PROPERTIES OF SURFACE RUNOFF IN THE HIGH ARCTIC

by

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ABSTRACT

The Mecham River is a stream near Resolute, Cornwallis I., N.W.T., which has an arctic nival regime. It flows only in summer, from June or early July to late August or early September. Discharge begins shortly after the snowpack becomes fully ripe under the influence of increasing temperature and radiation. Over a period of about two weeks it rises rapidly to a spring snowmelt peak from which there is a gradual recession through the rest of the summer. Superimposed on the recession there are (in most summers) occasional rainstorm responses. These floods are rapid to rise and to fall, because the bare terrain with shallow underlying permafrost has little capacity to detain or retain water. Diurnal fluctuations are also superimposed on the seasonal pattern of discharge; they can be related to diurnal fluctuations of incoming radiation, and although they are sometimes subdued late in the season they can be very pronounced before rainwater runoff begins to drown them out.

The most important findings of the study concern the annual water balance of the Mecham River basin. Several terms of possible importance in the water balance can be eliminated. Changes in frozen water storage and in lake storage may be neglected because the basin contains very few snowbanks and lakes. Annual changes in the moisture content of the active layer may also be neglected, although more understanding of how the active layer behaves hydrologically would be valuable for other purposes besides the solution of the water balance. The flux of windblown snow across the drainage divide has been suspected to be large, at least potentially, but an analysis of basin topography shows that this is unlikely to be so in basins as large as that of the Mecham River. The remaining terms in the water balance are snowfall, rainfall, evaporation and discharge.

Three of these are measured at Resolute or on the Mecham River itself. The fourth - evaporation - must be calculated indirectly. The method chosen in this study was an application of the combination model to the weather data collected routinely at Resolute. The hardest problem encountered in developing the model was the accurate estimation of surface wetness,

although improvements would also be desirable in measurements of the wind speed profile, the soil heat flux and perhaps in the estimation of basin snow cover. Nevertheless, comparisons of the results with pan evaporation measurements suggest that the model performs well. Comparisons with Priestley and Taylor's approach to the estimation of the vapour flux are also encouraging: their empirical coefficient is larger at Resolute (1.47) than elsewhere (1.26), probably because Resolute is a windy place.

An accounting of the water balance reveals gross discrepancies between inputs (precipitation) and outputs (evaporation and discharge). A small numerical allowance for unrecorded trace precipitation does not come near to removing the disparity, which is such that on average outputs exceed measured inputs by 400 per cent. Most of the disparity must be attributed to underestimation of precipitation, which is thus about four times greater than has been supposed in this "arid" locality.

Measurements of sediment concentration in the Mecham River suggest that, if the amount of water in hydrologic cycle has been underestimated, so too has its effectiveness as a geomorphic agent. Having a

gentle slope and a coarse-grained channel bed, the Mecham River carries little bed load, but its suspended sediment load is comparable with those measured in comparable physiographic and lithologic situations at lower latitudes. The largest fraction of its sediment load is made up of dissolved Ca, Mg and HCO_3. In comparison with warmer waters draining better-vegetated carbonate terrains, the Mecham waters have low solute concentrations, though the solute load is by no means negligible. Average Ca concentrations in the Mecham River and its basin are those which would be expected (allowing for an inverse relationship with discharge) in waters at equilibrium with atmospheric CO_2 and near saturation with respect to calcite. Even moderate admixtures of biogenic CO_2 can be shown to increase the aggressivity of these High Arctic waters towards calcite, and the notion that more limestone is dissolved at lower temperatures (sc., in cold periglacial regions) can now be considered dispelled.

For Jan

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SECTION 1

INTRODUCTION

> "We step and do not step into the same river. We are and are not. You cannot step twice into the same river, for fresh water is ever flowing in upon you."
>
> Heraclitus, fragment.

1.1 General Introduction

1.1.1 Preamble The phrase "water cycle" denotes a concept which is very firmly and deeply embedded in hydrological discourse. Water evaporates from oceanic surfaces, travels through the atmosphere and falls from the clouds as precipitation. The water reaching land surfaces in this way runs overland, collects in channels and flows through the channels back to the oceans. Hydrology is the branch of natural science which deals mainly with water in the "terrestrial" phase of this cycle.

The water cycle is an oversimplification of what is in fact a system of continual flux in all directions, involving changes of state between vapour, liquid and solid. Close examination reveals a wealth of detail in the flux which is obscured in the grossly conceived cycle. Water in, on and near the ground moves in different directions, in changing quantities, at continuously varying

rates; these movements may indeed be viewed at any of a number of scales of time and area, and at each of these scales distinctive properties become dominant.

The solid and dissolved impurities which are entrained by water travelling over and beneath the ground, or in open channels, form only a small part of the total mass in motion. Nevertheless, they are of great interest in themselves, not least to students of geomorphology, for water is the principal agent by which landscapes are worn down and the materials of the earth's surface shifted from place to place.

In addition to having exercised Man's curiosity since before the time of Heraclitus, all of these properties of water in motion near the earth's surface touch closely the daily life of every individual and every society. People in the civilized world take water and its availability for granted on each day of their lives, but in a place such as the High Arctic this can only be done where civilization has established toeholds in isolated settlements. These settlements need reliable supplies of water if they are to survive, and must be built with due regard to the vagaries of a not particularly accommodating water regime.

This study is an effort to make clear some of the properties of the water regime of the High Arctic, and in

particular the properties of surface runoff and sediment movement in the High Arctic.

1.1.2 Goals and History The goals of the study can be divided into two groups, those which are mainly hydrologic and those which are mainly geomorphic. At the outset, factual information on the hydrology and fluvial geomorphology of the High Arctic was very limited. One goal, then, was to make a start on the development of a data base which would be useful in future applied and basic research. Several simple questions will illustrate this point.

Are there any streams in the High Arctic which can be expected to flow throughout the year, or do stream channels invariably freeze to the bottom and fill with snow at the end of each summer? If there is no winter flow, what are the average dates on which flow begins and ends, and what range may be expected in the actual dates in a sample number of years? Given that a stream has begun to flow in spring, how long will it take for snowmelt discharge to rise to a flood peak before it begins to decrease? Should streams be expected to continue flowing until temperatures fall at the start of winter, or is it more usual for channels to dry up when they are no longer nourished by the winter snowpack? What happens to drainage systems when rain falls in the High

Arctic summer, bearing in mind the peculiarities of the terrain such as its barrenness and the underlying permafrost?

These are questions dealing with the hydrologic regime or seasonal pattern of discharge, but other questions naturally suggest themselves. In a given basin, what are the maximum water levels and discharges which may be expected in a given year, or in a number of years? Is the maximum annual discharge always produced during spring snowmelt, or do responses to rainfall sometimes exceed the snowmelt flood? Knowing that the thawed layer above the frost table is very thin, what can be learned about the capacity of High Arctic catchments to store water? Are ratios of snowfall to runoff and rainfall to runoff higher under these circumstances than in lower latitudes, where deep groundwater reservoirs are important in the flux of water? How do the particular characteristics of High Arctic terrains affect the rapidity with which hydrologic systems respond? Streams carrying meltwater from glaciers in temperate latitudes are known to discharge more water later in each day, some hours after the diurnal peak of solar radiation: is this true also of high latitude streams which carry mainly snowmelt?

Answers, some of them tentative, can now be given to most of these questions. Some could have been answered from general reasoning and from extrapolation of results gathered elsewhere. Others will only be answerable when more extensive data collection allows a study of regional differences within the High Arctic to be made effectively. But once a basic set of data is available it is natural to think of putting it to more effective use in prediction rather than simply in description.

The behaviour of drainage basins in general is now well understood theoretically, although success in practical application often lags behind the theoretical grasp which hydrology has of its subject matter. For practical purposes the hydrologic individuality of the High Arctic makes some of the tasks of prediction more easy and some more difficult. For example, the relative unimportance of groundwater means that the complications of deep storage and recharge can be eliminated. But since snow is such a conspicuous feature of high latitude drainage basins a second class of fluxes must be allowed for in the analysis of runoff: these are the energy fluxes of the snow pack, which control the immediate production of water during most of the hydrologically active year. To know about the amount of water "entering"

a drainage basin it is not enough simply to install a precipitation gauge or gauges, and even with instruments to measure incoming energy the problems of translating energy into its equivalent of meltwater and then routing the meltwater to the discharge measurement station are added complications.

The work done for this study on predicting runoff - in general, and in some of the detail hinted at in the questions posed above - has advanced less far than work towards a goal which was much less prominent at the beginning. One of the questions to be asked was, How is the flux of water disposed in the High Arctic on an annual basis? In other words, how accurately can the several terms in the annual water balance of a drainage basin be reconciled with each other? It became clear as time went on that this annual budget could not be balanced accurately at all, and that if the discrepancies could be resolved the effort of analysis would be repaid in insight gained. A number of terms required close attention to ensure that they were being estimated as accurately as possible, and to try if possible to put beyond doubt the implication of the discrepancy.

This implication is that precipitation is grossly underestimated in the area chosen for study, and probably

in high latitudes more generally. Since a number of convergent lines of approach (discussed later) all point to the same conclusion, the result of this study is not entirely unexpected. It does, however, corroborate earlier and current findings of climatic research (e.g. Hare and Hay, 1971), using more complete and accurate hydrologic data and focussing on events with relatively fine resolution.

There is a tendency, less marked in recent years, to think of the High Arctic as a climatic desert. The classical criterion that mean annual precipitation should not exceed 10 inches (254 mm) is amply fulfilled, for the 10-inch isohyet runs several hundred kilometres to the south, and in most schemes of climatic classification the word "arid" would be applied to the entire region. Yet this is unfortunate, not only because this study and others indicate that precipitation is greater than has been supposed, but also because the climatic view has influenced the percepts of the High Arctic taken in by other branches of science.

Geomorphology is one such discipline. The second major goal of this study, and the first-conceived in time, is to show that more geomorphic work is done in periglacial regions by running water than would, for

example, be suspected from a review of some standard texts on the subject. Of the hundred pages in "Geomorphology of Cold Environments" (Tricart, 1969) which deal with periglacial processes and landforms, only five are devoted to fluvial activity; the coverage is on a similar scale in Bird's "The Physiography of Arctic Canada" (1967), Embleton and King's "Glacial and Periglacial Geomorphology" (1968), Davies' "Landforms of Cold Climates" (1969) and Pewe's collection of papers on "The Periglacial Environment" (1969). Washburn's more recent "Periglacial Processes and Environments" (1973) devotes a full chapter to fluvial action, but the chapter occupies only six of over three hundred pages, and two pages of the six consist of photographs. A review of the relevant literature appears later in this section, but its small size should be mentioned here because it reflects a prevailing though ill-defined impression that, since most streams in periglacial areas only flow for a few months of the year, it is unlikely that they are capable of working so as to change significantly the form of the landscape.

Geomorphic writings on the periglacial environment, then, have often emphasized the cryogenic processes peculiar to cold climatic regions, at the expense of

processes which are familiar from lower latitudes. Much of what has been written is speculative, being based on sparse numerical data or none at all. Here, as with the hydrologic goals of the study, the first aim was to build a sound data set on which a more confident statement about the geomorphic role of fluvial activity could be based. Funds permitted only one summer of intensive data collection, and although this is scarcely adequate it is at least possible to place the results in context by comparing them with results from other climatic regions.

It transpires that the familiar processes of stream and wave action are responsible for most of the geomorphic work done in the High Arctic, at least if the definiton of "work" is restricted to the transfer of sediment from above base level (sea level) to below it. This statement can be seen most readily to be true when the landscape is viewed at scales which encompass areal units of the size of drainage basins or larger (Fig. 1:1). In units of the order in size of single hillslopes, mass movement under the influence of cold in the ground becomes discernible, although the role of water in this mass movement is also important (Wilkinson, 1972). It is not until the

Figure 1:1. A photographic mosaic of southwest Devon Island

(courtesy of P. J. Howarth)

scale of a few square metres is reached that cryogenic landforms come to dominate the frame of view (Fig. 1:2). While this periglacial "embroidery" (Carson and Kirkby, 1971, p. 325) is undoubtedly a challenge to analysis, and presents severe engineering problems, precisely the same is true of fluvial landforms and processes.

A particular question which illustrates the need for a corrective to the conventional view of the periglacial environment deals with carbonate solution. Bedrock in much of the High Arctic is limestone or dolomite; in particular, most of the work for this study was done in carbonate terrains. The question for consideration is, How do temperature, and climate generally, control rates and amounts of bedrock removal by solution? Until recently the opinion has been held that solution is more intense at higher latitudes: Williams (1949) argued that snowbanks should be especially favoured sites for chemical weathering, and Corbel (1959) made a vigorous case for increasing rates of solution in colder climates. The contrary view can, however, now be considered well established, and this study has helped in establishing it.

It is appropriate here to give a short description of the history of this study. As indicated, it

Figure 1:2. Frost-sorted stone circles

sprang from a desire to demonstrate the geomorphic significance of running water in the High Arctic, a desire which grew in turn from the increasing familiarity of McMaster University workers (in particular, the supervisor of the thesis) with the physical environment of southwest Devon Island. In the beginning, a summer of field research was conducted there in 1970 on the basin of "Jason's Creek", a small stream close to the base camp on Radstock Bay (Fig. 1:3). A feasibility study was also performed in that summer on the Mecham River near Resolute. Cornwallis Island, which was to become the main field site for research on the project (Cogley, 1971, 1972; Cogley and McCann, 1974, 1975; McCann and Cogley, 1971, 1972, 1973; McCann, Howarth and Cogley, 1972). In 1971 the Mecham River Project formally came into being as a Canadian International Hydrological Decade Project, having the number R-SR-21-N-11 and the title "Surface runoff of snowmelt and rainwater in the basin of the Mecham River, Cornwallis Island, N.W.T." A full programme of research was conducted on the Mecham River in summer 1971, and parts of the field programme were continued in 1972. The discharge of the stream has been measured each summer from then until the present (1975), although analyses have only been carried as far as 1974.

