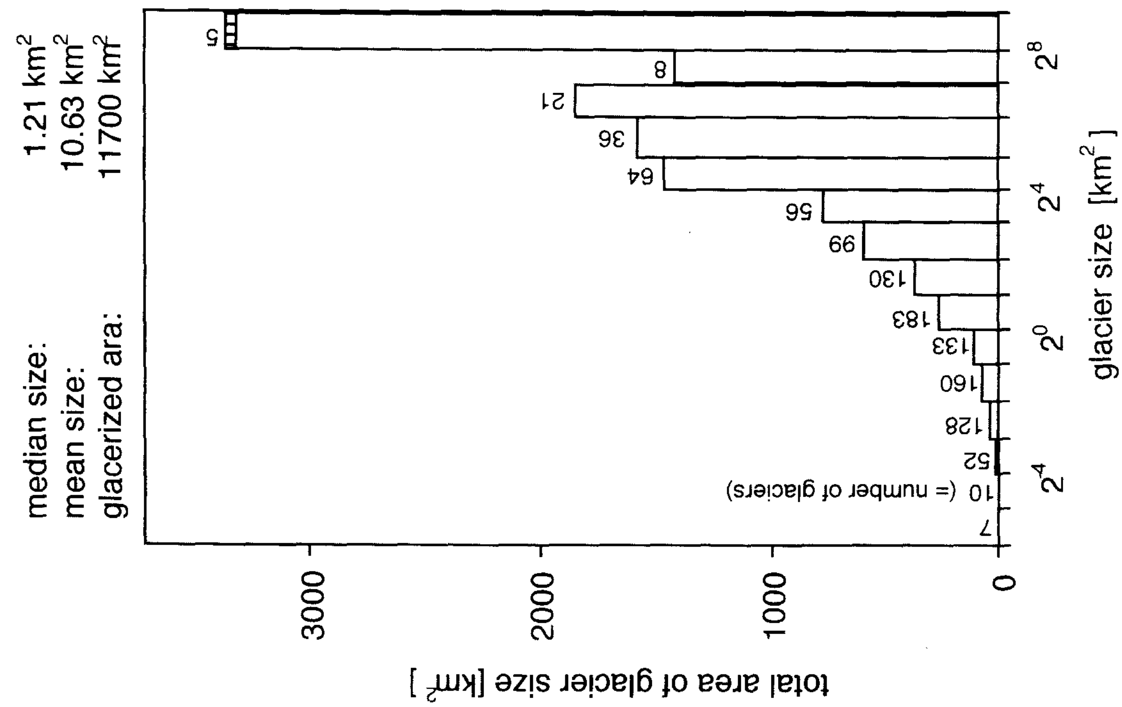


Fig. 13: Glacier inventories of the Americas and Greenland; numbers refer to the listings in IAHS (lCSI/UNEP/UNESCO, 1991). Dot - detailed, circle - preliminary

The international data basis on glaciers of the 20th century not only contains information about changes in time but also glacier inventory data describing the spatial variability of perennial surface-ice masses. The World Glacier Inventory, started in 1976 by ICSI, UNESCO and UNEP, was planned to be a snapshot of ice conditions on Earth during the second half of this century. Detailed and preliminary inventories were compiled all over the World (Fig. 13) to form a modern statistical basis on the geography of glaciers. Areal statisties, 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. 14). 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 - the main source of meltwater contributing to sea level rise - would goon for a long time and without much reduction due to glacier shrinkage. In the second case, however, most glaciers would rapidly be reduced in size and many would even completely disappear. Another example of glacier inventory data and their potential application is the information now available on mean glacier elevation (Fig. 15). 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).



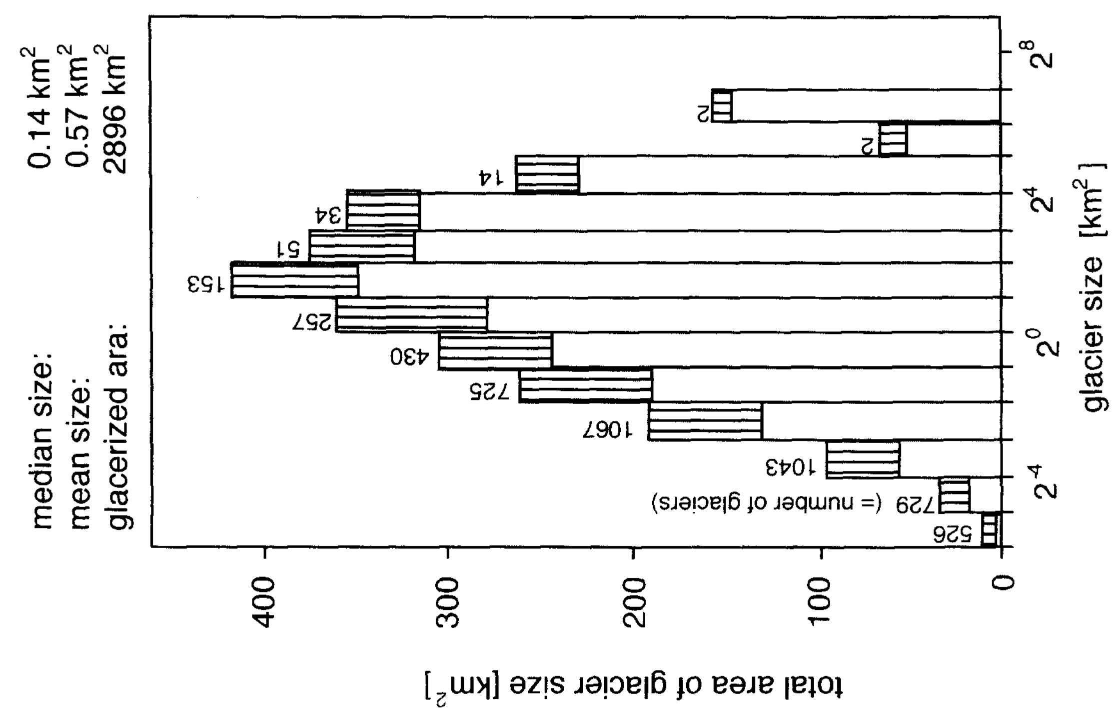


Fig. 14: Size/area-statistics from detailed inventories of glaciers in the Alps (left) and on Axel Heiberg Island, N. W. T., Canada (from IAHS(ICSI)/UNEP/UNESCO, 1989). The hatched areas in the left panel indicate debris cover.

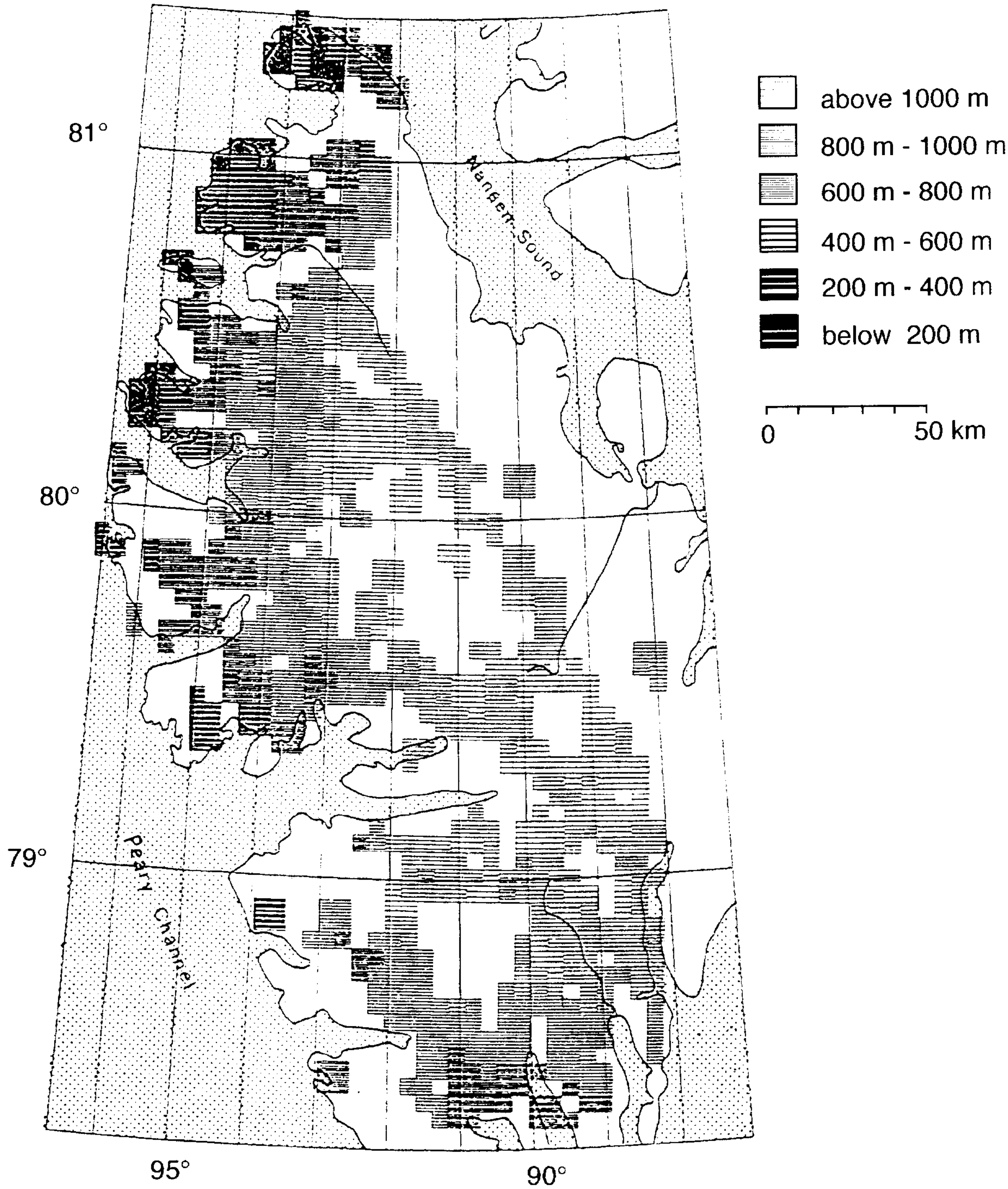


Fig. 15: Distribution pattern of glacier elevations on Axel Heiberg Island, N. W. T., Canada (from IAHS (ICSI)/UNEP/UNESCO, 1989).

Instructions and guidelines for the compilation of standardized glacier inventory data have been developed by UNESCO (1970) and were later updated by Müller et al. (1977), Müller (1978), and Scherler (1983). World Glacier Inventory - Status 1988, published in 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, other preliminary).