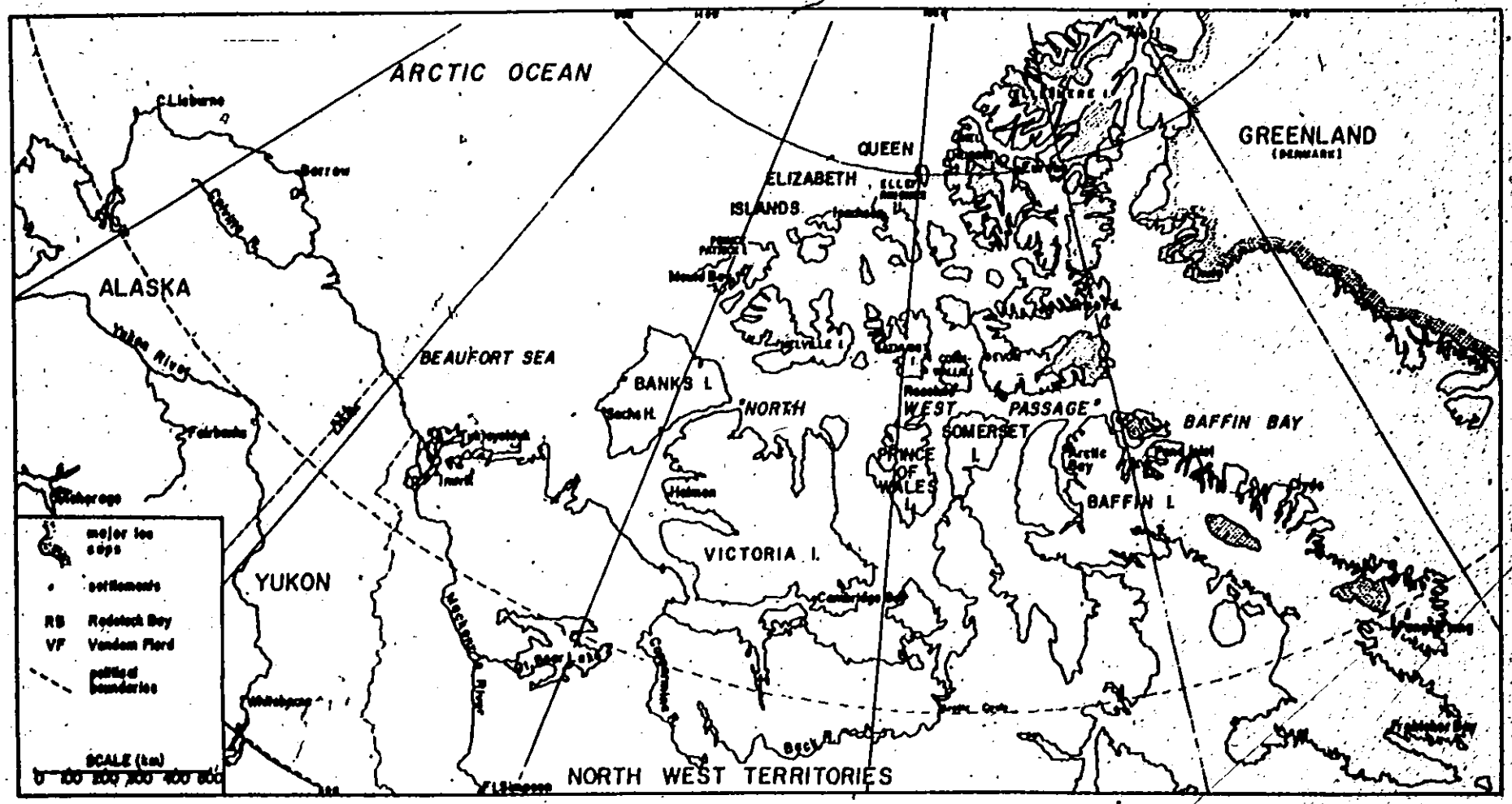


Figure 1:3. Location map for places in the High Arctic.

With the passage of time interest shifted from a principal concern with the second major goal to a principal concern with the first, until the proportionate emphasis on the two reached roughly the distribution found in this dissertation; the destination reached was thus not the same as the destination set out for. With time it was also realized that the present and near future held in store a number of practical problems to do with development of the High Arctic. While answers to specific problems of this nature have not been found (or sought), possible applications of the results of the research have been kept in view during the evolution of the project.

Most of the evidence on which statements have been based is drawn from research on "Jason's Creek", southwest Devon Island, the Mecham River, and streams near or adjacent to the Mecham River. After 1972, however, periods of time were spent doing research on the hydrology and fluvial geomorphology of southcentral Ellesmere Island ("VF" in Fig. 1:3), an area which lies in the path of a possible pipeline to carry oil or natural gas from the fields of the Sverdrup Basin. While few of the results of this research have been incorporated into this study, the statements made herein

are coloured by familiarity with conditions in south-central Ellesmere Island (an area contrasting with Cornwallis Island in several respects), and by brief acquaintance with conditions at a number of settlements elsewhere in the Canadian High Arctic and in western Greenland. In what follows, the context will usually give the degree of generality to be attached to any particular remark. Where this is not the case, a remark should be considered to apply strictly to the places and periods at which the data were collected. The basin of the Mecham River can be shown, where data are available for comparison, to be representative in many ways of other parts of the High Arctic. But undoubtedly there are substantial differences remaining to be discovered or confirmed between parts of the High Arctic.

1.1.3 Format of the Study The dissertation is divided into five sections: an introduction and a conclusion, which enclose sections on the hydrologic and geomorphic concerns which prompted the study. Although the first impetus was curiosity about geomorphic processes, the hydrologic results are presented first because they came to be seen as an essential background for apprecia-

tion of the geomorphic results. As related above, problems with the annual water balance assumed greater prominence during the evolution of the project; the bulk of the work done on these problems merits a section to itself, so that the second and third sections both deal with hydrology and the fourth with geomorphology.

In this introductory section some necessary preliminaries are given for a fuller understanding of the remainder of the text. The reasons for beginning the study, and something of how it developed, have been given in the preceding subsection. In the next subsection, although the bulk of the text is concerned only with findings from a circumscribed area near Resolute, there follows a summary description of the High Arctic. This is to put Resolute and the Mecham River basin in their regional context and to provide a background for some of the extrapolations which are made later. The next two subsections are bibliographical, dealing one with previous research on the subject of the dissertation, the other with relevant published sources of information. The rest of the introduction is concerned with how measurements were made in the field and with what data are collected at the Resolute weather station.

Section 2 deals with the surface runoff regime of the Mecham River. To explain this regime adequately it is necessary to consider the questions posed above in several categories. The most important aspects of the regime are the way in which runoff is first generated in spring and the course taken by the discharge curve as it rises to, and then falls from the snowmelt peak. These subjects are discussed in the second and third subsections, and are followed by discussion of the diurnal variability of discharge and responses to rainfall. Section 2 ends with a restatement of its main points, but the next to last subsection is important because in it is discussed the question of representativeness in time. Naturally the data series on which conclusions are based would be better for being longer, but the short hydrologic record can be placed in a valuable perspective by relating it to the much longer climatic record, which dates from 1948. Questions about representativeness in space are more difficult to answer, but answers are attempted in comments at several other places in the text.

Section 3 is devoted entirely to the annual water balance, and to the assessment of the terms in it. It begins with an explanation of the basic equation

which states the principle of the water balance. This equation is an expansion of the classical equation of hydrologic continuity, and its eight terms are discussed in the next two subsections. Windblown snow is given detailed attention in section 3.3.1, since the existing literature suggests that it may be an important flux of water in High Arctic basins. The term in the water balance which is treated at greatest length is, however, evaporation. It is the most difficult of the recognized terms to estimate, and a detailed defence is needed for the version of the "combination model" which is adopted here. The results of calculations of the vapour flux are presented in section 3.3.2.4; the method of calculation, using routinely collected weather data, has some intrinsic interest apart from the aims of this study, and in section 3.3.2.5 reason is shown for hoping that reliable evaporation estimates may be obtainable from published monthly summaries.

The other "large" terms in the balance are snowfall, rainfall and discharge. They are measured directly, and the direct measurements must be accepted for better or worse. The validity of the discharge data, collected specially for the study, is discussed first in section 1.2.1.3, but all the terms in the water balance are re-

examined in section 3.4. Each of its subsections is important for a just assessment of the conclusion which is drawn: that the High Arctic water balance involves considerably larger mass transfers than are indicated by available data on precipitation.

Section 4 takes up some of the geomorphic implications of this conclusion, although the framework of the section is an exposition of the data on stream load collected mainly during 1971. The basis for the framework is the convention of dividing a stream's sediment load into three fractions. With the information which was gathered little can be said of bedload, but the other fractions can be considered in more detail. To throw a different light on the results for the Mecham River, suspended sediment data are presented from two streams in southcentral Ellesmere Island, both of which have glacerized catchments. Since it is notoriously difficult in geomorphology to make sound inferences from short samples to long periods, the significance of these data from High Arctic streams can only be assessed comparatively and by implication. This is done in section 4.2.3. A similar remark applies to dissolved load, although the sources here are more varied. If solute concentrations, as

opposed to loads, are used, the waters of the Mecham River basin can be compared with waters sampled in several investigations in other parts of the world. The results of the comparison have their own intrinsic interest, apart from an attempt to estimate stream load; they are given in section 4.2.3.2 and elaborated upon in section 4.2.3.3.

Section 5 has three purposes: to restate as succinctly as possible the findings of the study; to dwell briefly on some of the implications of these findings; and to recommend steps which should be taken to deal with the problems which have been elaborated or identified.

1.1.4 A Geographical Description of the High Arctic

The "High Arctic" is defined to be the Queen Elizabeth Islands together with a number of islands to the south of Parry Channel (the "Northwest Passage"): Banks Island, Victoria Island, Prince of Wales Island, Somerset Island and Bylot Island, together with (at least) the northern part of Baffin Island.

1.1.4.1 Climate The near-polar situation of the High Arctic is responsible for a climate which is everywhere

cold and dry. Above the Arctic Circle solar energy is received only during a portion of the year. At 75° N, the sun is below the horizon from early November to early February and above the horizon from late April to mid-August. The sun is never high in the sky, and although the total of available solar energy is approximately the same in summer as at lower latitudes, little of the available total is used to heat the surface and the lower atmosphere. High reflectivities from cloud surfaces, sea ice cover and ground snow cover typify the region even in summer, and winter is long and extremely cold.

Synoptically, the meteorological year may be said to begin during September and October with the development of an anticyclone over the Beaufort Sea in the west. Cold, anticyclonic air flowing over relatively warm surfaces is apt to become unstable; September and October are stormy months, and considerable amounts of snow fall at this time. As winter progresses the anticyclone becomes stronger and expands to cover most of the High Arctic. A well-developed area of low pressure over Baffin Bay tends to weaken through the winter, but nevertheless a circulation from NW to SE is maintained over the High Arctic until springtime. Precipitation is light after the establishment of the anticyclone, although this is less the case in eastern Baffin Island

than elsewhere. The high mountains of Baffin Island act as a barrier to moist cyclonic systems moving northward from the Atlantic, and these systems deposit precipitation on the mountains and the eastern coast throughout the year.

In May or early June pressure drops over the entire area, and cyclones migrate into the High Arctic from the west and southwest with increasing frequency. Some inward advection of heat is associated with the cyclones of spring and summer, but this phenomenon is less marked in the High Arctic than in "mid-Arctic" areas such as the Mackenzie Delta. The major effect of cyclonic migration is to bring precipitation. Almost everywhere July and August are the wettest months of the year (Fig. 1:4) and since temperatures usually rise above freezing point under the influence of solar heating at this period, most of the summer precipitation falls as rain.

As a consequence of the radiative energy distribution, temperatures are well below zero throughout the year until May, and only in July and August are temperatures consistently above zero for any length of time. By early September temperatures are again below the freezing point, and the annual round begins anew.

The climatological summary above is drawn mainly from "Climate of the Canadian Arctic" (Meteorological

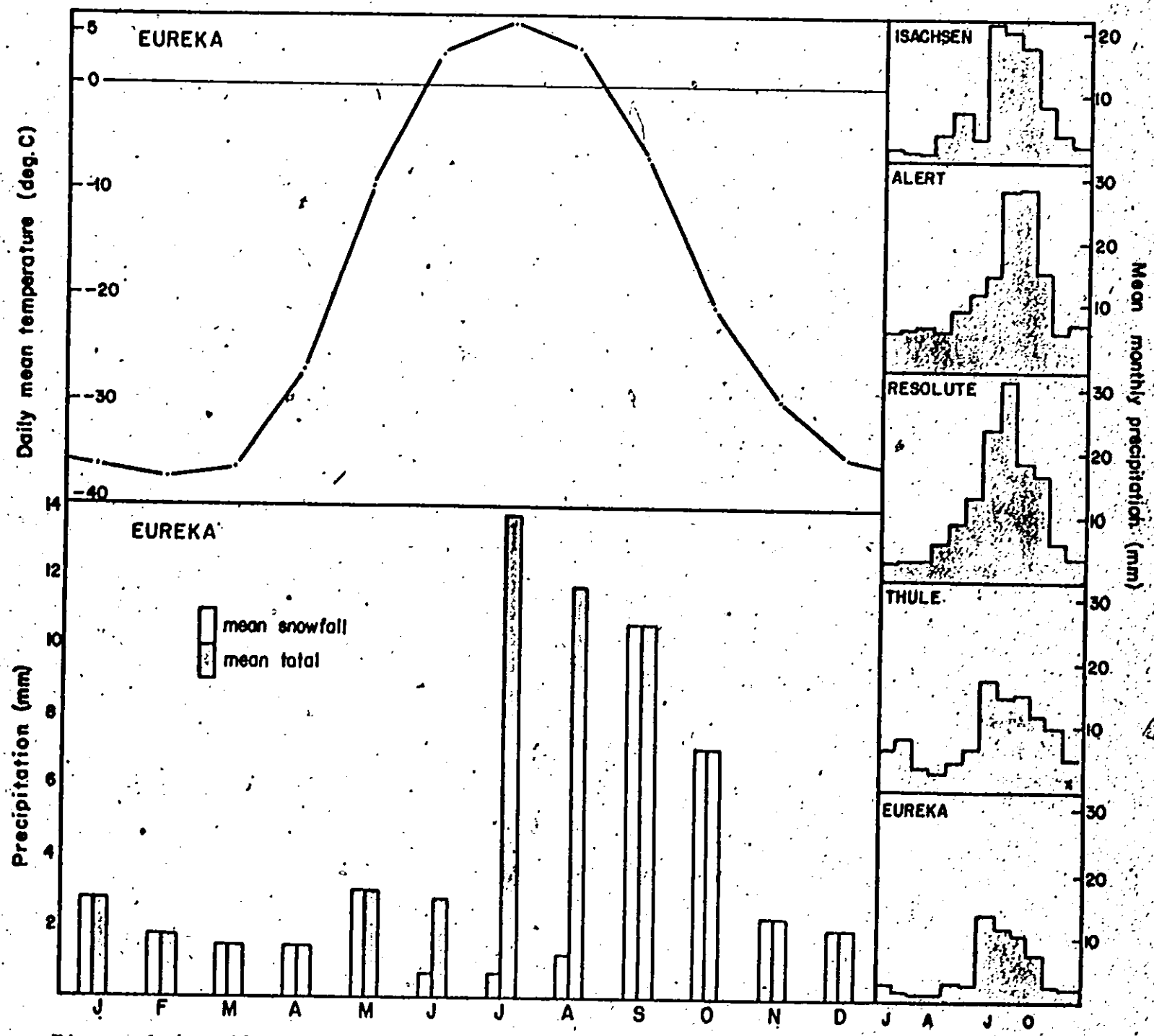


Figure 1:4. Climatological summary graphs for Eureka and five other High Arctic stations.

Branch, 1970a); up-to-date climatic normals are published in "Temperature and Precipitation: The North" (Atmospheric Environment Service, 1971). However, for the most important hydrometeorological variable, precipitation, no convenient or up-to-date summary is available, at least on an annual basis. Figure 1:5 has been prepared to remedy this situation. It is based on reliable and readily accessible data, and it is believed to be the most accurate map of measured annual precipitation over the High Arctic which has been prepared to date. No account is taken of orographic influences on the arrangement of the isohyets, so the map is inaccurate in areas such as eastern Baffin Island and perhaps the higher mountains of Ellesmere Island. The important features of spatial variation in precipitation are, however, clearly brought out. The dominant feature of the map is a progressive decrease in amount of precipitation from south to north, reflecting a transition from the mid-latitude zone of low pressures to the stable polar high pressure region. Eureka is the driest meteorological station in Canada, with a mean annual total of measured precipitation of only 58.4 mm. In the western Arctic, a feature of the precipitation distribution is the steep gradient indicated by convergence of isohyets across the mainland

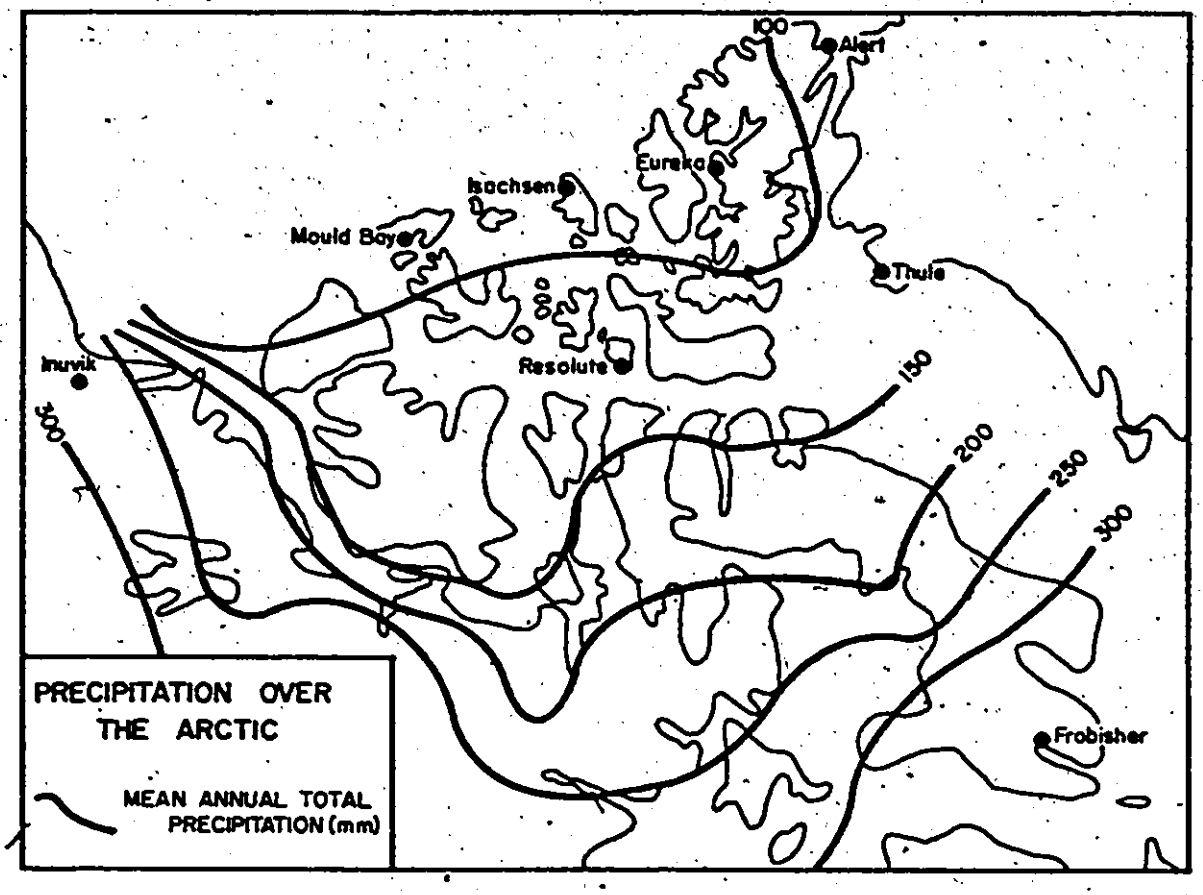


Figure 1:5. Precipitation over the High Arctic.

coast. This gradient is well documented, for a number of D.E.W. line stations were located in the area, but it is difficult to explain; it is not known whether the southward salient of the 150 mm isohyet over Victoria Island reflects an actual phenomenon or merely the absence of data from the north of that island.

In general, the most important facts about Arctic precipitation are that it decreases to the north and that orographic effects influence its distribution in the eastern islands.

1.1.4.2 Physiography The sediments from which the High Arctic is formed have been laid down at intervals throughout geological time (Fortier et al., 1963, Thorsteinsson and Tozer, 1960). Broadly speaking, their age decreases from SE to NW; shield rocks appear in Baffin Island, part of Somerset Island, eastern Devon Island and southeastern Ellesmere Island. Abutting on this basement are Palaeozoic strata, which appear at the surface over wide areas in northern Baffin Island, Somerset, Devon, Ellesmere and Cornwallis Islands. In the outer islands such as Ellef Ringnes Island Mesozoic outcrops predominate, and through these younger strata of the Sverdrup Basin evaporites have intruded as diapirs. The evaporite beds and intrusions are important traps for natural hydrocarbons, which may

be exploited in the near future. Tertiary sediments, particularly of Eocene age, are widely distributed in the High Arctic, many being of very-shallow-water or terrestrial origin, and in the extreme west are found extensive deposits laid down during the Quaternary or late Tertiary.

The Laramide orogeny of the middle Tertiary (Kerr, 1960) affected the High Arctic by folding or re-folding the beds of its eastern margin into high mountain ranges, which persist today and form the "backbones" of Baffin, Devon and Ellesmere Islands, and also of Axel Heiberg Island. Both altitude and relief decrease to the west. Extensive plateaux, but little dissected, characterize western Devon Island and Cornwallis Island, while further to the west the topography is lower still and very subdued. The westward decrease of altitude may be thought of as a ghost of the "Barrow Surface", a hypothetical plane of gentle slope which developed in the High Arctic following the main phase of Laramide activity. The extensive Quaternary and late Tertiary deposits of the western Arctic, the Beaufort Formation, are products of the development of the Barrow Surface.

Probably the most important single event in the recent geological history of the High Arctic was glaciation. The events preceding glaciation may be

placed, very generally, as follows, on a time scale beginning arbitrarily in the early Tertiary:

$5\text{-}4\times10^7$ yB.P.:	partial marine transgression, and widespread deposition of sediments which remain, as a rule, weakly indurated and incompetent;
$4\text{-}3\times10^7$ yB.P.:	Laramide folding and uplift of the eastern mountain ranges;
$3.5\text{-}0.5\times10^7$ yB.P.:	development of the Barrow Surface, possibly followed by dissection and growth of a topographic system which now finds expression in the pattern of islands and seaways forming the Arctic Archipelago (Fortier and Morley, 1956).

The glacial history of the High Arctic is not well known. There is abundant evidence that most of the region was covered by glacier ice at least once during the Pleistocene, and until recently it was held that most of the islands supported local ice caps. Blake (1970), however, has argued that the Queen Elizabeth Islands were glaciated by an "Innuitian Ice Cap" centred on the Norwegian Sea, roughly in the middle of that island group. At present the largest ice caps in North

America are to be found on the highlands of eastern Devon and Ellesmere Islands, and these, together with an ice cap or caps centred on Baffin Island, were presumably more extensive at the glacial maximum. The Laurentide Ice Sheet, at its maximum extent, reached southern Somerset Island.

The evidence for former glaciation includes mantles of till and other sediment of glacial origin, the carving or enlargement of valleys and, in some respects most important of all, continuing rebound of a land mass which was depressed by the weight of overlying ice, to depths up to 150 m below present levels with respect to the sea. Return to isostatic equilibrium continues today, although it was substantially completed some 5000 years ago. In the area of southwest Devon and southeast Cornwallis Islands, for example, the ice disappeared 8-9000 years ago, and shortly after deglaciation sea level was 130 m higher with respect to the land than it is now. By 5000 years ago the sea was only 20 m above its present level. Isostatic rebound has been marked by the development of suites of raised beaches and strandlines in favourable locations, and these sequences have been incised by streams whose origins probably date back no further than

the end of glaciation. An example of such a stream is "Jason's Creek", whose valley is cut roughly perpendicularly across the raised beaches flanking Radstock Bay.

During the Holocene, the geomorphic evolution of High Arctic land surfaces appears to have proceeded in ways not differing greatly from those observed today. Glacial conditions were replaced some 9000 years ago by periglacial conditions, and periglacial landforms have been generated on surfaces subjected concurrently to the denudational activity of running water and other agents such as winds and waves.

The ice caps of Devon and Ellesmere Islands have been referred to above. They contribute significantly, on a continental scale, to the water balance of North America, and more particularly to the water balance of the High Arctic, by holding large amounts of fresh water in storage. Their effect on the hydrological and geomorphic regimes of the drainage basins to which they contribute is also substantial, and this fact must be borne in mind when making generalizations about the hydrology of the High Arctic. In particular, the Mecham River and "Jason's Creek" have unglacierized

catchments, and glacier-fed streams may be expected to differ from them in characteristics such as regime, precipitation response and sediment distribution and transport.

1.1.4.3 Permafrost All basins in the High Arctic are profoundly affected by the persistence of freezing temperatures in the ground. In summer the active layer at the surface reaches depths which rarely exceed 1 m, but throughout the greater part of the year temperatures are below zero from the surface down to at least 400-500 m (Brown, 1970, 1972). Permafrost is deep and continuous everywhere in the High Arctic, so that any groundwater which may be found beneath the permafrost is efficiently sealed off from hydrological proceedings at the surface. The surface or, to be more precise, the permafrost table is of course completely impermeable, and surface runoff is similar in a general sense to runoff from other impermeable terrains. The existence of the active layer in summer generally makes little difference to the capacity of a drainage basin to retain water, but the shallowness of the active layer implies that saturation of the surface is common. Lubrication of the active layer by an excess of water often leads to instability, and many of the mass movements which occur on slopes

result from this instability; the permafrost table forms a rigid substratum which serves as a locus for failures and a channel bed for flows.

The High Arctic contrasts with sub-arctic regions in the continuousness of its permafrost. In areas such as central Alaska, the Mackenzie Valley and northern Quebec, discontinuities in the perennially frozen ground and greater active layer depths combine to complicate the hydrological situation, by allowing groundwater to interact with surface water and by providing larger reservoirs for retention of water. Sub-Arctic localities are also characterized by more frequent development of segregated layers of ground ice and the thermokarstic phenomena which often result from ice segregation; moreover, the sub-Arctic happens to be a zone of low relief in the main, and impeded drainage, tracts of muskeg terrain and widespread superabundance of lakes are more common in the sub-Arctic, and especially the Canadian Shield, than in the High Arctic.

<u>1.1.4.4 Vegetation</u> All of the High Arctic is located well to the north of the tree line, and no plants grow to heights above about 0.2 m. Many areas are almost

completely barren, and climatic inhospitability keeps total biomass low even where supplies of minerals and other nutrients can be drawn upon readily by growing plants. On the calcareous terrains found, for example, over most of Cornwallis and Devon Islands, vegetation cover is especially sparse and the flora more than usually restricted. Terrains with different lithologies are more completely covered with vegetation, but for the entire Arctic archipelago Porsild (1957) was able to enumerate only 346 species of vascular plant.