Analysis of information from remote sensing of glaciers in regions of difficult access has started in several places (cf., for instance, USGS, 1988, 1989, 1991; Zakharov, 1991). Extensive use of satellite imagery was made especially for compiling preliminary glacier inventories.

Data analysis and results

Glacier mass balance data going back to the 19th century and predominantly collected by the geodetic/photogrammetric method (repeated precision mapping) are available for the European Alps (Table 8). Mean values of specific mass balance b (total mass balance divided by glacier area) over the entire period vary between -0.2 and -0.6 meters per year. The overall reduction in ice thickness since the end of the Little Ice Age is thus measured in tens of meters. This quantitative information represents the main standard concerning secular mass loss of glaciers. Worldwide comparison with this standard is possible via the interpretation of glacier length changes and through statistical analysis of mass balance measurements.

Glacier length variations reflect changes in glacier mass balance over time intervals of years and a few decades; the complex chain of dynamic processes relating mass balance and length changes can be understood and numerically simulated only for a few individual glaciers studied in great detail (cf., for instance, Kruss, 1983; Oerlemans, 1988). The complica - tions, however, disappear if long enough time intervals are analyzed. Over time intervals longer than the response time of the glaciers considered, a simple continuity approach allows calculation of mean mass balance from the easily measured glacier length change (Al). The response time of a glacier is related to the ratio between its maximum thickness and its annual ablation at the terminus (Johannesson et al., 1989). Values for Alpine glaciers are typically several decades to slightly less than a century. This important fact opens the possibility of quantitatively estimating secular trends in mass loss by comparing steady-state conditions; assessing glacier reactions to changing energy input over climatically relevant time intervals therefore does not require full treatment of the complex dynamics related to transient glacier response. Mechanically unstable (surge-type, calving) glaciers 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.

Fig. 11 presents such data on secular length change from the Swiss Alps. The overall picture is a pronounced retreat of glacier tongues, the absolute amount of which being related to gla - tier length. The magnitude of an assumed initial step change in mass balance (Ah) that results in a given Al after full response can be estimated from the ablation at the glacier margin multiplied by $\Delta l/l$, where I is total glacier length (Paterson, 1981). Values of Δb calculated in this way for all categories A through D in Fig. 11 are -0.5 to -1.0 meters per year. Roughly half this amount or -0.25 to -0.5 meters per year corresponds to the average mass loss for the more realistic case of steadily adjusting glaciers. Hence, the observed glacier retreat confirms the representativeness of the directly measured mass balance values. Assuming typical alpine gradients of mass balance with altitude b (0.5 to 1.0 meters per year per 100-m altitude), the observed Δb over the past century corresponds to an increase in equilibrium line altitude (ΔELA) of roughly 50 to 100 m). This result is further confirmed by direct comparison of glacier geometry at the maximum extent of the Little Ice Age and today (Maisch, 1988). Most if not all of this change can be explained by an air temperature increase of about 0.50 C (Kuhn, 1989) as measured on average for the northern hemisphere (UNEP, 1991). About one third of the surface area and half the ice mass originally existing in the Alps at the end of the Little Ice Age has already disappeared as a consequence of the 20 -century warming; further reduction to about one fourth of today's glacierized area and to a few percent of today's volume must be anticipated for the 21st century if the coming decades are markedly warmer than now.

Comparison of glacier mass balance and length changes at low latitudes with much less complete observations from polar regions must take into account climatological aspects and glacier morphology. Changes in precipitation cause stronger ELA-shifts under especially dry or especially wet conditions (Kerschner, 1985). Moreover, b’ increases by more than one order of magnitude from continental to maritime glaciers (Kuhn, 1981). Because b depends on the product of ΔELA times b’, the climatic or regional sensitivity of a glacier’s mass balance is generally much higher in humid / maritime than in dry / continental areas. It is therefore not surprising that principal-component analysis of spatio-temporal distribution patterns classifies long-term glacier mass balances first of all according to “continentality” (Letreguilly and Reynaud, 1990) and that glaciers can gain mass as a reaction to increased precipitation despite simultaneously rising air temperature (Mayo and Trabant, 1984). The relatively homogeneous long-term signal greatly facilitates the interpretation of glacier fluctuation data and confirms the global significance of the early started observational series from Europe and Scandinavia. Topographical aspects cause the local or individual sensitivity of a glacier’s mass balance and relate to the hypsography (Furbish and Andrews, 1984) as well as various feedback mechanisms (cf. Oerlemans, 1988).

Mass loss and retreat of small mountain glaciers were strongest in the first half of the century. This general tendency slowed down considerably and in humid areas even inverted after about 1950 (Haeberli et al., 1989). The overall rate of melt since the past century is nevertheless very high and - as far as the Alps are concerned - considerably exceeds average melt rates during the late Pleistocene vanishing of Ice Age glaciers (Table 9). This fast glacier shrinkage can be assumed so far to have been predominantly natural. It could, however, also easily include man-induced effects. In any case, further continuation or even acceleration of the trend would probably create a situation without historical on Holocene precedent.

Roughly one third of the observed secular sea level rise of 10 to 20 cm percentury may be caused by the melting of mountain glaciers (IPCC, 1990). Due 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 in Iceland are the main contributors (Meier, 1984) and will continue to produce meltwater throughout the next century. In the cold firn areas of the Antarctic and Greenland ice sheets, of smaller ice caps on polar islands, and of mountain glaciers in continental climates at high altitudes, atmospheric warming first causes warming of the firn and does not directly produce additional meltwater runoff. Over the whole accumulation area of Greenland as well as in the marginal parts of Antarctica, temperature gradients in boreholes are clearly inverted between 50 and 100 m depth (Robin, 1983). The magnitude of these anomalies is comparable with those observed in polar permafrost (Lachenbruch et al., 1988), confirming the pronounced secular warming trend. The more recent decadal cooling trend can also be found in many boreholes of the Greenland ice sheet in qualitative accordance with the temperature history as derived from weather-station or oxygen-isotope records (Blatter, 1987; Johnson, 1977; Clausen et al., 1987). In such cold firn areas, refrozen melt layers represent important evidence of meteorological conditions at the surface during dry weather (Beck et al., 1988) and hence are complementary to the oxygen-isotope record, which reflects precipitation temperature. Stratigraphic investigation of firn cores from the Canadian Arctic (Koerner, 1977; cf. also Herron et al., 1981) reveals a pronounced increase in melt-layer frequency since the middle of the 10th century and especially in the first half of the **20’”** century. Secular warming of cold firn seems to be less pronounced at lower latitudes (Haeberli and Funk, 1991).

Monitoring strategies for the 21st century

Today’s strategy of worldwide glacier monitoring attempts to combine various types of information. Observation of glacier length changes remain the most important key to the past and the only 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 have been built up in the Alps or Scandinavia. 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 parameter being the mass balance gradient b’, cf. Kuhn, 1984). Such available long series of extensive mass balance measurements form the basis of our understanding concerning the climate / glacier relationship.

Mass balance measurements on many more glaciers can be made 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 at a regional scale. Glacier inventory data will help in assessing 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 unexplored glaciers in various parts of the world helps in documenting the changing state of glaciers. At the same time, systematically collected reports about special events such as glacier surges, instability of tidal glaciers, glacier floods, ice avalanches and eruptions of ice-clad volcanoes enable glaciologists to share experiences about glacier-related hazards and natural catastrophes.

It is now widely recognized that mass changes of glaciers are among the surest evidence of ongoing changes in the energy balance at the earths surface and therefore that they can be considered to be key parameters for early detection of possible human-induced warming in the near future. Prospects would therefore appear to be good for maintaining and even expanding the existing observational network. With respect to glacier mass balance observations, however, problems and even danger 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 with respect to long-term commitments;
- the growing financial problems, especially in developing countries with changing socio-economic systems;
- the difficulty of finding scientists among the younger generation who are willing and able to carry through the sometimes rather unattractive work for a long time and with the necessary care;
- the 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.

Based on the experience collected so far in the course of operating the World Glacier Monitoring Service and running an international programme of glacier mass balance observation, the following steps may be suggested as reflecting the most urgently needed international activities:

- 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;
- make clear to governmental agencies that long-term observations in glacierized regions are an important part of continuous environmental monitoring.

A century of systematic observations clearly reveals a general shrinkage of mountain glaciers at a global scale. The documented ice melt represents one of the most reliable pieces of evidence for a secular warming trend that is developing at least as fast as - if not even much faster than - the rapid change at the Pleistocene-Holocene transition. The observed development is so far most likely to be mainly natural but could already contain man-induced effects. Within the coming decades, continuation or even acceleration of the trend could create a situation without historical or Holocene precedent. Efforts should not only be continued but strengthened to document ongoing changes adequately within internationally coordinated monitoring programmes.

6. Specific features of the snow and ice regime under the conditions found in Central Asia

The cryosphere in Central Asia plays an important role in the global climate. For example, it is well known that Eurasian snow cover in winter controls the Indian monsoon in the succeeding summer, and recent studies have revealed a relation between the monsoon and the El Nino event in the eastern Pacific as the coupled interaction between cryosphere, atmosphere and ocean. Also, recent studies have pointed out the importance of hydrological processes for the relation between Eurasian snow cover and the summer monsoon through the heat budget for snow melting and evaporation of its melt water. Soil hydrology in the active layer of permafrost on the Tibetan Plateau is considered to play a specific role in this process.

Glaciers in Asia comprise one third of the world glacier coverage in total area excluding Antarctica and Greenland (IAHS (ICSI)/UNEP/UNESCO, 1989). These glaciers are important as a hydrological element of the water cycle at a continental scale and as water resources for Asian countries, especially for arid regions. Also, the contribution of Asian glaciers to global sea-level change is considered to be appreciable. However, one of the major uncertainties in its assessment arise from the relation between climate and glacier variations in this area (National Research Council, 1985).

Many glaciers in the Asian high mountain areas are strongly characterized by their mass balance affected by the summer monsoon and much supraglacial debris, and the uncertainties mentioned above arise from these characteristics. Such specific features of the glacier regime, which play an important role in variations of glaciers in the past and at present, will be described in this chapter.

Glacier mass balance

In Asia, both maritime and continental glaciers are distributed under humid and arid climates, respectively with greatly different annual water exchange. From the viewpoint of the seasonal water cycle, a common characteristic of most of these glaciers, particularly in the Himalayas, the Tibetan Plateau and the surrounding areas is that the annual accumulation is provided mainly in summer - "summer-accumulation type", in contrast to the "winter-accumulation type" glaciers, well known in Europe and North America.