1.1.4.5 Settlements and Communications The High Arctic has a resident population of 2800-3000, scattered among not more than twelve settlements, ranging in size from small meteorological stations housing fewer than 20 people to Eskimo communities of 600 or more. The regional "capital" is Resolute, which serves principally as an airline terminus and route centre for the air charter operations by which almost all travel within the High Arctic is conducted. Heavy freight is brought into the High Arctic by ship during summer and autumn. Overland communications are still in a very rudimentary state; there are no tarred roads anywhere in the region, and although

gravel and dirt tracks radiate from most settlements to distances of a few kilometres, these tracks do not lead to other settlements. The time, however, is not far distant when freight movements by road will become practicable and/or necessary, and already vehicle movements in connection with scientific research, resource exploration and exploitation are increasing annually. As the pace of High Arctic development quickens, increasing importance will attach to the safety of roads and other overland transportation systems (especially pipelines) which are subject to hydrological hazards.

Possibly of even greater importance are the demands which will be placed on water supply as settlements grow. In summer the population of the High Arctic increases substantially, and settlements which drain water from streams rather than from lakes can experience difficulties in late summer when streamflow decreases. In winter the absence or inaccessibility of supplies of unfrozen water can also cause difficulties, particularly in the less well-equipped (i.e. Eskimo) settlements.

1.1.4.6 Resolute and the Mecham River Basin Resolute is the most accessible place in the High Arctic. Partly

for this reason, a considerable amount of research, including the bulk of the work for this study, has been done there (Fig. 1:6). Comprehensive programmes of data collection have also been conducted at Resolute since the settlement was established in 1947.

The climate of Resolute is summarized in Figure 1:7 and Table 1:1. The table is from Atmospheric Environment Service, 1971; the figure from Meteorological Branch, 1970a. There are small differences in the two sources. Range of temperature is greatest in the winter months but temperatures are generally above the freezing point only from mid-June to late August; mean daily temperatures in February and July, respectively the coolest and warmest months, are -33.1° and 4.2°C, and the absolute extremes recorded at the station are 18.2° and -51.7°. Mean annual measured precipitation is 136 mm, with a seasonal distribution such that most falls in the summer half-year. On the average 43 per cent of annual precipitation is rain, falling almost entirely between mid-June and early September. The remainder of the precipitation falls as snow, and the seasonal snowfall distribution exhibits two maxima, a greater in September-October and a lesser in April-May.

The basin of the Mecham River and its environs are floored by predominantly calcareous sediments of

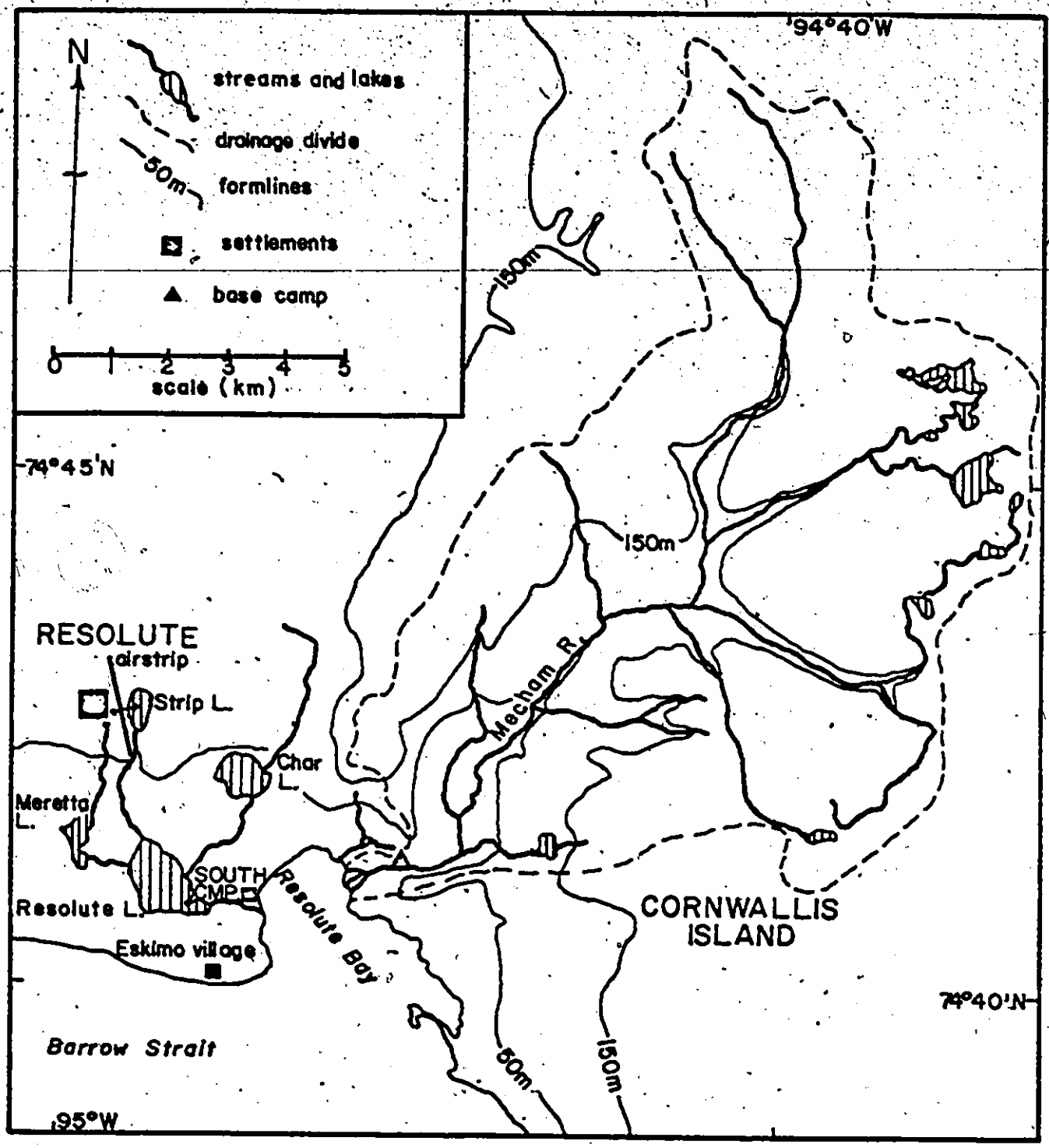


Figure 1:6. Location map for the Resolute area.

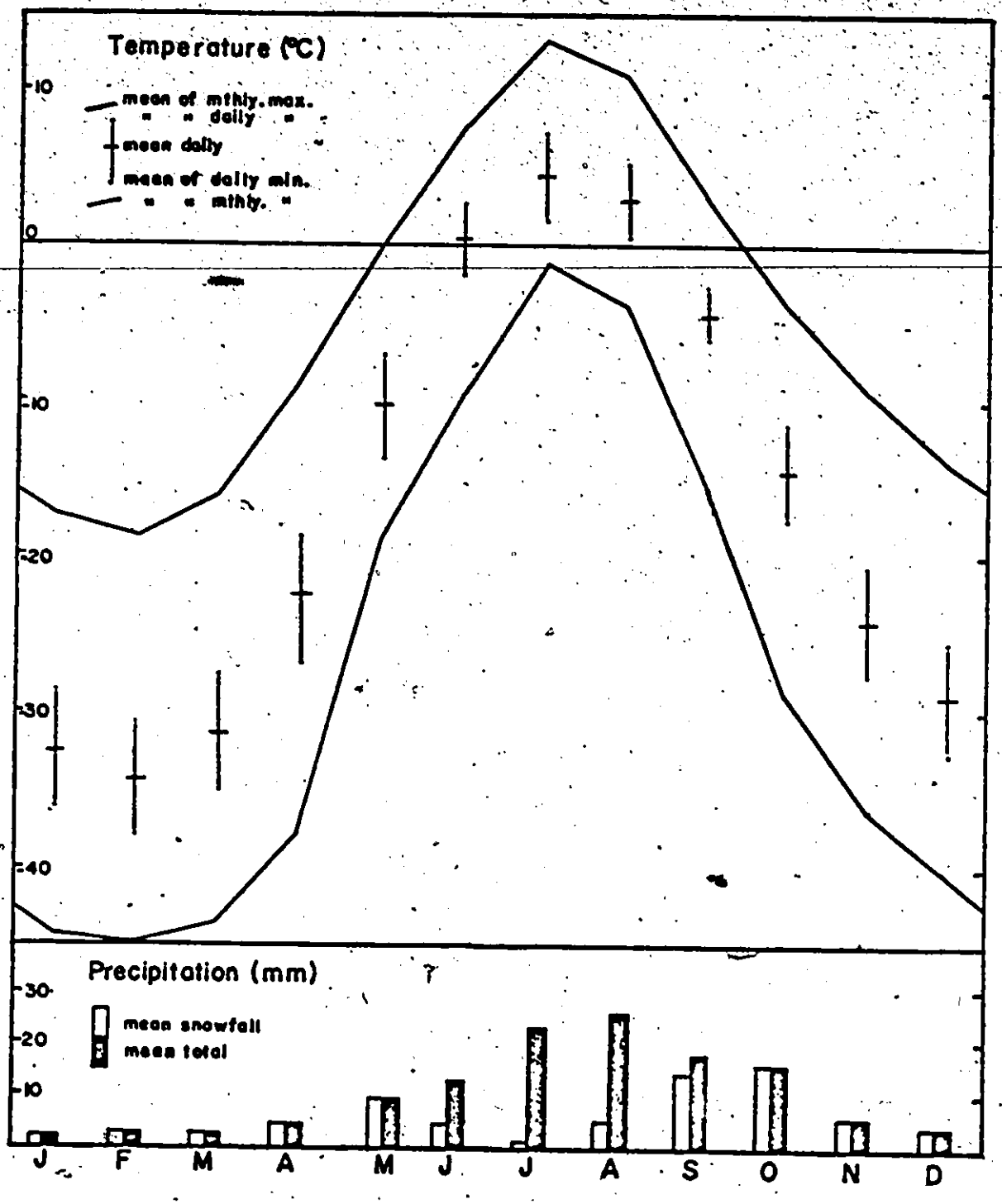


Figure 1:7. Climatological summary graph for Resolute, 1948-60.

TABLE 1:1

Climatic Means at Resolute, 1948-70

Month	Daily mean temperature (degrees C)	Mean total rainfall (mm)	Mean total snowfall (mm)	Mean total precipitation (mm)
January	-32.2	0	27.9	2.8
February	-33.1	0	33.0	3.3
March	-31.0	0	33.0	3.0
April	-22.8	0	58.4	5.8
May	-10.5	0	88.9	8.6
June	- 0.3	5.8	66.0	12.4
July	4.2	23.4	30.5	26.4
August	2.7	25.6	48.3	30.5
September	- 4.8	3.8	142.2	17.8
October	-14.6	0	155.0	15.2
November	-24.0	0	55.9	5.6
December	-27.4	0	48.3	4.8
Annual mean or total	-16.2	58.7	787.4	136.4

Upper Ordovician to Lower Devonian age. The two formations found in the basin (Fig. 1:8) are the Allen Bay and Read Bay Formations (Thorsteinsson, 1958; Thorsteinsson and Kerr, 1967). The older Allen Bay Formation occupies small areas near the mouth and in the extreme north of the basin; it consists mainly of dolomite, thin- to thick-bedded, dense to porous, with minor dolomitic limestone, limestone, argillaceous limestone and calcareous shale. The contact of the Allen Bay rocks with the overlying and more extensively exposed Read Bay Formation is transitional, but the Read Bay beds are mainly grey, brown or yellow limestone with lesser thicknesses of argillaceous limestone, dolomite and calcareous shale, which is locally bituminous or petroliferous, and minor sandstone and siltstone.

The strata into which the Mecham River is cutting are gently folded, the lower reaches of the river flowing across the axis of the Resolute Bay Anticline, to the east of which lie the Barrow Strait Syncline and the Assistance Anticline. In the north of the basin the dip is generally westward towards the Resolute Bay Syncline. Robitaille (1960) considered the present Mecham River to be the descendant of a

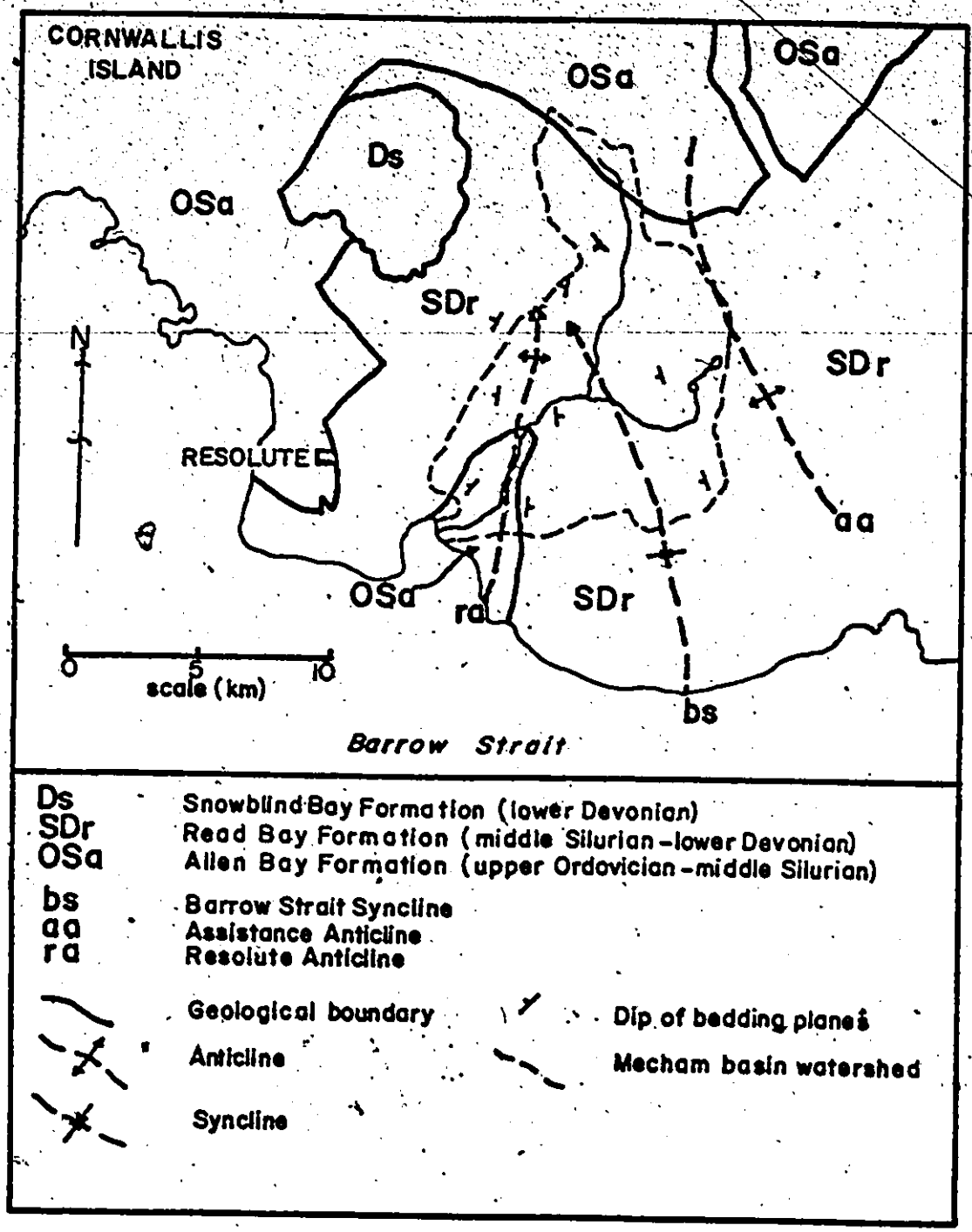


Figure 1:8. Geology of the Mecham River basin.

stream which first flowed before glaciation, probably in the Pliocene. At this time dissection of the Barrow Surface was beginning, and Robitaille's conclusion seems a reasonable one, particularly when the Mecham River and other streams like it on Cornwallis Island are contrasted with streams such as "Jason's Creek" on Devon Island. The inward curvature of raised strandlines near to the mouth of the Mecham River, and the considerable amount by which the stream has widened its valley, both suggest that erosion of the valley was resumed rather than initiated during the Holocene. There is convincing evidence, in the form of mollusc shells resting in the position of death, for believing that the Mecham valley was a marine inlet at least 8 km long at the time of deglaciation, when sea level was near to its maximum against the land.

The valley of the Mecham River is carved into a plateau with an average elevation of 150-180 m a.s.l. (above sea level), the highest point in the basin being 194 m a.s.l. (Fig. 1:9). The drainage area tributary to the mouth of the river is 97.7 km^2. Overall relief is less pronounced than on the plateaux of Devon Island where, for example, "Jason's Creek" drains an area of 2.26 km^2 rising to an altitude of 320 m a.s.l. The valley of "Jason's Creek" is steeply

Figure 1:9. The Mecham River basin from the air.

Figure 1:10. The lower valley.

incised and reaches depths of 150 m below its immediately surrounding plateau, but no comparable gradients are found in the more subdued topography of the Mecham basin (Fig. 1:10). Some tributaries of the Mecham River are indeed sharply incised, as might be expected during the early stages of dissection of a plateau surface. However, since base level is never far away in a vertical direction, depths of incision are always small in comparison with those observed in the higher plateaux to the west.

Terrains within the Mecham basin vary from coarse gravel tracts in the neighbourhood of the river (Fig. 1:11) to areas of fine-grained, silt and clay soils on the plateau and on gentle valley slopes. Frequently these finer soils have admixtures of coarser material which is segregated into nets, stripes and other patterns by frost sorting. In the language of the pedologist, the surface of the basin is covered with lithosols and polar desert soils, with tundra gleys and bog soils where drainage is poor (Cruickshank, 1971). There has, however, been very little time for the slow processes of pedogenesis to produce mature soils in the accepted sense of that word. Material of former littoral origin is predominantly of gravel size (Fig. 1:12), though there are sandy deposits, which originated

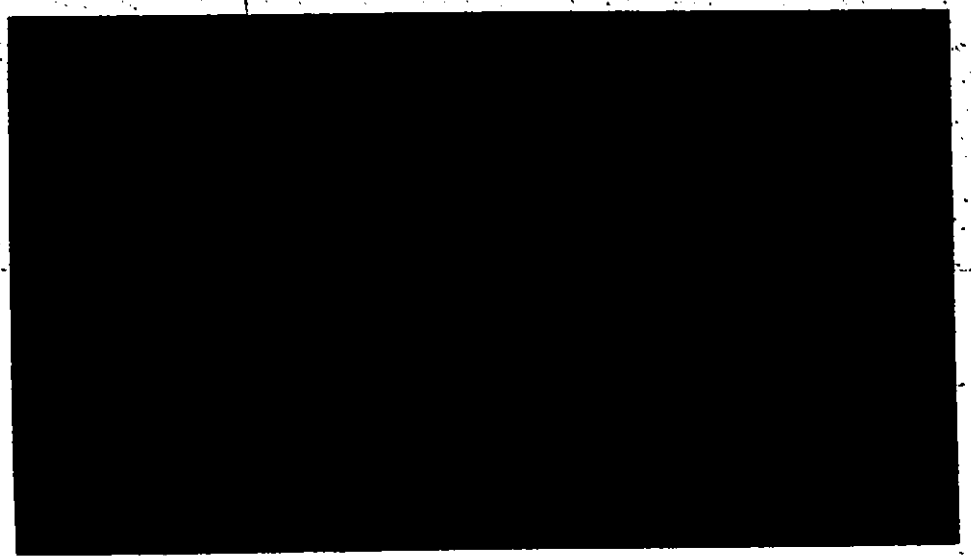


Figure 1:11. Gravel terrain near to the river.

Figure 1:12. Coarse material on a slope subjected to frost sorting.

under probable lacustrine conditions, in the lower parts of the valley. Bedrock outcrops usually give rise to sheets and cones of very coarse talus on steeper slopes, and to blockfields on the plateau; this very coarse material tends to be angular and platy, particularly where bedding is thin, and individual "grains" may reach lengths of 1 m and more. Most of the plateau, however, is covered by a generally fine regolith of unknown depth, which may be in part glacial till and in part the product of weathering in situ.

The basin as a whole is nearly barren of vegetation. Individual plants are distributed sporadically, but continuous vegetation is confined to small areas of impeded drainage, for example at the bases of talus cones. These areas probably cover one or two per cent of the drainage basin. Arkay (1972) studied primary production of carbon by the terrestrial flora of the Char Lake drainage basin, adjacent to that of the Mecham River. She estimated primary shoot production at the very low figure 0.09 kg C m^{-2} y^{-1}, and found a vascular flora comprising only 45 species. The most common species in the Mecham basin are Salix arctica (arctic willow), Saxifraga oppositifolia (purple saxifrage), Dryas integrifolia (mountain avens) and a number of grasses and sedges.

1.1.5 Previous Research on Arctic Hydrology Hydrological research in the High Arctic has been limited until recently. The behaviour of water in sub-Arctic environments such as those of central Alaska, the Mackenzie Delta and Labrador-Ungava is better, though still not thoroughly, documented. In the High Arctic itself, more attention has been focussed on glacierized than on unglacierized basins.

A review of the published literature on Arctic hydrology formed part of Chapter 1 of Cogley (1971). Some of that material is repeated here in summary, together with notes on work which has appeared or been uncovered since then.

Two articles had been devoted, before 1971, to surface runoff in the High Arctic. Writers such as St.-Onge (1965) appreciated its importance, but the first to study it in its own right was Pissart (1967), who dealt with conditions at Mould Bay, Prince Patrick Island, and Cook (1967), who wrote a description, with limited supporting data, of the annual regime of the Mecham River. Cook's description is good. He divided the hydrological year into five parts, beginning with a long period of snow accumulation and no discharge in winter. In June a brief span occurs during which the snowpack "ripens" or

becomes saturated with water; this span is followed by a still shorter one which is defined by the start of open channel flow, a gradual increase in discharge through the snow-lined channels, and the onset of the flood. The annual snowmelt flood is the most imposing feature of the annual hydrograph, for although it lasts only ten to twenty days it carries three quarters or more of the year's discharge. It is succeeded by the final period, lasting from some time in early July to the end of August or the beginning of September, during which there is a gradual decrease in discharge. The decrease in discharge reflects a decrease in the amount of snow left in the basin, and it is liable to be interrupted by floods resulting from summer rainfall; the floods are sharply peaked because the basin is effectively impermeable and lacks a vegetation cover. The end of the hydrological year comes when water remaining in the channel freezes down to the bed, and snow accumulation begins again.

Streams in Spitzbergen have been described by Vivian (1964) and Czeppe (1965). Spitzbergen receives more precipitation than does the High Arctic, but its hydrological regime seems to be similar in outline. In Alaska, Dingman (1966) and Likes (1966) have studied drainage basins, near Fairbanks and Cape

Lisburne respectively, which were vegetated and which lacked continuous permafrost. Rapidity of basin response is slower under these conditions than on the barren permafrost terrains of the High Arctic. The same holds true for the basin at Barrow studied by Brown, Dingman and Lewellen (1968); the permafrost beneath it is continuous, but small lakes and ponds on the nearby flat surface of the basin delay considerably the transmission of flood waves through it. Slaughter (1971) describes a recent extension, to a larger basin, of Dingman's work near Fairbanks. Studies of larger rivers in northern Alaska are those of Schallock, Mueller and Gordon (1970), who give a preliminary analysis of runoff hydrology and water quality for the Chena and Sagavanirktok Rivers, and Arnborg, Walker and Peippo (1966) and Walker (1973), whose work deals with the Colville, a river draining more than 50,000 km^2. The same regime prevails in the Colville as described by Cook (1967) for the Mecham, with the important exception that the river does not freeze completely in winter but maintains sluggish flow beneath a surface ice layer. This is a characteristic of large, northward-flowing streams whose headwaters lie in more temperate regions than do their lower reaches. The Yukon, and more clearly the Mackenzie, have this property, but in the case of the Colville

River the persistence of an unfrozen channel can be related simply to the size of the channel. It is not known whether any streams in the High Arctic are large enough to remain unfrozen all year.

Streams which drain glaciers, or which receive significant contributions from glaciers, have been studied in Antarctica (Davis and Nichols, 1968), Iceland (Gudmundsson and Sigbjarnasson, 1970, 1972), northern Sweden (Stenborg, 1970), southern Norway (Østrem, 1970), Baffin Island (Østrem, Bridge and Rannie, 1967) and northern Alaska (Wendler and Trabant, 1972), as well as in the High Arctic itself. Diurnal fluctuations in discharge are noted by all of these authors, the fluctuations being in response to diurnal fluctuations in the energy sources used for melting ice and snow. The main energy source is solar radiation, normally with a lesser contribution from sensible heat fluxes; Østrem noted also high correlations of wind speed and humidity with glacier discharge, suggesting that condensation might play a significant role in the budgets of some Norwegian glaciers. Glaciers are, in general, more complicated phenomena than ephemeral snowpacks. Stenborg, for example, found that the Mikka Glacier required periods of the order of weeks rather than

days or hours to transmit gross input fluctuations to its outlet.

In the High Arctic itself, glacio-hydrological studies have been conducted by Keeler (1964), Adams (1966) and Maag (1969). Maag's work is concerned with the behaviour of ice-dammed lakes; such lakes often have the habit of emptying catastrophically at regular or irregular intervals, with potentially disastrous results. The studies made by Keeler of the Sverdrup Glacier, Devon Island, and by Adams of the White Glacier, Axel Heiberg Island, are of wider scope. Keeler established that the largest heat source for melting was radiation, the two other principal sources being sensible and latent heat fluxes. Interestingly, the latent heat flux was directed downward almost all the time; in other words condensation, and not evaporation, was the rule. Heat conduction into the ice mass accounted for 12 per cent of heat supplied to the surface, and the remainder was assumed to have been used in melting.

Adams' study was oriented more towards hydrological observation than was Keeler's meteorological work. He provided a detailed description, based on a number of continuous data series, of ablation on and discharge from a glacier system in which ice-dammed

lakes provide sizeable storage facilities for water flowing off the glacier surfaces. This factor contributes to a lowering of the spring discharge peak in favour of a higher and longer hydrograph spread over much of the summer. The main reason for this more even distribution of discharge, however, is that a glacier surface does not "dry up", as does a land surface with a thin cover of snow, when it is heated above the freezing point. Glacier melt continues unchecked by diminution in the area covered by snow or ice.

Holmgren (1971) has produced an extensive study of ice-cap climatology for the Devon Ice Cap, noting among other things small amounts of net evaporation in summer. The meteorology of the same area studied by Adams and Maag, namely west central Axel Heiberg Island, is described by Ohmura (1972 a, b, c). The energy exchanges at vegetated tundra surfaces are such that most of the heat supplied by radiation is used to warm the air. About 60 per cent is used in this way, 30 per cent in evaporation and 10 per cent in warming the active layer. On a seasonal scale, heat consumed in melting the snow cover is a small fraction of the total which is received.

It should be noted that, by analogy with the iceberg, the papers on glacial hydrology and hydrometeorology reviewed above are simply the more pertinent surface layer of a large body of glaciological research. The remainder of this large body has been passed over only because its direct relevance to the study of surface runoff from unglacierized terrain is limited.

The study of fluvial landforms and fluvial sediments, both in situ and in motion, is again further advanced at lower latitudes than in the High Arctic. In Alaska, Brown, Grant, Ugoloni and Tedrow (1962) measured concentrations of a number of solute species in a variety of localities in northern Alaska, concluding that these concentrations were, in general, comparable with those observed in comparable lithological situations further south. Rainwater and Guy (1961) reached a similar conclusion for the Chamberlin Glacier in northeast Alaska. Arnborg, Walker and Peippo (1967) presented data on the suspended sediment concentration and load of the Colville River, noting a concentration of 1658 parts per million (ppm) at a flood discharge of 5050 m^3s^{-1}; most of the annual load was carried during the snowmelt flood.