On such glaciers, accumulation and ablation mainly occur simultaneously in summer. Consequently, separate measurements of accumulation and ablation on the summer-accumulation type glaciers pose technical difficulties.

Snowline altitude distributions compiled by Shi and Li (1981) show an inverse in the Himalayas and the southeast Tibet which show a decrease southward in contrast to a decrease northward in north Tibet, the Tienshan and the Altai. Such inverse tendency reflects strong influence of the summer monsoon from the south which provides glacier accumulation, and maritime glaciers develop in this area. In these mountain ranges, glaciers develop more on the south slope which faces the direction of vapour supply than the north slope on the same mountain ridge (Ageta et al., 1991). Consequently, it can be said that the distribution of the maritime glaciers is mainly controlled by accumulation in glacier regime.

On the other hand, the continental glaciers in north Tibet and the Tienshan develop on the north slope which receive weaker solar radiation than the south slope, and their distribution is mainly controlled by ablation in glacier regime. These features of accumulation and ablation of the maritime and continental glaciers in the Himalayas and Tibet will be described in the following sections.

Glacier accumulation

Much accumulation in summer is supplied by the Indian monsoon to the maritime glaciers, which are distributed in the east Himalayas and southeast Tibet. A part of the monsoon precipitation during the warm summer season is provided as rain on the maritime glaciers which extend to lower, warmer altitudes.

Using observed linear relations between the probability of occurrence of solid precipitation and surface air temperature at Glacier AXO1O in east Nepal, accumulation during summer (solid precipitation from June to September) on the whole area of this glacier was estimated to be about 50 % of total summer precipitation (Ageta et al., 1980). This glacier is small and classified into a sub-maritime type. Although this glacier has a small aptitudinal span from 5380 to 4950 m, the portions of solid precipitation in total summer precipitation showed a wide range between about 80 % and 20 %, at the highest and the lowest altitudes, respectively. This example indicates summer air temperature (altitudes of glaciers) as well as total precipitation is an important factor in controlling accumulation (the amount of solid precipitation).

On the other hand, the monsoon circulation does not reach far inland and has no direct influence on the continental glaciers. However, local orographic convection in the arid inland areas provides precipitation mainly in summer to the continental glaciers in central and north Tibet and the Tienshan. Since such glaciers are located at cold altitudes and latitudes, almost all precipitation even in summer is supplied to the glaciers as solid precipitation. However, there is a possibility that climate warming will change snowfall to rain.

For the continental and sub-continental glaciers in Central Asia, which are generally subpolar type, internal accumulation due to refreeze of melt water in/on cold snow and ice (superimposed ice) contributes considerably to accumulation (Ageta et al., 1989, 1991). Consequently, air temperature is also an important factor for the internal accumulation of cold glaciers, since it controls snow and ice temperature.

Glacier ablation

The main heat supply in the surface heat balance of glaciers in Asian high mountain areas is normally solar radiation. The contribution of net radiation to heat sources is more than 80 % for the arid continental glaciers and less than that for the humid maritime glaciers (Shi and Li, 1981). New snow cover with high albedo has the effect of reducing ablation through the radiation balance, especially for the summer-accumulation type glaciers, since much snowfall occurs during the season favorable for ablation. However, summer snowfall sometimes changes to rain under warm air temperature conditions in the maritime glaciers area as mentioned in the last section, and new snow cover melts away quickly under such conditions. Consequently, summer air temperature is an important factor which controls ablation through albedo variation at the glacier surface of the maritime glaciers.

Characteristics of ablation of the continental glaciers, attributable to the arid inland climate, can be seen in comparison with the relations between ablation and air temperature of glaciers in the West Kunlun and Nepal Himalayas. As shown in Fig. 16, similar amounts of daily ablation occur on the continental glacier at air temperatures around 4 “ C lower than that in Nepal, under air temperature conditions below about -20 C averages over 6 day periods in this case. This suggests a large contribution of evaporation-sublimation to ablation due to the continental climate. Since the latent heat of evaporation-sublimation is much larger than that of melting, such heat use is effective to reduce total ablation of the continental glaciers.

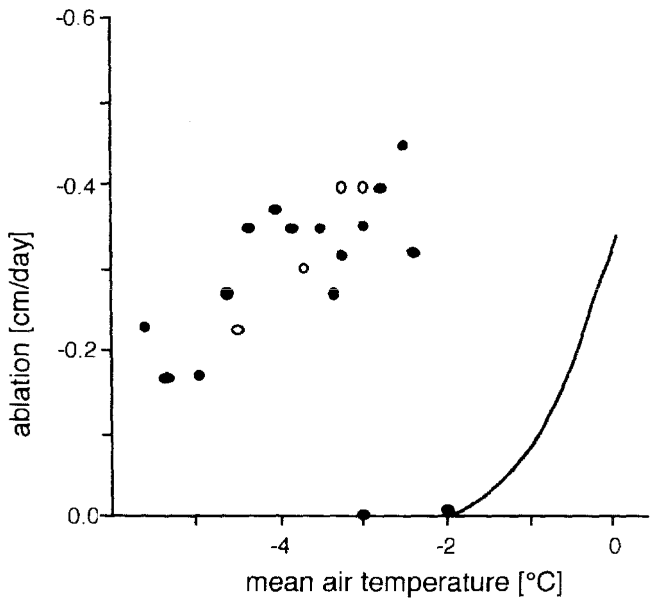
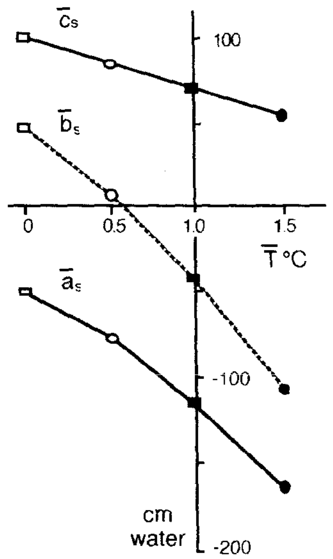


Fig. 16: Relations between ablation and mean air temperature during the period from July 29 to August 4 at every station along 2 lines (solid and open circles) of the Chogce Ice Cap in the West Kunlun. A curve shows the relation in each half month during summer of Glacier AX 010 in the Nepal Himalaya (Ageta et al., 1989).

Mass balance of glaciers and its variation



Fig, 17: The relations between summer mean air temperature (T) and accumulation (es), ablatiou (as) and balance (b) during summer in the area-averaged values for the whole area of Glacier AX010 in east Nepal (Ageta, 1983).

As described in the above two sections, summer air temperature is one of the important factors for the controlling of both accumulation and ablation, especially of the maritime and sub-maritime glaciers. Estimated relations between summer mean air temperature and accumulation, ablation and balance during summer are shown in Fig. 17 in the area-averaged values for the whole area of Glacier AX010 in east Nepal in the case of fixed summer precipitation. As seen in the figure, accumulation decreases with temperature rise reflecting an increase in the proportion of rain in the total precipitation, and ablation increase is accelcrated by the lowering of surface albedo. As a result, from the negative variations of accumulation and ablation both with temperature rise, variation of balance shows high dependency on summer air temperature.

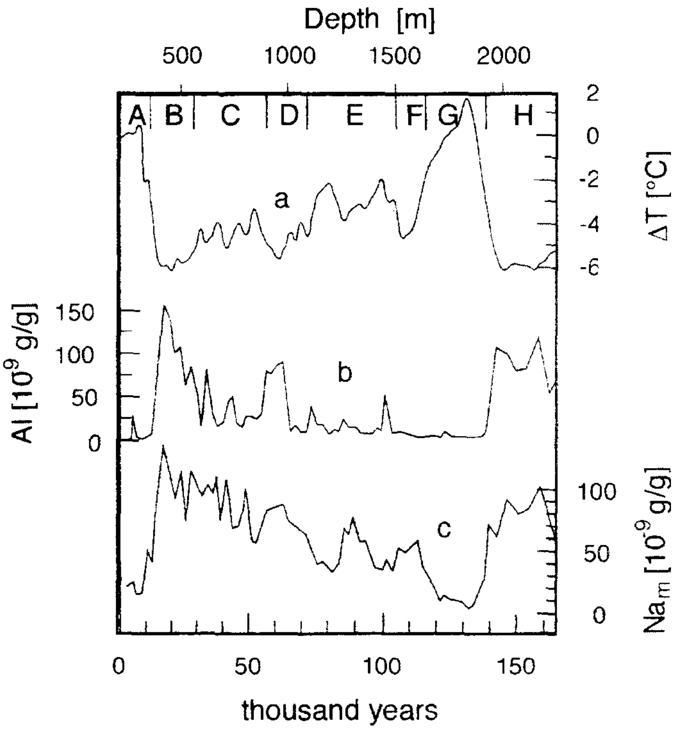


Fig. 3: Concentration of continental and marine aerosols in an ice core extracted from a borehole at the Vostok Station (De Angelis et al., 1987; Legrand et al., 1988):

a) Paleotemperature deviations from the present-day temperature values;

b) Aluminum concentration in ice;

c) Sodium concentration in ice.

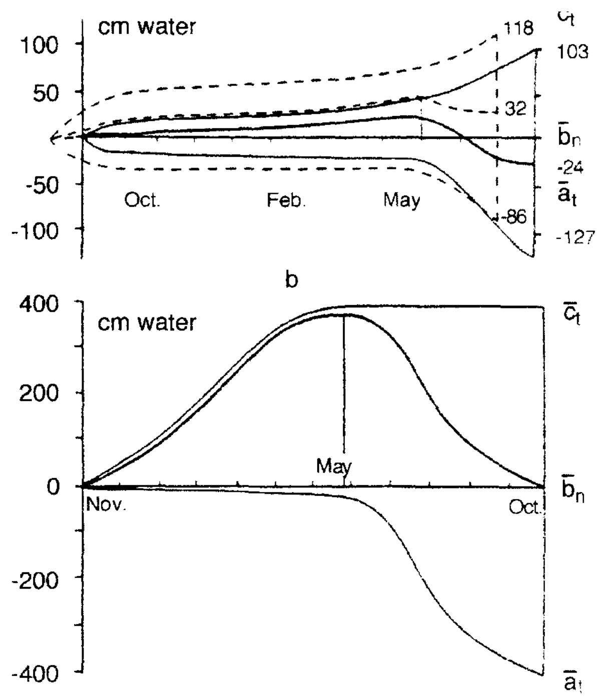


Fig. 18: Variation of the cumulative value of mass balance component in the area-average through the 'balance year'. c_t— total accummulation, a_t— total ablation, b_n— net balance (Ageta and Higuchi, 1984).

a) the summer-accumulation type: Glacier AX010 in Nepal Himalaya. Dashed line: case of 0.5 °C lower temperatures (Ageta, 1983).

h) the minter accumulation type: et, at and the maximum value and its time of balance are referred to the results of South Cascade Glacier in U. S. A., 1964 -05.

Smoothed variation of the cumulative value of accumulation, ablation and balance through a balance year on Glacier AX010 is shown in Fig. 18 a, as an example of the summer-accumulation type glacier. For comparison, an example in the case of the winter-accumulation type glacier is shown in Fig. 18 b. It can be seen in these figures that the amplitude of the variation of the cumulative balance through a balance year is small in the case of the summer-accumulation type, since much accumulation and ablation occur simultaneously in summer. Besides, it can be said on the summer-accumulation type that the period (length and season) of "winter", "summer" and "balance year" which are defined by UNESCO/IAHS (1970) is unstable and sensitively changeable depending on air temperature, as shown with the solid line and dashed line in Fig. 3 a. This result suggests that the "fixed data system" has more generality than the "stratigraphic system" in the definitions of mass balance terms by UNESCO/IAHS (1970).

Some characteristics of mass balance of the summer-accumulation type glaciers have been described comparing the maritime and the continental glaciers in Asia in this chapter. Seasonal water cycles through glacier mass balance have some variance in each glacierized area and those may change interannually in some areas due to variation of the general circulation. Therefore, further studies on the effect of seasonal features of glacier mass balance to variation of the regional water cycle are required.

Debris-covered glaciers

There are debris-covered glaciers, sometimes called moraine-covered glaciers, in many regions of the world. Regions with high occurrence are the Andes, the Himalayas, southeast Tibet, the Karakorum, the Pamir and New Zealand. These are glaciers in which debris are accumulated in the ablation area. These regions are located at the periphery of the tectonic plate of the Earth's surface where it corresponds to subduction zone or site of frequent deep earthquakes, which enhances debris supply to the glacier surface mainly in the accumulation area. There are debris-covered glaciers in other regions including North America, the European Alps and Antarctica, but they seem to be formed due to uplifting of the basal tills in the ablation area or accumulation of volcanic dust.

The study of the debris-covered glaciers has been made mainly by scientists in the Former Soviet Union, China and Japan. Not much work has been done by scientists in North America or Europe, as there are not many debris-covered glaciers in their regions.

The importance of the debris cover lies in the fact that their existence either enhances or suppresses ice melting in the ablation area. This affects the mass balance and aptitudinal distribution of glaciers. Furthermore, by changing the morphology of the glacier surface, this in turn has an effect on the dynamic characteristics of the glacier. Due to these processes, such glaciers show different responses to climatic changes in comparison with debris-free glaciers.

This section will concentrate on the characteristics and problems related to debris-covered glaciers mainly referring to studies in the Nepal Himalaya made by Japanese scientists.

A glacier inventory has been made for the Dudh Koshi Region in Eastern Nepal (Higuchi et al., 1978). Statistical analysis for three drainage areas, Bhote Koshi, Dudh Koshi and Imja Khola near Mt. Everest, shows that 35 % of the total glacier area (287 sq.km) is covered by debris. Assuming 40 % accumulation area ratio, nearly 50-60 % of the total ablation area is covered by debris. This is due to the fact that generally large glaciers longer than 2 km all have debris-covered parts in the ablation area (Fuji and Higuchi, 1977). These facts mean that debris cover cannot be dismissed in the discussion of glacier hydrology, ice melting and glacier response to climate.

From an intensive study on the Khumbu , the debris on the glacier was determined to be supplied from the rock wall in the accumulation area as shown in Fig. 19 (Fushimi et al., 1980). The amount of debris is small at higher altitude (9 km from terminus) and gradually increases to 1.2 m thick at 4.2 km from the terminus, and it is much thicker near the terminus.

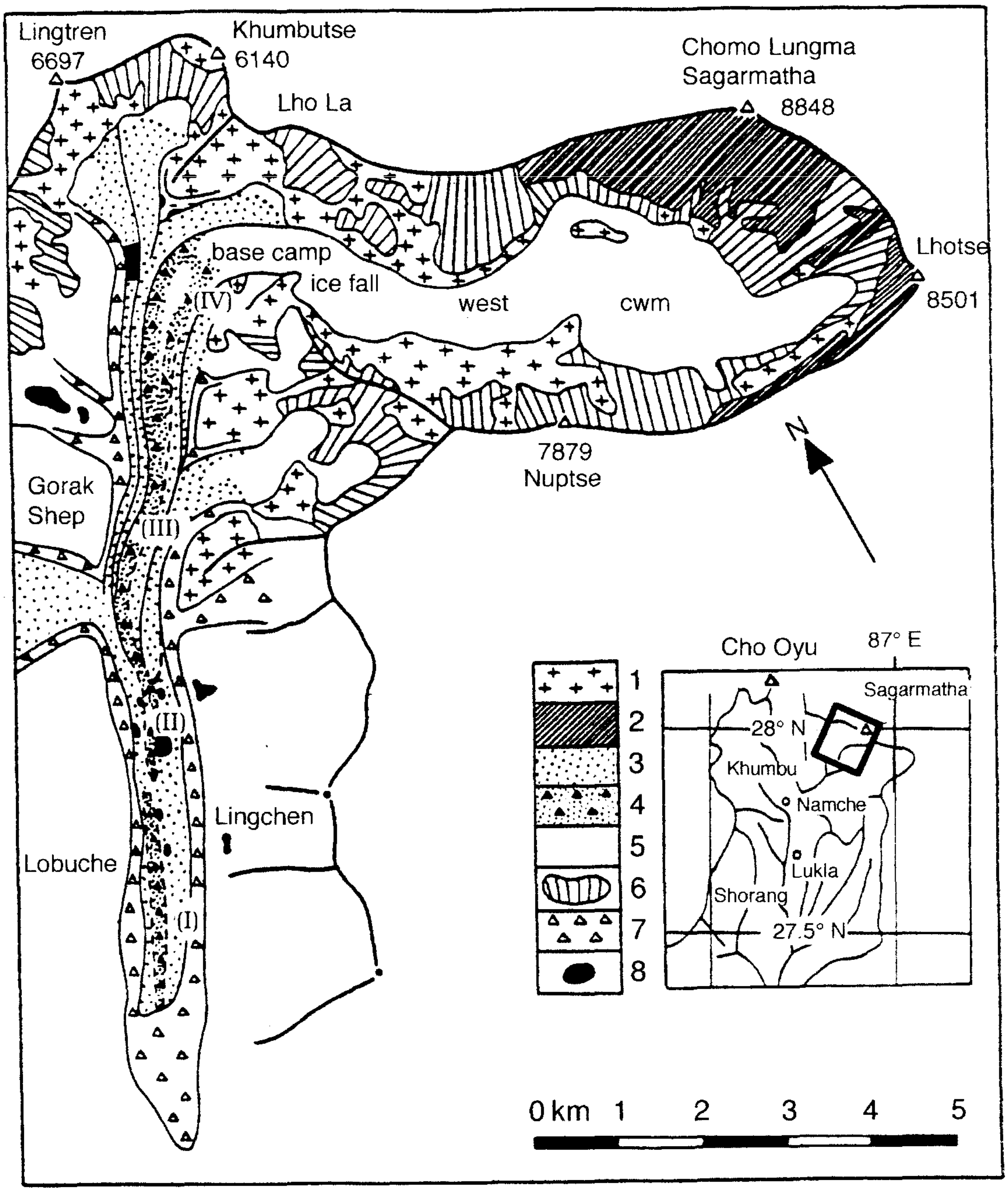


Fig. 19: Distribution of the supraglacial debris and the rock types, Khumbu Glacier, Khumbu region, east Nepal. 1) ogranitic rocks, 2) schistose rocks, 3) supraglacial granitic debris, 4) supraglacial schistose debris, 5) debris-pee part, 6) snow cover, 7) side- and end-moraines and 8) ponds and lakes, (Fushimi et al., 1980).

Observations (Dolgushin et al., 1972; Demchenko and Sokolov, 1982) and field experiments (Fuji, 1976) have e shown that sand or rocks on the glacier surface in general enhance melting when the layer is thin and retard melting when it is thick. The thickness of debris whose melting amount drops to half value of debris-free condition is in the range of 5 - 10 cm. Model calculation has been made to discuss this effect depending on climatological and debris conditions (Bozhinsky et al., 1986; Nakawo and Young, 1980). Areal observation results on Khumbu Glacier (Inoue and Yoshida, 1980) show similar melting amount where debris is thin and increase drastically as debris thickens. There is a limit in the one dimensional treatment, as the debris-covered area shows uneven surface topography and existence of ice cliffs (Iwata et al., 1980)

Up to present, there has been no work done on the variability of the debris-covered glaciers. Melting amounts on steady-state glaciers are usually comparable to the accumulation amounts. The existence of debris changes the distribution of melting, and influences the determination of the position of the terminus. For example, the area up to 4-5 km from the terminus of Khumbu Glacier has negligible melting, and also the flow is considered to be small (Kodama and Mae, 1976). This means that, if there were not debris, the terminus of the glacier would be 4-5 km shorter than the present 19 km, excluding the dynamic effect. It can be said that the glacier size is maintained due to the existence of debris. There are other glaciers in the same region where the debris-covered surface dominates 60-70 % of the glacier (c. f. Lhotse Glacier), and in this case, such effect should be much greater. Thus the length and volume of the debris-covered glacier is enlarged by their existence, and probably has a weak relationship with the present climate.

The variability of the debris-covered glaciers is a poorly studied subject. Debris-free glaciers are said to have steady state under stable climate apart from surging glaciers. However, debris on debris-covered glaciers cannot be considered to be stationary, because debris is constantly supplied to the surface through ice melt, although at a small rate. In order to clarify this subject, the debris budget in the glaciated area need to be investigated.

We need to answer the following questions in order to understand the variability of debris-covered glaciers.

(a) Is the depth of debris thicker near the terminus for all debris-covered glaciers? Does it have any relation to the inclination of the surface?

(b) What is the debris concentration in glacier ice?

(c) What are the movement processes of debris within the glacier, and from glacier surface to lateral moraines and vice versa? If there were inclination at the surface or uneven melting predominated, debris would be moving in a certain direction.

(d) How will the suppressed surface melting in the lower part affect the surface morphology and dynamics of glaciers?