Investigations of sediment transport by high latitude and/or glacial streams outside Alaska are few; those of Østrem, Bridge and Rannie (1967) on Baffin Island, and of the Norges Vassdrags og Elektrisitetsvesen (Østrem, Ziegler and Ekeman, 1970; Ziegler, ed., 1972, 1973, 1974) in Norway are noteworthy. Active glaciers supply large quantities of detritus to streams issuing from them, and their loads are commonly high, and certainly higher than those of streams with comparable discharges and gradients which are not nourished by glaciers.

Fluvial landforms in arctic regions have been studied by G. S. Anderson and Hussey (1962) and Legget, Brown and Johnston (1966), who dealt with alluvial fan formation, the former in the Sagavanirktok basin, Alaska and the latter near Aklavik in the Mackenzie Delta. Other studies of fluvial geomorphology in the Arctic part of the Mackenzie basin include those of Henoch (1960) and J. R. Mackay (1963, 1970, 1972).

The most important work on the hydrology and fluvial geomorphology of glacial outwash streams was written by Church (1970). This monumental dissertation, through its published

version (Church, 1972), makes important contributions to fluvial sedimentation, glacial geomorphology, Quaternary and particularly Holocene geology, and possibly other subjects. Church shows that rate of alluvial activity is a function of the presence or absence of active glaciers upstream, and that where glaciers are absent or unimportant rate of alluvial activity decreases with time since deglaciation. He errs in concluding that "the characteristic by which Arctic fluvial activity is distinguished from that of all other regions" is the small role played by dissolution in comparison with removal of clastic material. Nevertheless, his work is and is likely to remain the standard reference on the fluvial landforms and processes associated with discharge of water and sediment from glaciers.

Interest has quickened recently in the hydrology of the North American Arctic, particularly because of the prospect of development in the region. Two large volumes containing reports on hydrologic studies in the Mackenzie basin have been published by the Glaciology Division of Environment Canada (1973, 1974). Most of the papers are preliminary or progress reports on work which ranges from ice-

jamming on the Mackenzie River to tree-ring analysis of historical floods and oil pollution on ice-covered rivers. The most relevant to this thesis are papers by J. C. Anderson and D. K. MacKay (1973,1974) on small basins in the Mackenzie Delta and by Jasper (1973,1974) on a small basin in the Mackenzie Mountains; Anderson's study (1973) of the water balance of large tributary basins should also be mentioned. The stream studied by Jasper lies geomorphically in a more energetic environment than does the Mecham River, and appears to transport more sediment; summer rainstorms appear to be more significant components of its hydrologic regime, as would be expected of a stream much further south than the Mecham River. The delta streams of Anderson and MacKay resemble the Mecham River more closely and preliminary results indicate that they too have water budgets which cannot be balanced using existing measurements of precipitation.

Another sign of increased interest in Arctic hydrology is the appearance of two review articles. MacKay and Løken (1974) survey the field in a wide sense, summarizing Russian work and discussing major rivers and lakes, river and lake ice conditions,

groundwater and water resource management. Church (1974) has provided a schema for classifying the different hydrologic regimes of northern North America, distinguishing proglacial and muskeg regimes and nival regimes which he subdivides into subarctic and arctic variants. "Jason's Creek" (cf. Cogley, 1971, and McCann and Cogley, 1972) is taken as one type of the arctic nival regime, and the Mecham River may also be considered representative of it. The volume in which Church's paper appears (Can. Natl. Comm. Intl. Hydrol. Decade, 1974) also contains a number of other articles on the subject "permafrost hydrology", and in particular one by Ambler (1974) which provides another example of the arctic nival regime.

1.1.6 Sources of Data on Arctic Hydrology The largest bibliography on arctic hydrology is by Hartman and Carlson (1970). It cites some thousands of references on all aspects of the subject and from all parts of the Arctic. A more specialized bibliography is that of Dingman (1973a). Discharge data are collected from a number of streams in the North American Arctic, in the Yukon and Northwest Territories by the Water Survey of Canada (Inland Waters Directorate, 1968) and in

Alaska by the U. S. Geological Survey and Weather Bureau (Office of Water Data Co-ordination, 1970). The Yukon River is gauged at a total of sixteen cross-sections, and several of its tributaries are also gauged, but the Mackenzie River fares less well: of twelve surface water stations on the section of the Mackenzie north of 60°N, only two record discharge. Two surface water stations are located on streams, other than the Mackenzie, which drain into the Arctic Ocean. The two streams in question are the Coppermine and the Back; there are no stations on the Arctic Slope of Alaska.

Two recent additions to the Canadian network represent the first official attempts to collect water data in the High Arctic, for Freshwater Creek near Cambridge Bay, Victoria Island and for the Allen River some 12 km north-west of Resolute. The Allen basin has an area of 470 km^2 and its upper reaches are adjacent to those of the Mecham River. Records from the Allen River have been examined for comparison with data for the Mecham River, and as might be expected for two neighbouring streams draining similar terrains, patterns of response are closely similar.

1.2 Methods of Measurement and Analysis

The measurements made in the Mecham River basin were of two sorts, hydrological and geomorphological: most of the latter type were measurements of the chemical and sedimentary character of waters entering, flowing through and leaving the basin. The hydrometric programme was dominated by measurement of stream discharge and precipitation, although records of air temperature and other meteorological variables were maintained and a limited amount of data was collected on snow cover and snow density. Considerable use has been made in analysis of the records of the Resolute meteorological station.

Table 1:2 summarizes the variables which were monitored in the field and the methods used to collect data on these variables. Details of the methods are given in the following paragraphs, and where practicable an attempt is made to evaluate the accuracy of the method. The weather station at Resolute records an unusually complete series of meteorological data, and although the accuracy of these data may fall short of the maximum attainable in a careful study, the continuity and completeness

TABLE 1:2

Measurement of Variables, Mecham River Project

Variable	Unit	Instrumentation or method of measurement
precipitation	mm	tipping bucket rain gauges
snow density	kg m^{-3}	NRC snow sampler and Ohaus triple beam balance
snow temperature	deg C	mercury-in-glass thermometers
stream discharge	m^3 s^{-1}	Scientific Instruments of Wisconsin current meters and wading rods; Rhodamine WT fluorescent dye
water level	m	Leupold and Stevens A-35 and other stage recorders
suspended sediment concentration	ppm	DH-48 suspended sediment sampler
bed load movement		estimated
Ca	mol m^{-3}	titration with E.D.T.A.
Mg	mol m^{-3}	titration with E.D.T.A.
Na	mol m^{-3}	Perkin-Elmer 303 and 290 spectrophotometers
SO_4	mol m^{-3}	titration with $BaCl_2$ in presence of thorin
Cl	mol m^{-3}	titration with $AgNO_3$ in presence of K_2CrO_4
HCO_3,CO_3	mol m^{-3}	potentiometric titration with H_2SO_4
pH		Metrohm E280A portable pH meter with combination glass-reference electrodes

TABLE 1:2 cont'd.

Variable	Unit	Instrumentation or method of measurement
specific conductivity	$\mu mho\ cm^{-2}$ at 25°C	Barnstead PM-70CB conductivity bridge with Yellow Springs cell
water temperature	deg C	mercury-in-glass thermometers

SI units are used throughout this dissertation, with the exception that the micromho per square centimetre is preferred to the Siemens per square metre.

of the Resolute record more than compensate for this deficiency. A considerable deflection of energy from other aspects of the project would have been entailed, had reliance not been placed on official weather records.

1.2.1 Variables Measured in the Field Most data for the study were collected near to the mouth of the Mecham basin (Fig. 1:13). The river debouches on the eastern shore of Resolute Bay, and a base camp was maintained 250 m upstream from the bay on the right (north) bank of the river. Meteorological variables were monitored on level ground close to the base camp. 100 m upstream from the camp there is a gorge through which the Mecham River flows in a constricted rectangular channel; water level was recorded at a particularly stable cross-section within this gorge, and discharge measurement by dye injection was conducted through the gorge; area-velocity measurements of discharge were made either within the gorge or across the wider channel section at the base camp downstream, as were measurements of suspended sediment and solutes.

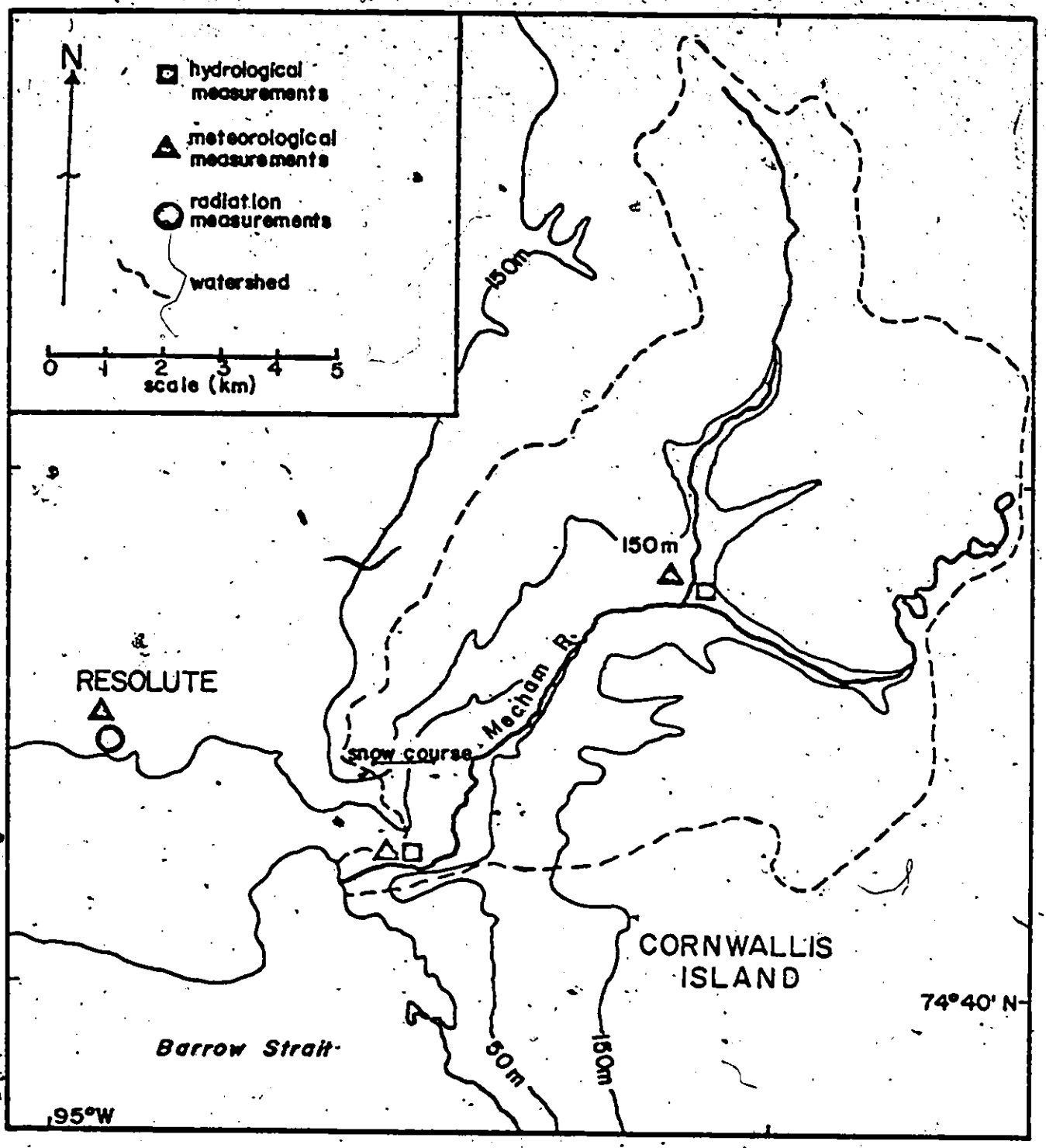


Figure 1:13. Sampling sites.

A number of samples were taken at a variety of points in the interior of the basin, to characterize the chemistry of waters flowing over the calcareous terrain. One site was visited repeatedly; at this site, the confluence of the two major headwater streams of the Mecham River proper, discharge and water quality measurements were made weekly and a record of water level maintained. On a plateau surface about 30 m vertically and 200 m horizontally from the confluence, precipitation and temperature were recorded continuously. This programme of observations was intended to serve as a basis for hydrological subdivision of the Mecham basin, but as such it was not wholly successful. It did, however, contribute useful information on spatial variations of the variables which were monitored at the confluence of the northern and eastern headwaters. (The location of this confluence is indicated on Figure 1:13).

<u>1.2.1.1. Precipitation</u> Precipitation was recorded in a tipping bucket rain gauge, with a daily chart at the base camp and with a weekly chart at the site

in the interior of the basin (Fig. 1:14). The buckets are calibrated to tip upon the accumulation of 0.254 mm (0.01") of rainfall, and each tip is represented as a small jump in the line traced on the chart by a pen. During rainfall events, resolution of one hour is achieved easily for rate of rainfall, which is calculated as the number of jumps per time unit on the chart. The rainfall totals, as indicated on the charts, were consistently reproducible by measurement of the rain gauge catch in a graduate cylinder. This reproducibility implies that the tipping bucket rain gauge performed comparably with the standard rain gauge for which the measuring cylinder was manufactured. It is said that recording rain gauges underestimate rainfall because, among other things, a finite amount of time is required for the buckets to tip, and during this time rain entering the receiver is lost. Reproducibility of the rain gauge catch in a cylinder is taken to be sufficient ground for ignoring this problem of underestimation, and the rainfall records from the standard gauge at the Resolute weather station are regarded as comparable with the records collected in the Mecham basin.

Figure 1:14. Tipping bucket rain gauge

Figure 1:15. The recording cross-section and stage recorder.