(e) Is the amount of debris now increasing, decreasing or constant on the glaciers?

(f) How are the response characteristics of debris-covered glaciers different from the debris-free glaciers under the same conditions of climate change?

In order to investigate these problems, intensive measurements on characteristics of debris on glaciers, comparative study of relation between debris and glacier morphology, and simulation studies using a model are needed. Furthermore, there is need to standardize technical terms used for characterizing debris-covered glaciers and related physical processes.

7. The role of snow and ice in natural processes on local, regional and global scales[1]

7.1. Introduction

Some of the most distinct climatic contrasts, for example between winter and summer or ice ages and interglacial, are associated with changes in snow and ice at the earth's surface. The presence of ice on land has effects of a magnitude similar to or greater than those of vegetation or soil moisture as far as the boundary conditions for atmospheric circulation are concerned. In the hydrological cycle snow has a remarkable seasonal storing capacity. Ice sheets can increase surface elevation by kilometers, their development leads to sea level changes and on a 100,000 years time scale they depress the earth's crust and trigger lithospheric feedbacks.

Apart from their direct and indirect climatic influence glaciers and ice sheets are of interest to climatologists both in predicting future and unriddling past climatic situations: their slow motion at meters to kilometers per year, long turnover times and large thermal inertia make long term forecasts in the cryosphere more meaningful than in the atmosphere. On the other hand, moraines and other geomorphic traces of past ice extent have significantly promoted paleoclimatic reconstruction. Finally the polar ice sheets contain ice that originated in the atmosphere hundreds of thousands of years ago. With the appropriate technology ice cores are now being used as archives containing information on temperatures, precipitation, dust and trace gas content of the atmosphere as far back as 150,000 years.

7.2. An inventory of Ice and Snow

Global ice masses and snow covered areas have been summarized by Kotlyakov in chapter 1. The mass in long term storage is roughly 30.1018 kg and if it were melted and evenly distributed over today's oceans would form a layer 83 m thick. It would take roughly 1025 J to melt this ice, and, supporing this was accomplished in 10,000 years, would require a heat flux density of 0.06 W m^{-2} when averaged globally, or 2 W m^{-2} when averaged over the land area permanently covered with ice today.

Of the global average annual precipitation of nearly 1 m (1000 kg m^{-2}) it is estimated that about 6 percent (60 kg m^{-2}) falls as snow. A correspondingly much higher amount of water is taken into seasonal storage in those areas actually affected by snow fall. A spring snow pack of 1 m water equivalent, which is not uncommon in midlatitude mountains, requires 64 W m^{-2} in order to melt in a period of two months while on a global, annual average solid precipitation turns over 0.6 W m^{-2}.

Both sea ice and seasonal snow cover extent fluctuate from year to year. An evaluation of satellite records by Ropelewsky (1986) resulted in a 1973-85 mean northern hemisphere snow cover of 32 . 10^6 km^2 with maxima ranging from 42 to 52 . 10^6 km^2.

[1] This chapter contains material that has been published in *Understanding Climatic Change,* Geophysical Monograp h 52, IUGG Volume 7, edited by A. Berger, R.E. Dickinson and J.W. Kidson, copyright American Geophysical Union 1989.

7.3. Boundary conditions

As a boundary of atmospheric circulation and of energy and mass exchange between air and soil, snow and ice covered surfaces have the following characteristics that distinguish them from bare soil:

- High albedo
- Evaporation at the potential rate
- Stable surface boundary layers
- Strong thermal insulation, suppression of soil heat flux

All of these properties make snow and ice colder than bare soil under comparable external conditions. Their combined climatic effect is:

- Local temperature inversions
- Baroclinic boundary layers and associated small scale circulation
- Large scale differential heating and associated synoptic effects.

7.3.1 The surface energy balance of snow and ice

We shall treat the surface energy balance in the conventional way, i.e. express the distribution of energy gained from net radiation R to sensible heat flux H, latent heat flux of evaporation LE and heat fluxes associated with temperature changes S or melting M in the snow or underlying ground.

$$R = H + LE + S + M \qquad (1)$$

where the individual terms are most conveniently taken as energy flux densities (W m^{-2}). Table 1 presents examples for melting snow and for Antarctic mid-winter conditions. In all four of these examples the turbulent flux of sensible heat is directed towards the surface. In the cases a and b sensible heat and net radiation supply heat for snow melt, in the two antarctic cases sensible heat has to replace the energy loss of net radiation. In all four examples energy fluxes into the snow and evaporation from the snow surface play a subordinate role.

Table *1* *Examples of snow surface energy balances.*
R net radiation, H sensible heat, LE latent heat of evaporation, M latent heat of melting, S heat flux into snow, all expressed as heat flux densities, W m^{-2}

	R	=	H	+	LE	+	S	+	M
a) Hintereisferner, 2960 m, 15 July -18 August 1971	66	=	-32	+	3	+	0	+	95
b) Bad Lake, Saskatchewan 27-30 March 1976 (Male and Gray 1981)	14	=	-8	+	1	+	2	+	19
c) Plateau Station (80°S, 40°E, 3625m), June + July 1967	-15	=	-13	+	0	-	2		
d) Maudheim (75°S, 11°E, sea level), June + July 1951	-23	=	-15	-	4	-	4		

The formulation chosen here applies to the snow surface only. Since the snow pack is translucent to shortwave radiation and permeable to air, water vapor and liquid water, energy and mass is being exchanged internally with the effect that the size of snow grains increases with age.

7.3.2 Albedo

Increasing grain size, liquid water content and impurities lead to a lowering of snow albedo. As the snow metamorphism progresses faster at higher temperatures, an absorption-temperature feedback will accelerate the ripening of the snow pack. Reflection from the snow does not take place in a specular fashion at the surface but mainly by multiple scattering within the snow. The emerging flux shows selective absorption by ice and liquid H_2O bands and displays anisotropic spatial distribution.
Low sun and high albedo make measurements and modelling of polar radiation fluxes highly susceptible to errors: at 10 degrees solar elevation and an albedo of 0.8, the value of the direct component of absorbed shortwave radiation is in error by 10 percent if either albedo is wrong by 0.02 or or the level of the instrument is off by 1 degree.
Nonetheless, for the annual, energy weighted albedo of dry snow which ranges from 0.82 to 0.85, a modal value of 0.84 seems to be a reliable upper value (Kuhn et al. 1977, Koerner 1980). For obvious reasons there are no other fixed points in the albedo scale. The albedo of „ wet" snow, whatever that is, will range widely about 0.7, and alpine summer snow (with a density of 500 to 600 kg m^{-3} and a grain diameter of 1 to 5 mm) has an albedo typically between 0.5 and 0.6 depending on dust contamination. Clean, snow free glacier ice has albedo values near 0.4, depending on surface roughness, cracks and air bubbles, while the albedo of dust or debris covered ice may go down to 0.15.
Vague as these figures may seem, they are easier to handle in a global model than the spatially and temporally highly variable albedo of midlatitude model grid points, representing an area, say, 100 by 100 km. A study of late winter development of albedo of Eastern US surfaces by Robinson and Kukla (1986) shows farmland albedo changing from dry snow values in February to a snow free 0.2, while forest albedo declines from 0.2 to 0.1 and a variety of surfaces range in between.

7.3.3 Thermal Insulation by the Snow Cover

The equations of molecular heat conduction and ensuing temperature changes are sufficiently well known. For given density, heat capacity and conductivity, the thermal diffusivity and the downward progression of a harmonic temperature change at the surface can be computed. Table 3 summarizes relevant values for snow, ice, sand and rock. The depth D_a at which the amplitude of the annual temperature wave is damped to 1 / e of its surface value illustrates the insulating properties of the material, low density snow being the best insulator of the example given in *table 3*. Higher frequency fluctuations penetrate less efficiently, the daily wave, for instance, to a depth D_d= 365-'/2 Da= 0.05 D_a. Dry winter snow, which generally is of density less than 300 kg m^{-2} is thus a very efficent insulator between soil and atmosphere. Compared to wet soil, rock or ice, low density snow effectively reduces subsurface heat storage. This situation can be broadly summarized by saying that the dry snow energy balance is an unbuffered, quasi-instant conversion of sensible heat gain into radiative heat loss (see *Table 2, c and d)* and vice versa.
To give an arbitrary example, I found the February heat content of an alpine snow pack at 3000 m elevation to be only 15 MJ m^{-2} less than that of a zero-degree snow pack of same mass, a difference that can be restored by the energy gain of only two summer days.

Table 2. Penetration of Annual Temperature Wave Damping of Amplitude to 0,37 of its Surface Value at Depth D_a.

Material	Density ka m^{-3}	Conductivity m^{-1} K^{-1}	Diffusivity $m^2 s^{-1}$	Dam
Snow	50	0.007	0.07.106	0.8
	100	0.03	0.15	1.2
	300	0.27	0.45	2.1
Ice	917	2.5	1.28	3.6
Sand	dry	0.19	0.22	1.5
	40% water	2.2	0.76	2.7
Rock	Granite	3.7	1.9	4.4

7.3.4 Stability of Cryospheric Boundary Layers

The thermal decoupling of cold surfaces from the warmer atmosphere is enhanced by a temperature-stability feedback. Consider the midwinter situation of an antarctic energy budget where approximately R = H *(see Table 1)*, net radiation loss being a function of surface temperature T_o, and H being determined by the free atmosphere - surface temperature difference T_a -T. as well as by a stability dependent transfer coefficient α_H.

$$R = L_d - \sigma T_o^4 = \alpha_H \, (T_o - T_a) \tag{2}$$

A disturbance in the radiative forcing (where L_d is the longwave downward flux) will lead to a change in To that includes a gain of several percent from the stability dependence of α_H.

7.4. Synoptic and Large Scale Effects

While local modification of the surface energy can be expressed in simple, analytical terms, the synoptic and planetary wave scale effects of snow cover have to be simulated in general circulation models. In this section two examples of large scale simulations will be discussed.

Dewey (1986) demonstrated that in winters with extensive snow cover the number of cyclones increases over the southern part of the US. Increasing cyclonic activity accompanies the southward extension of the snow cover. In Eastern Canada in turn, cyclones become less frequent when the margin of the snow cover moves south.

A far reaching experiment was recently carried out by Barnett et al. (1988) who investigated the effect of Eurasian snow cover on global climate, especially on the Indian monsoon. When doubling the Eurasian snow cover in their simulation they found that it retarded the warming of the Asian land masses, thereby increasing surface pressure by up to 8 hPa, cooling the surface to 200 hPa layer by 1.7 to 3.6 K. Since ocean temperatures changed little, this caused a weaker meridional temperature gradient and a weaker subtropical easterly jet. The results of these simulations, especially the reduction of Indian precipitation by about 300 mm, agree well with what one observes in poor monsoon years. Reduction of Eurasian snow cover to half yielded similar results with opposite sign.

The authors found teleconnections via Rossby waves to the pressure field in the west Pacific and in Western Tropical Africa. The similarity with winter seasons of ENSO events leads them to remark that “snow fall perturbations may act as trigger for some ENSO events. They cannot directly force an ENSO since their characteristic time scale is about one season.. .“

While the two reports quoted above treat seasonal snow cover, the growth of ice sheets to heights above 3000 m adds a third dimension to the forcing af atmospheric motion by snow and ice. By inserting ice age (18 ky BP) topography into a simple two-layer model Lindemann and Oerlemanns (1987) found that orographic forcing was more important than thermal forcing. According to their model the Laurentide and Fennoscandian ice sheets produce a wave number three anomaly in the 500 hPa height. Note that the Greenland and the Ural ice sheets do not have any significant influence on this pattern. The negative anomalies over the eastern part of the large ice sheets develop further by a planetary wave feedback.

The fact that in the wave number three pattern of this circulation model a third, large ice sheet is missing while Greenland, the Ural Mountains and the Central Asian Highlands bear ice covers underlines the importance of land surface elevation for ice sheet development.

7.5. Glacier Mass Balance

The mass balance of an existing glacier or ice sheet is dominated by atmospheric processes such as solid precipitation P, redistribution by snow drift D, deposition or erosion of snow by avalanches A, Evaporation E, melt M and calving C. P, D, C and E are strongly influenced by topographic features, E and M only to the extent that surface aspect governs the energy budget. The mass budget of an entire ice body is

$$B = P + D + A + E + M + C \qquad (3)$$

where P, D, A are usually positive (accumulation) and C, E, M negative (ablation).

Specific mass balance terms are referred to unit horizontal surface area and customarily y denoted by lower case letters. The line connecting all points where the annual value b = 0 is called the equilibrium line. Further details are given in earlier papers (Kuhn 1981, 1986).

A stationary or steady state glacier keeps its shape by annually transporting net specific mass gain from the accumulation area across the equilibrium line to the ablation area. By dividing annual mass gain in the accumulation area by the cross-sectional area at the equilibrium line, a characteristic velocity of the ice flow is derived. This has typical values of 10 to 10^2 m yr^{-1} in mountain glaciers and 10^3 to 10^4 m yr^{-1} in polar outlet glaciers. Considering the sizes of glaciers and ice sheets one finds mean residence times of ice of less than 1000 years in polar ice caps, with extremes probably reaching 200 000 years.

From the areas and volumes given in chapter 1, the typical thickness of a glacier is of order 100 m, that of ice sheets is in the km-range. Comparing the life times and dimensions of these ice bodies to those of seasonal snow one finds a remarkable gap in the size and age spectrum. This impression is confirmed by glacier inventories and by individual observations.

While the seasonal snow cover may be gradually dissolved in smaller patches and vanish in spring, the majority of glaciers and ice sheets have a sharp boundary. In the time domain this is paralleled by rapid transitions from glacials to interglacial or the switching back and forth at higher frequent y between less long-lived stable states of the climate system.

An obvious requirement for such sharp or rapid transitions is the existence of feedbacks involving snow and ice, promoted by the potential of ice to suffer catastrophic, dynamic changes.

7.6. The Albedo - Temperature Feedback

We have mentioned so far two feedbacks, one involving surface layer instability and the other involving orographic forcing of planetary waves. The one most often quoted in connection with snow and ice is the albedo-temperature feedback. With rising global temperature, planetary albedo is diminished by the melting of ice thus adding to the original forcing. This feedback was the central theme of the classical papers of Sellers (1 969) and Budyko (1969).

Hansen et al. (1984) have modelled the effects of doubling CO_2 or increasing the solar constant by 2 % and found that an original temperature change of about 2 K (similar in both cases) was amplified by various feedbacks to a total of more than 4 K. about 0.4 K of this being due to the surface albedo changes.

As has been mentioned, planetary albedo is higher than surface albedo over most of the globe but lower (about 0.75) than surface albedo (0.84) over the dry snow facies of the Antarctic

Plateau. This masking of surface contrasts by atmospheric backscattering and absorption decreases the efficiency of the albedo feedback. Shine et al. (1984), among others, have pointed out that cloudiness is higher over snow free than over snow covered surfaces, a fact that further contributes to reduce the surface albedo feedback.

7.7. Snow and Ice in the Hydrological Cycle

Apart from the various ways in which ice and snow cover modifies the climate via the surface energy budget, some climatically relevant processes involve mass transport or mass storage in the cryosphere. These range from the daily freeze and thaw cycle to glacial sea level changes.

7.7.1 The Seasonal Cycle

As quantified in chapter 1, an appreciable amount of water is seasonally stored on the land surface as snow. Compared to no-snow conditions on the same soil this means delayed runoff and prolonged evaporation at the potential rate not only from the snow but also from the soil moisture that is recharged by meltwater. The hydrologic effect is not restricted to the local scale: many semi-arid areas receive an important fraction of their water balance from ice and snow melt in distant mountains. This is particularly important in summer-dry regions with sufficient winter precipitation that would otherwise drain unused.

While the peak of annual runoff occurs during spring in seasonally snow covered land it is further shifted to July and' August in glacierized mountain regions. Under alpine conditions the following relation between the glacierized fraction of a basin and the month with peak runoff was observed:

glacierized fraction	83 %	Peak runoff in	August
	44		July / August
	11		July
	2		June

7.7.2 A Possible Runoff Change in a Warming Climate

To a useful approximation the limit of snow fall is determined by the O°C-isotherm. Following a climatic rise of mean temperatures, then, considerable fractions of a region may lose their seasonal snow cover and consequently change their annual river runoff characteristics.

An example can be given on the basis of the mean monthly temperatures recorded in Austria. The present climate conditions favor a seasonal snow cover over the entire country. Frequent valley inversions cause nearly isothermal conditions from 1000 to 1500 m elevation at minimum monthly means of -3 to -4 °C. A five-degree warming would be sufficient to shift the lower limit of winter snow cover to elevations above 1500 m. This effect may be experienced in other mountains or highlands as well. It is best developed where the changes with elevation of temperature are small and those of area are large.

7.7.3 Sea Level Changes

Difficult as the determination of recent sea level changes may be on a global scale, there is evidence that since 1900 sea level has risen about 10 cm. Of this rise more than half is caused by thermal expansion of the oceans, the rest being attributed to glacier melt.

From a survey of the restricted number of mass balance measurements of sufficient length Meier (1984) found that 30 mm could be attributed to this century's melting. The possible future contribution of these glaciers to sea level rise following a hypothetical four-dgree global warming is expected to be limited to another 120 mm (Kuhn 1992).

The additional contributions to sea level rise from Greenland and Antarctica are still largely unknown. Whether South Greenland surface ice melt, wasting of antarctic ice shelves from below and storage of possible increased snow fall in polar regions are all insignificant compared to extrapolar glacier melt or whether they simple have cancelled in this century is still a matter of stimulating discussion.

8. The future of glaciers under the expected climate warming[1]

The results of modeling carried out on the basis of global circulation models (National Research Council, 1985; Intergovernmental Panel, 1990) as well as experimental works on the fact of a parallel course of changes in paleotemperatures and in the concentration of atmospheric carbon dioxide throughout the last glacial-interglacial cycle (Kotlyakov and Lorius, 1992; Lorius et al., 1990) indicate that the rapid increase in the concentration of green house gases causes on equally rapid rise in global and local temperatures. A twofold increase in the content of atmospheric CO_2 may result in a global temperature increase by 2 - 4° C; such conditions are predicted for the 2030s.

Let us consider two consequences of such warming: one characterized by a rise of the level of the World Ocean, and the other bearing on changes in the glacier-derived runoff on the continents. The aim of this chapter is to show that when estimating the amplitudes of the forthcoming rise of the ocean level, it is important to avoid mistakes stemming from (i) neglecting the role of iceberg detachment in overall losses of ice and (ii) the unwarranted assumption of a linear dependence between the temperature rise ant the glaciation decrease.

The mechanism of glacio-eustatic changes of the ocean level, due to be activated by the imminent warming, will amount not only to changes in the rates of glacier melting but, by and large, to fluctuations in intensity of iceberg detachment from marine ice sheets; i. e., those whose morphology and dynamics are largely determined by their interaction with the ocean.

The main cause of polar ice sheet evacuation onto the ocean lies in the instability of inland ice sheets (Maze, 1989) which either collapse under changing conditions or, if they are expanding, adopt a new mode of existence characterized by additional stabilizing mechanisms of ice discharge. When inland ice sheets flow down high-latitude trough valleys, their stabilization occurs not until they reach the ocean and start interacting with it, thus turning into “marine” glaciers. An important element of this interaction is the stabilizing of icebergs into the ocean.

At present the overall mass of the largest “marine” ice sheet in West Antarctica accounts for 15 percent of the contemporary glaciation of the Earth, while in the latest maximum of glaciation, the “marine” parts of the East Antarctic, Pan-Arctic and South Bering Ice Sheets must have constituted over 50 percent of the entire Pleistocene ice (Grosswald, 1983). Besides the large glacier complexes of the Antarctic type, the Arctic ice-cap islands and the tidewater glaciers of North and South America now have a significant role to play in interacting with the ocean.

Marine glaciers (ice sheets) are the most unstable element of glaciation; the indicator and frequently the cause of this instability are related to the position of the grounding line depending on the floating of glaciers, i.e., conditions of the transition from the inland portions of marine glaciers to floating ice shelves. A rapid retreat of the grounding line entails catastrophic consequences, including discharges of a large ice masses from the inland regions of a marine ice sheet to the ocean.

The grounding line cannot be stabilized at all on a horizontal bed. In this case a marine ice sheet will shrink until it leaves the ocean, that is, becomes an inland ice sheet, or will grow until it reaches the margin of a continental shelf where the sea bottom inclination increases abruptly, a factor that stabilizes marine ice sheets.