The more general problem of accurate estimation of rainfall is an intractable one. Aside from the question of spatial variability in rainfall amounts, measurement of rainfall at a point is complicated by the two necessities of minimizing turbulence in the neighbourhood of the gauge and eliminating "splashback" of bouncing raindrops. The compromise which is usually effected, that is, to raise the gauge orifice some height above the ground, succeeds in eliminating splashback at the expense of increasing turbulence, and consequently the gauge underestimates "true" precipitation. The magnitude of the underestimate increases with wind speed, and high wind speed is often associated with high precipitation intensity. No acceptable rule is known for correcting this underestimate, and the measured values of precipitation have been accepted for purposes of calculation and analysis along with an unknown amount of error.

1.2.1.2 Measurements on Snow. Snow depth was measured in spring 1972 along a snow course of twenty snow stakes (Fig. 1:13), spaced at 50 m intervals on a line running eastward down the western flank of the Mecham valley. The 1.5 m long stakes were driven into the ground in summer 1971. Alternating decimetre sections of each stake were painted in red and white, and centimetre values

within the relevant decimetre subdivision read from a ruler. The snow course began on the flood plain, continued up the valley side and ended with a number of stakes on the plateau. Areas of snow accumulation and drifting were thus sampled in a way which was, subjectively speaking, representative of the basin as a whole, although no deep gullies were included. The depths obtained were used for comparison with figures derived in water balance calculations.

Snow density was measured for samples taken from the walls of snowpits with an N.R.C. sampler. The pits were dug in snowbanks near to the base camp. The sampler is a handheld device for procuring cores of volume 250 cubic centimetres; usually two cores were taken at each of from 4 to 12 positions in the pit wall, and the weight of the cores measured to the nearest 0.01 g with a triple-beam balance. An alternative procedure was to melt the sample and measure its volume to the nearest cubic centimetre in a graduated cylinder. The density was calculated as the ratio of water equivalent volume to core volume, the first two digits of the result being regarded as significant.

Snow temperature was measured at the same positions in the pit wall from which density samples were taken. A

mercury thermometer was pushed into the snow and left for 2 min, the temperature reading being taken immediately after the thermometer was withdrawn.

1.2.1.3 Discharge Stream discharge was measured by one of two methods: the conventional area-velocity method, using current meters to measure velocity, and dye dilution gauging with the fluorescent dye Rhodamine WT. The current meters were standard and "pygmy" type meters manufactured by Scientific Instruments of Wisconsin Inc., and most velocity readings were taken with a meter mounted on a top-setting wading rod. For most measurements of discharge, velocity was measured at 0.6 of total depth from the water surface at 10-20 verticals across the channel section. The current meters measure velocity with an accuracy of ± 1 per cent, but figures for discharge are less accurate than this. The reason lies in the difficulties inherent, firstly, in the estimation of channel shape and, secondly, in measuring total cross-sectional discharge quickly enough that the discharge can be considered constant throughout the period of measurement. Channel width is easy to measure, but where the channel is shallow and/or has an irregular bed a single measurement of depth may

have little value for extrapolation; it is vital that discharge should not change significantly while it is being measured, and this usually means that a measurement must be completed within 30 min. The compromise which must be made between these difficulties leads almost always to a very generalized model of the cross-sectional shape, and usually to a smaller than ideal number of velocity measurements. In a vigorously flowing stream of very cold water, the physical act of measurement is often accompanied by some discomfort and no little risk so that, although difficult to evaluate, operator performance is probably below its optimum under these conditions.

Discharge, therefore, is difficult to measure accurately. An estimate of the accuracy is unlikely to be more than an approximation, but the reproducibility obtained from repeated measurements at identical stage suggests that, for the work reported here, the accuracy was better than ± 10 per cent and of the order ± 5 per cent. Measurements made by dye-dilution gauging were more accurate than area-velocity measurements; for an explanation of the principles of the technique see Church and Kellerhals (1970). Slugs of a solution containing 20-60 ml of the pink dye Rhodamine WT were thrown into the Mecham River at the entrance to the gorge,

and sequential samples of the stream water were taken as the dye cloud passed a point 500-600 m downstream from the point of injection; complete mixing of the dye in the water had occurred before the sampling point was reached, and the time of travel between the two points varied from 7 to 12 min. The fluorescence, and thus the dye concentration, of the samples (numbering about 15 to a gauging) was measured later with a Turner 111 fluorometer, and the discharge calculated from the shape of the dye wave passing the sampling point.

Discharge was measured mainly to calibrate the stage record obtained from a Leupold and Stevens A-35 water level recorder (Fig. 1:15). Stage was measured with a float mounted in a stilling well, and its value transmitted to a pen moving along a strip chart; as with discharge itself, the instrumental accuracy of the stage recorder (± 0.06 mm across a 0.3 m chart width) exceeds the accuracy of any interpretation about discharge which can be made from its readings. The cross-section at which the stage of the Mecham River was recorded has as nearly a rectangular shape as could be desired, and a simple power law gives a very close fit for the rating curve which describes discharge as a function of stage. Since the

stage recorder was attended continuously (throughout summer 1971 and during parts of the summers of 1970, 1972 and 1973), or at least visited frequently, it was possible to adjust the rating curve when minor changes occurred in the cross-section, for example after the rainstorm flood of 3-4 August 1971. A new rating curve was also calibrated afresh each year until 1972. That for 1972 is graphed in Figure 1:16; in this case discharges exceeding the highest measured value occurred while the recorder was unattended, and the curve had to be extrapolated. Although this is undesirable, it is probably the most accurate solution of the problem.

Unfortunately the rating curves can be applied to the stage record only when the channel shape approximates that which prevailed when the discharge measurements were made. This is not the case in the earlier periods of any given flow season, for when integrated channel flow begins in the spring the channel is choked with snow, and as the season progresses the channel grows by destroying its walls and bed of snow. In consequence the channel shape changes continually and discharge is not a function of water level. This situation prevails for 5-15 days, and for a part at least of the annual flood a reliable, continuous record of discharge is hard to obtain,

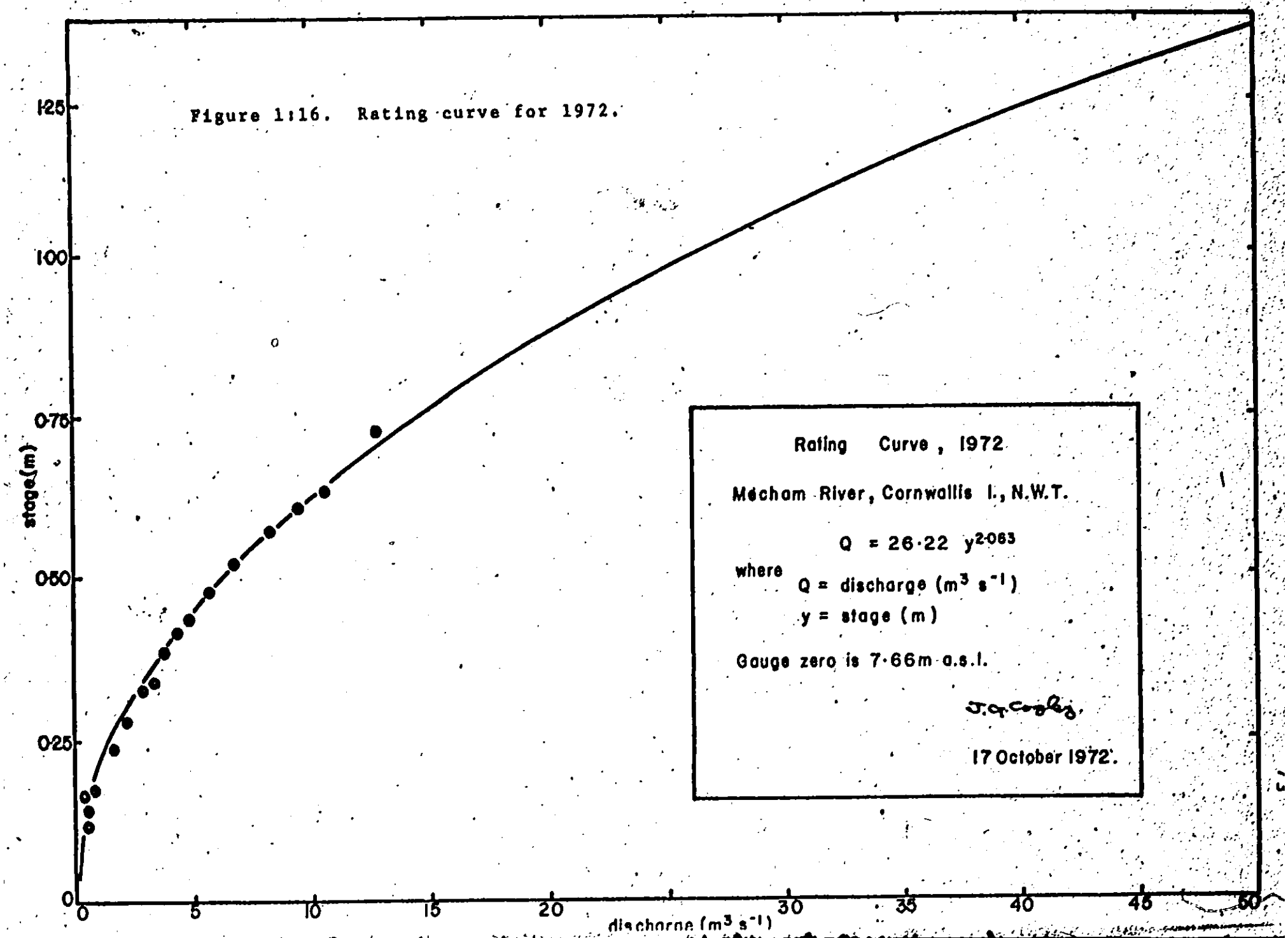


Figure 1:16. Rating curve for 1972.

both because it is difficult to install and maintain a recording device and because a record of stage, if obtained, is difficult to interpret in terms of discharge. In 1970 and 1971 the stage recorder was not installed until after the peak of the snowmelt flood, and the continuous discharge graphs presented in later sections for these two summers contain interpolations between spot measurements of discharge. For 1972 and 1973 continuous stage records cover the flood period, but their corresponding discharge graphs contain interpolations and some subjective interpretation, at least up to the time when all the snow was cleared from the channel. In 1972, for example, this occurred on 12 July.

It is almost impossible to state the accuracy of discharge figures obtained from the rating curve, for the reasons given above and also because there is no standard against which to check the figures. At worst the accuracy may be as bad as ±25-30 per cent, although the great majority of the figures are more accurate; at best the accuracy approaches that of the individual discharge measurements used to derive the rating curve.

1.2.1.4 Measurements of Sediment Suspended sediment concentrations were measured by filtering samples, taken with a DH-48 hand-held sampler, through a Sartorius filtration apparatus, and weighing the accumulation of sediment on the filter. Only one bottle was filled at each sampling, at a point as near to the middle of the channel as could safely be reached by wading. This method was adopted to minimize the time consumed by the filtration procedure, but was considered reasonably representative under the prevailing turbulent flow conditions. The filters had a mean pore diameter of 0.45 μm. Except for concentrations close to zero, where changes in filter weight during drying begin to confuse the issue, the determination of the suspended sediment concentration in the sample bottle is relatively precise: it is possible to discriminate between differences of 0.01 g in a sample weight of ~400 g, but extrapolation to the entire cross-section and to periods longer than the sampling interval naturally introduces some error.

Waterborne movement of sediment in contact with the channel bed is almost impossible to measure under field conditions. Very crude estimates of its magnitude may be obtained by observing the accumulation of coarse sediment in natural traps in the bed, and by other visual methods. For the Mecham River a record was kept

of when bed material was or was not in motion, and an estimate of bed load transport derived from an application of the Meyer-Peter bed load equation.

Most of the major ions in solution were measured in the waters of the Mecham River, as well as in waters flowing through its basin and those of other streams. Samples for analysis of water quality were taken in polyethylene bottles which were immersed in the stream of water, allowed to fill completely and capped while still under water. Water temperature was measured with conventional thermometers at the time of sampling. The procedures for analysis of Na, SO_4 and Cl are rather complicated, and these analyses were done on samples returned to McMaster University at the end of each field season. Analyses of Ca, Mg, alkalinity (as HCO_3), pH and specific conductivity were done as soon as samples could be returned to a tent at the base camp. In the case of samples from the Mecham River, this involved a time lag of a few minutes at most, although the delay reached 3-4 hours for samples from more distant points. Returning samples to the tent caused an increase in the temperature of the sample, of 0-2°C for Mecham River samples and of up to 5°C for some others, but this was considered acceptable because of the difficulty of working efficiently and accurately in the open.

As a rule, samples for analysis were not filtered, since a pump was not available. Most samples contained very little suspended solid material, but in the turbid conditions prevailing during the snowmelt flood some overlap undoubtedly occurred between the separate analyses for solid and dissolved sediment. In general it is desirable to measure dissolved ions in filtered samples, but experienced practitioners recommend that, for instance, alkalinity be measured in the unfiltered sample. Moreover, it is difficult to make any reliable interpretation of the pH of a natural, turbid water if the pH is measured in a filtered sample of that water. A few tests in the field suggested that Ca and Mg determinations differed by little or no more than the claimed accuracy of the method used, when the determinations were repeated on filtered and unfiltered samples of the most turbid water encountered.

Ca and Mg were measured by titration with E.D.T.A. (disodium dihydrogen ethylenediamine tetracetic acid), using materials retailed by British Drug Houses Ltd. The method is similar to that described by Rainwater and Thatcher (1960), and is considered accurate to $\pm$ 0.02 mol m^{-3}. Alkalinity was measured potentiometrically, by titrating a sample aliquot of 50 ml

with a 0.0164 N solution of H_2SO_4 and observing the rate of decrease of pH. The volume of acid consumed on reaching the end-point of the titration (where rate of decrease of pH was greatest) was used to calculate total alkalinity, and the species HCO_3^- and $CO_3^=$ were separated by calculation using Wigley's (1972) computer program (v. infra, where the accuracy of the alkalinity determination is discussed under measurement of pH).