[1] The present chapter is based on an article contributed by V. Kotlyakov and coauthors (1991)

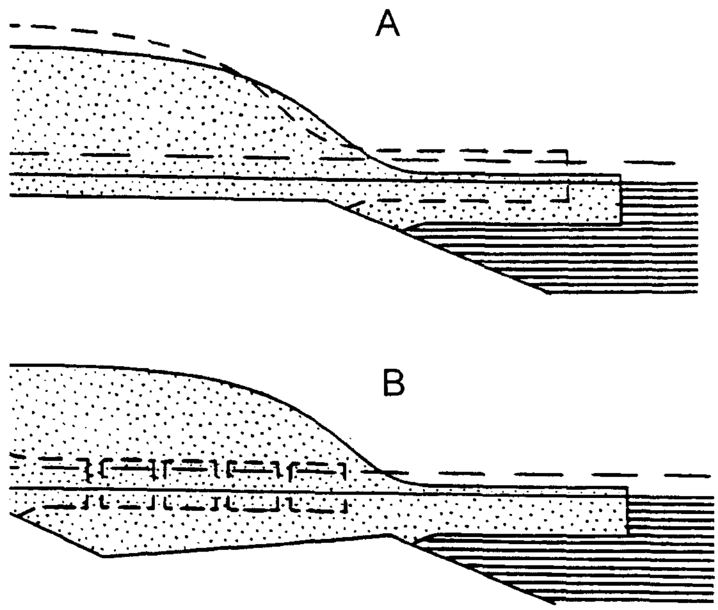


Fig. 20: *Stable (a) and unstable (b) response of a marine ice sheet to changes in the sea level with respect to different glacier bed profiles.*

The dynamic factors accounting for the instability of marine glaciers maybe activated only in specific geomorphological conditions typical of glacial shelves, that is, when an ice shelf has a marginal riegel separating the inverse surface gradients (Fig.20). The grounding line ought to lie immediately after the marginal riegel separating the inverse surface gradient and the continental shelf slope. In that case the instability of a marine ice sheet will lead to its disastrous degradation, or shrinkage.

Thus a marine ice sheet is stabilized if the grounding line is on a subjacent surface inclined seaward; but it is unstable if the grounding line lies on a subjacent surface inclined inward. The reaction of a marine glacier to changes in external conditions depends on the grounding line position: when the latter is critical, even insignificant changes in external conditions may elicit dramatic changes in the state of a marine glacier. The consequences thereof may be catastrophic indeed: a rapid retreat of the grounding line, a discharge of significant masses of ice into the ocean, and a relatively fast rise of the ocean level.

The shrinkage of ice sheets in consequence of climatic warming may exhibit a nonlinear dependence and, therefor maybe catastrophic due to their structural instability. For this reason all scenarios postulated for changes in the ocean level will be erroneous if they proceed from the linearity of the glaciation decrease/temperature dependence. Thus the basal scenario, used by many scientists for prognostication and assuming that the sea level will keep rising at the same rate as at present, provides for a rise by only dozens of centimeters toward the middle of the next century (117 cm by the maximal model, and 5 to 24 cm according to the minimal one).

Changes in mountain glaciers (those connected, either directly or via rivers, with the ocean) may likewise be responsible for ocean level fluctuations. Such glaciers account for the bunk of mountain glaciation; their total area is close to 500,000 km² (IAHS (ICSI)/UNEP/UNESCO, 1989). But there are also glaciers without an outlet to the ocean; the largest of such regions is in Central Asia: the mountain glaciation of the Tien Shari, the Pamirs, the Kunlun Mountains, as well as parts of the glacial systems of the Karakoram and Tsiliang Shari Mountain Ranges, and those of the Tibetan Mountains. The overall area of glaciers in this region is close to 40,000 km², with the Tien Shari alone accounting for 19,000 km².

There are various different methods for the prognostication of changes in mountain glaciation. The most common one is to estimate ice melting or ablation by the prognosticated data on meteorological characteristics (air temperature, precipitation) and also according to the snow line dynamics. The problem consists in computing only a part of the glacier-derived runoff, exceeding the runoff value under the conditions of glacier mass balance stationarity and, consequent y, the stationarity of glaciation. This means that with respect to climatic warming and glacier shrinkage it will be necessary to compute the value of the meltion-derived runoff Rd, equal to the glacier mass balance. The Rd value affects changes in the ocean level and those in the humidity of a particular territory. In computing the glacier meltion-derived runoff, one ought to proceed from data on changes in the glacier mass which tends to shrink in warmer periods.

Observations carried out in the Tien Shari Mountains show that the area of glaciers located in regions with dispersive glaciation shrinks by 0.5 percent annually, while that of glaciers in regions of compact glaciation by 0.1 percent. Accordingly, the annual reduction of the Central Asian glacier area may be assumed at 0.5 percent; as to the extrapolar glaciation of the Earth connected with the ocean, the annual shrinkage figure could be assumed to stand at 0.3 percent (though the decrease in the height of Arctic glaciers may follow a highly irregular pattern due to their island position and the dome-like form).

0.5 percent for annual shrinkage of Central Asian glaciers looks like a high figure. But the point is that in warmer periods the rate of glacier shrinkage (degradation) increases with the continentality of the climate, for the summer mass balance of glaciation increases its contribution to changes in the annual mass balance and in the glacier-derived runoff. This is related to a slow but steady growth of the proportion of winter precipitation in the total annual amount under conditions of warming and stronger cyclonic activity during winter. Besides, as it is shown in Fig, 21, the glaciers of continental regions tend to receive less nourishment from atmospheric precipitation. A decrease in snow accumulation and a tendency toward warming are typical of Central Asia's inland regions. This most unfavorable combination of factors accounts for the anomalously high rates of glacier shrinkage and the high runoff value R_d as it is evident from Fig. 22.

In order to determine the degradation-derived runoff R_d and predict its possible dynamics relative to the extrapolar glaciation of the Earth. One has used data on the mass balance of Norwegian glaciers: this precise and most reliable evidence collected over many years reflects the standard conditions for a possible response of glaciation to global warming. The glaciation of Scandinavia's coastal regions with their characteristic maritime climate (Garseg, 1955) bears some similarity to the glaciation regime of the US and Canadian Coast Range, and of the offshore mountain ranges of Alaska, Iceland, Spitsbergen etc. Therefore the averaged tendency, obtained for six Norwegian glaciers (Fig. 23), has been adopted as a basis for predicting the possible waste of the entire, 500,000 km^2 in area.

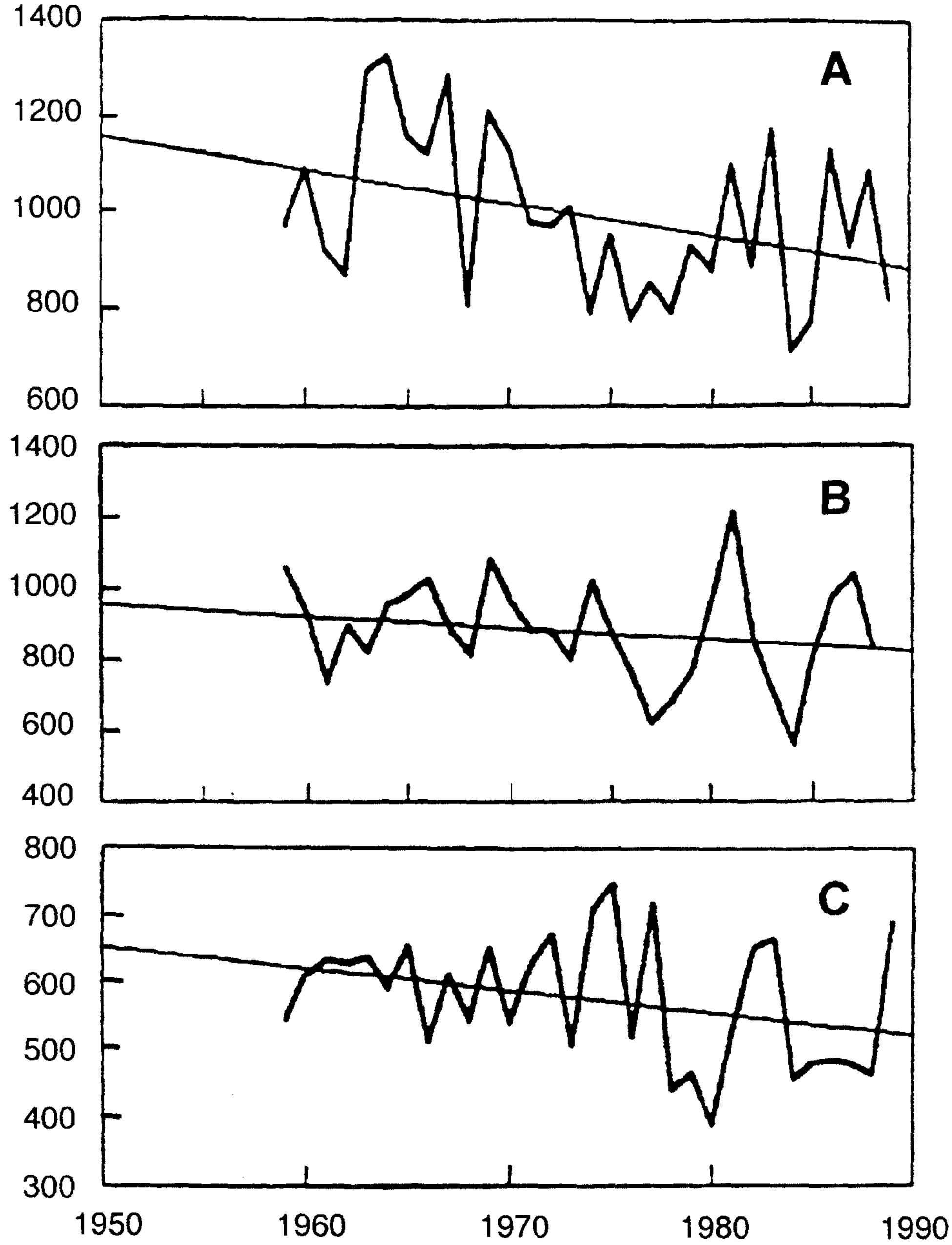


Fig. 21: Time series of the annual accumulation of snow on the glaciers Tuyusksu in the Ala Tau mountains (A), Sary Tor in the Inner Tien Shan (B) and No. 1 in the Chinese Tien Shan (C).

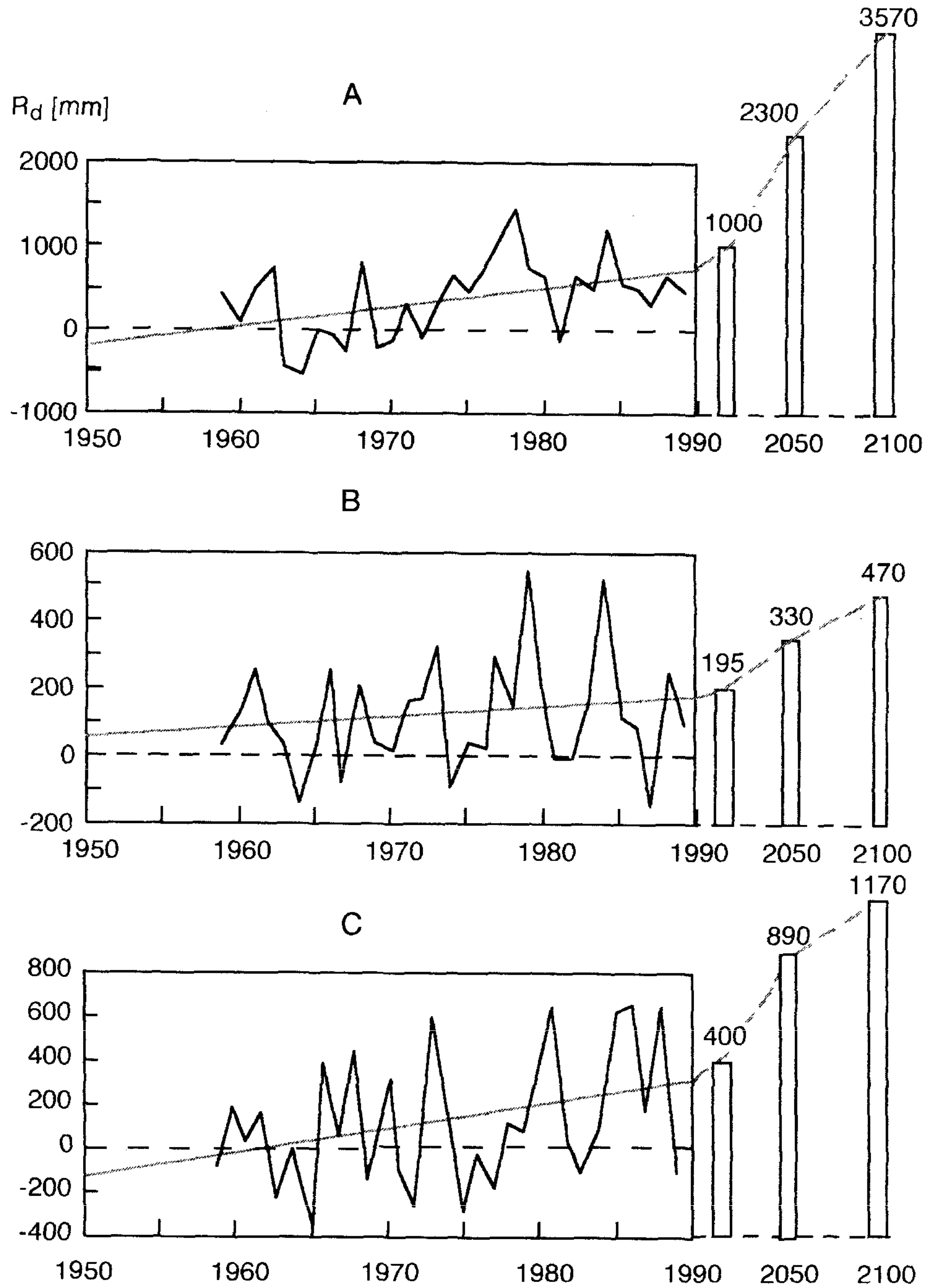


Fig. 22: The present trend in the glacier degradation-derived runoff R_d and the prognosticated values computed relevant to this trend according to the data on the glaciers Tuyuksu (A), Sary Tor (B) and No 1 (C).

Taking air temperature records of the Tien Shari Weather Station and assuming that the same linear trend is to keep up (unfortunately, linear extrapolation is inevitable here), we may find that the mean annual temperature in Central Asia may go up 1.5° C by the year 2350.
Proceeding from these rough estimates, in Tables 10, and 11 we have derived figures for glaciation shrinkage DS, changes in the specific glacier melt runoff R_d and the volume of this runoff QR_d, as well as for the overall rise of the ocean level Σ SL. These data apply to Central Asian glaciers and all of the extrapolar glaciation of the Earth.

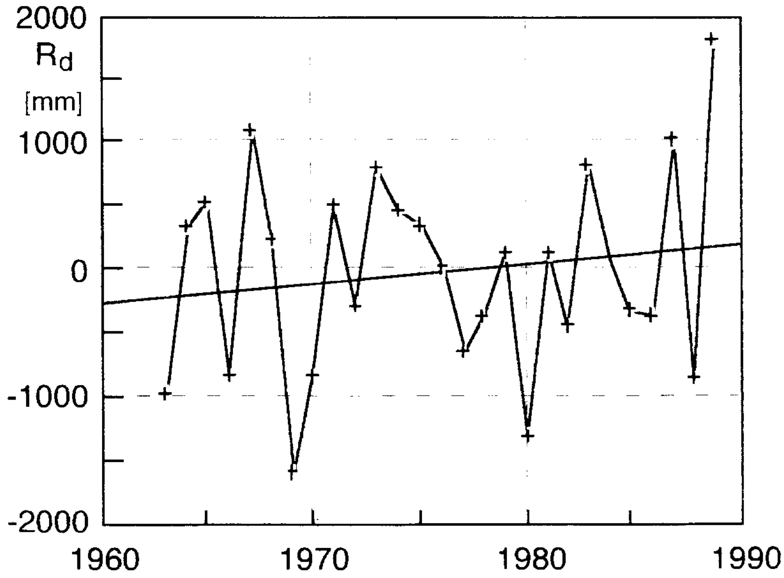


Fig. 23: Degradation-derived runoff R_d according to data on the mass balance of six Norwegian glaciers.

There has been a dramatic shrinkage-----d late even at catastrophic rates—in the glaciation area of the Tien Shari Mountains and within Central Asia. This results in a growing intensity of deglaciation-derived runoff and in its volumetric increase. The increment maximum (relative to 1975) is to be attained by the year 21100, with a 3.5- fold increase in the absolute volume of melt-derived runoff; this volume will be declining with subsequent warming. The scope of the glaciation area decrease is impressive indeed. Glaciers will remain only in the high mountain parts of Central Asia, in compact glaciation areas.

With the further progress of warming or stabilization of the present climate, the territory, particularly its low and mid-mountain parts, will be turning into a high-mountain desert land with periodic spells of rain and snowfall in the spring and summer seasons.

We should also take into consideration the extremely unfavorable conditions for the snow-melt runoff. It will decrease dramatically, with the peak of snow melting shifted to the early spring period. Under like conditions, the pathway of transit runoff from glacier regions to lowlands is bound to increase and thus activate melt-water channeling and filtration processes. The runoff increment will diminish in real terms, also will the increment of the runoff of mountain rivers. It is only thanks to the water yielded by melting glaciers that the water sup - ply problem will not become critical.

The degradation of the extrapolar glaciation of the Earth will be apparent in rising ocean level already by the year 2050, and there will be a drastic rise of the ocean thereafter caused by the deglaciation-derived runoff (see Table 11). This period will last from 200 to 300 years. The extrapolar glaciation of the Earth will be decaying at rapid, catastrophic rates—its total area will shrink from 500,000 to 100,000 km² by the year 2350. Glaciers will survive only in the mountains of inner Alaska, on some Arctic archipelagos, within Patagonian ice sheets, in the Karakoram Mountains, in the Himalayas, in some regions of Tibet and on the highest mountain peaks in the temperature latitudes.

Thus the glaciation shrinkage as a result of the “green house” effect will entail highly negative implications in geoecological terms. Changes in the structurally unstable ice sheets of the “marine” type may culminate in their catastrophic decay and, as a consequence, in a relatively rapid rise of the ocean Ievel (5 to 7 m in several decades). The mountain glaciation of temperate and subtropical regions, as well as the Arctic ice-cap island will, under such extreme regime conditions, develop an abruptly negative mass balance (-3 or even -5 m/year) and will practically disappear. Accordingly, the glacier-nourished runoff of rivers will drop dramatically, with the ensuing negative effects for the farm industry.

9 Tasks for Future Glaciological Research

Present day glaciological studies are motivated both by the need for better basic understanding of the fluctuation of glaciers and by the need to manage ice and its meltwater as resources of ever increasing value. While it seems obvious that future research will continue to follow these two general objectives there is less certainty about details of practical approaches.

Experience shows that research teams at universities tend to adjust their planning for future investigations to their individual man power, knowhow and technical capabilities, national agencies tend to concentrate on regional problems. A global problem like the fluctuation of glaciers requires global management of scientific work and monitoring, Through its IHP IV project 4.2 UNESCO has set a good example of guidelines for future research. Through the establishment of the World Glacier Monitoring Service a basic step was taken toward unification and homogenization of world wide monitoring of glacier fluctuations.

In view of these facts I believe that the first task for future activities is to work out a program for scientific research connected with glacier fluctuations, a frame or guide lines that may direct ongoing and future studies. In view of the scientific capabilities of its membership, the International Commission on Snow and Ice is ideally suited for this purpose.

There exist many examples of how national and international organizations have attacked the problem of evaluating the consequences climatic change could have for certain natural processes or human activities:

All of them have started with an account of what the scientific community knows about the problem. This is exactly what is being done with the present ICSI Technical Report on Variations of Snow and Ice in the Past and at Present on Global and Regional Scales.

All of them have recognized the need for a sound data base but few were in such an axcellent position as the climate/glacier community who have at their disposal direct mass balance studies of four decades, direct observations of glacier lengths of one century, direct climate records of nearly two centuries, and moraines, tree rings and proxy data to cover the millenium.

All acknowledged the importance of continued and intensified data collection but few had anything equivalent to the World Glacier Monitoring Service established in anticipation of the development in a wise act of science policy.

Most used these assets and developed scientific programs to fill the gaps in present day knowledge of their problems, many joined forces with other groups.

If the International Commission of Snow and Ice would design a science program to study glacier fluctuations it would most likely contain, among others, the following points.

Improve inventories of glacier length, area and volume. Establish typical volume/area ratios for various topographic, climatic and dynamic conditions. Prepare profiles of volume/altitude distribution for selected glaciers as basis for modelling.

Compare time series of mass balance and length variations and work out response times, suitable two and three dimensional models, simulate known surging glaciers, verify short term events as the alpine advance of 1965-83.

Complete late and post glacial data sets. Concentrate on medieval fluctuations and on minima in Little Ice Age. Correlate past fluctuations with past climate records and proxy data.

Establish global patterns of advance and retreat on long and short time scales. Explain or model regional and local differences in glacier fluctuations.

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