Na^+ was measured with a spectrophotometer. The machine was calibrated with standards having concentrations in the range of 0.0 - 0.2 mol m^{-3}, which bracketed all of the samples analyzed. The Na^+ concentration of the sample was taken as the mean of at least three readings from the spectrophotometer. The accuracy of the values used was ± 0.01 mol m^{-3}. $SO_4^=$ was analyzed in the laboratory using the visual thorin method described by Rainwater and Thatcher (1960) and involving titration of an acidified sample aliquot with $BaCl_2$, in a medium of 1,4-dioxane and the dye thorin. The results were accurate to ± 0.001 mol m^{-3} (= ± 0.1 mg ℓ^{-1} $SO_4^=$). Cl^- was analyzed, also in the laboratory, by the volumetric Mohr method of Rainwater and Thatcher, in which the

sample is titrated with $AgNO_3$ using K_2CrO_4 as an indicator. The accuracy of the method is ± 0.02 mol m^{-3}.

The determination of pH in arctic field conditions is difficult. Since pH is an important variable because it is used in calculations of several other variables, and since the method of pH measurement used in this study was relatively successful in adverse condtions, the method is described here in considerable detail. Its use is recommended in similar situations.

Since there is no line power a portable, battery-operated meter is essential. Ambient temperature is rarely above 10°C, and relative humidity is generally 60 to 80 per cent or greater; in these circumstances the meter is slow to warm up and its response remains sluggish even when it is warm. The sluggishness of response is aggravated, moreover, by undesirably high resistances in electrode glasses at low temperatures. Essentially a pH meter is a precise potentiometer, and its response is usually expressed as a number of millivolts per pH unit. At 25°C the ideal response is 59.157 mV/pH unit, decreasing with temperature to 54.197 mV/pH unit at 0°C (Bates, 1964). Very few meters display this

ideal response, and their actual response is a function of temperature (allowed for with a built-in compensator), a deviation attributable to manufacturing imperfection and a deviation of a random kind, attributable to ambient conditions at the time of the pH measurement. Figure 1:17 shows the result of a calibration of the meter used in this study, under near optimum conditions in the laboratory. The deviation from the ideal response may be thought of as a change in the slope of the e.m.f. (electro-motive force) - pH relationship; changes in the position of this relationship are compensated easily by a simple mechanical adjustment of the meter's zeroing switch, but changes in slope are insidious and may pass unnoticed if pH measurements are not done carefully. Under the near-optimum conditions of Fig. 1:17 deviations from ideal response are seen to lead to an over-estimate of true pH of about 0.06 pH unit at pH 8.00, and of 0.26 pH unit at pH 4.01, using as primary standard the 0.05 M borax buffer solution of pH 9.18 at 25°C. In field conditions the over-estimate is rarely less and sometimes much more than this, reaching more than 1.00 pH unit at pH 4.01 in the worst cases experienced.

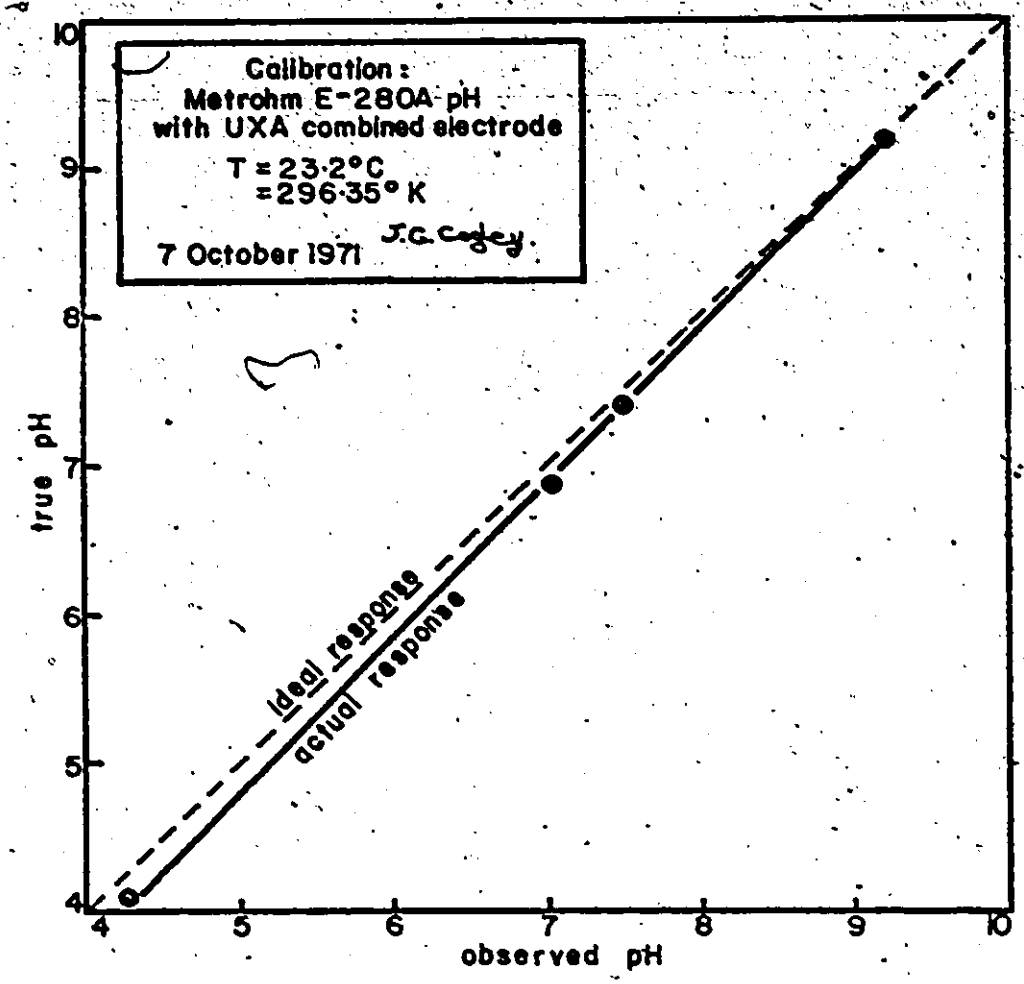


Figure 1:17. Calibration of pH meter.

To correct for this defect in meter performance a modification of the procedure of Barnes (1964) was adopted. The procedure incorporates improvements in the determination of both pH and alkalinity: the correction for deviation from ideal response, and a more exact location of the end-point of the alkalinity titration. Combined glass-reference electrodes of standard and low-temperature (lower-resistance) types were used. The standard buffer solutions used for calibration were the borax solution (pH 9.18 at 25°C) and the phthalate solution (pH 4.01 at 25°C). The electrode was washed twice in de-ionized water (specific conductivity = 2.7 μmho cm^{-2} at 25.0°C) and once in the solution to be measured before each actual measurement, and the temperature compensator was reset as and when necessary. The procedure was as follows:

i) allow meter to warm up for about 60 min;

ii) read pH of borax buffer solution, then return to "standby": repeat as often as needed to obtain satisfactory readings for both standby and buffer pH at the relevant temperatures;

iii) read pH of sample solution without stirring, then return meter to standby and adjust mechanically for

any zero shift: repeat until at least two consecutive readings yield the same sample pH;

iv) titrate, with minimal stirring, with 0.0164 N H_2SO_4 to the alkalinity endpoint in the region pH 4.5 - 5.3, adding acid drop by drop near to the endpoint and recording decrease in pH with each increment of acid;

v) return meter to standby, then read pH of phthalate buffer solution.

Step iv) yields a better estimate of the volume of acid consumed in the titration, and an accuracy of $\pm$ 0.016 mol^{-3} (= $\pm$ 1 mg ℓ^{-1} HCO_3^-) is obtained. Step v) yields an estimate of the actual meter response and enables the following correction to be applied to observed pH:

$$pH_w^t = pH_b(pH_w^o - pH_p^o) + \frac{pH_p^t\ (pH_b - pH_w^o)}{pH_b - pH_p^o}, \qquad (1:1)$$

where the superscripts t, o indicate "true" and "observed", and the subscripts w, b, p indicate "water sample", "borax buffer" and "phthalate buffer" solutions respectively. The claimed accuracy of the pH meter is $\pm$ 0.05 pH unit, but it is readable to $\pm$ 0.02 pH unit and most measurements were reproducible

to this figure also. With the correction of eq. 1:1 applied, pH values are regarded as accurate to + 0.04, - 0.02 pH unit. The higher positive error is assumed because the meter reading approaches a stable value from below (for pH above 7.00, which is almost always the case in this study) and it is possible that this stable value may not always have been attained due to sluggishness of response. Under the worst conditions encountered, the accuracy was probably somewhat worse than this; on one or two occasions meter performance was so poor that a measurement had to be abandoned. Turbidity of the water sample had a generally adverse effect on performance, and this was transmitted to the alkalinity measurements. pH values for turbid samples, such as those taken during the snowmelt flood, are probably as accurate in themselves, although acquired with greater effort, as the values obtained for clear samples; however, the interpretation of the pH of a turbid water is difficult, as mentioned above, for the hydrogen ion activity of a water-sediment mixture is in general different from that of the same water with sediment removed. The accuracy of alkalinity measurements on turbid samples is affected by a tendency

towards instability in the meter reading during the titration: having registered a decrease, the meter needle tends to creep upwards again. It is difficult to estimate the rate of decrease of pH, and consequently difficult to fix the end-point precisely. Alkalinity values reported for the more turbid waters have an accuracy somewhat less than that quoted above of $\pm$ 1 mg ℓ^{-1} HCO_3.

Specific conductivity was measured in the field, shortly after sampling, with a meter having a digital readout system. The specific conductivity of a water is strongly dependent on its temperature, and the usual procedure is to refer all measurements to a standard temperature of 25°C. This is done by employing an approximate numerical relationship or by allowing samples to adjust to "room" temperature before measurement. The values used in this study were obtained by an improved procedure involving rapid heating of a sample aliquot over a stove, to about 26°C, and observation of the value read out by the meter as the sample temperature fell through 25.0°C. The claimed accuracy of the instrument is $\pm$ 1 per cent, but in laboratory calibrations using standard solutions of KCl the accuracy achieved was only better than $\pm$ 3

per cent for the range of values encountered in the field.

1.2.2 Meteorological Data from Resolute The officially collected weather data used in this study include records of daily rainfall, daily snowfall and their sum, daily precipitation; daily minimum, mean and maximum air temperature; twice daily ground temperature; six-hourly precipitation; hourly windspeed, wind direction, wet bulb temperature, dry bulb temperature, global solar radiation, reflected solar radiation and net radiation. Different time standards prevail in the recording of these data, and where necessary all observations were referred to Eastern Standard Time, which is used for all time series data in this study. Certain changes have been made in collection practices during the period of record, and these have been allowed for where appropriate. They are documented, together with details of data storage formats, in publications of the Meteorological Branch (1964, 1966, 1967, 1970b).

Standard measurement methods are used at the Resolute meteorological station. All of the data used have been accepted at their published values, although manipulation of the data is naturally limited by the resolution of the measurement method in each case.

1.2.3 Computing Most of the results of analysis described herein would have been unobtainable without the use of a digital computer and off-line graphic plotter. The more important computer programs used in this study are listed in Table 1:3.

TABLE 1:3

Computer Programs Used in Analysis

Name	Purpose	Source/Authorship
AUTO	to compute the autocorrelation function	S.S.P.
AUTOP	to plot the autocorrelation function	J.G. Cogley
MBD02T	to do spectral and covariance analysis	BMD
CROSS	to compute the crosscorrelation function	S.S.P.
CROSSP	to plot the autocorrelation function	J.G. Cogley
EVAP	to compute evaporation with the combination model	J.G. Cogley
FOUREX	to compute and extract Fourier harmonics from a data series	M.K. Woo
GRAFF	to plot graphs	J.J. Drake, J.G. Cogley
HAYREG	to estimate the parameters of a multiple regression	S.S.P.
MAGIC	to compute and plot statistical regressions	J.G. Cogley, J.J. Drake
MTYP	to read and (using subroutine ID) to unpack meteorological data	M.K. Woo

Name	Purpose	Source/Authorship
MVAP	to compute monthly evaporation with the combination model	J.G. Cogley
NLINSQ	to estimate non-linear parameters by the least squares method	D.W. Marquardt
TOMCHEM	to derive measures of aqueous chemistry such as saturation indices with allowance for ion pairing.	T.M.L. Wigley

SECTION 2

THE RUNOFF REGIME

2.1 Introduction

2.1.1 Rationale As far as is known, very few geomorphic events happen in the High Arctic winter. This may be simply because very little geomorphic research has been done in winter in the High Arctic, but more probably it is because water only flows in the summer. If flowing water is an important geomorphic agent at high latitudes, the hydrological events of summer assume a corresponding importance as the backdrop of the geomorphic stage. They also have an intrinsic interest because water itself is an important arctic resource, and because flowing water is capable of damaging structures and thwarting plans whether or not it has the aid of moving sediment.

As work on the goals defined in sec. 1.1.2 progressed, certain aspects of them became more prominent because unsuspected peculiarities came to light. In consequence some of the others have yet to be analyzed exhaustively; questions raised at the beginning of the study remain unanswered, and to them new ones raised during the study have been added. Section 2 is intended to define some of these questions, to suggest ways of

answering them and to provide a context for the sections which follow.

2.1.2 The Annual Flood The shortness of the annual flow season, and the concentration of most annual runoff into an even shorter period, were among the first things to excite interest in High Arctic hydrology. Cook's five-part schema (1967) has been mentioned earlier: it gives a good conceptual framework on which to build an explanation of Figure 2:1, which is a graph of the discharge of the Mecham River for the five years 1970-74. The shortness of the hydrological "year" is emphasized by the omission from Figure 2:1 of most of the first of Cook's five periods, the winter period of no flow, which occupies about four-fifths of the year. The second period is that during which the winter snowpack is prepared for melting by the processes of warming and ripening. There is no evidence in the hydrographs for the existence of this period, but it is important and will be referred to again. The hydrographs represent the last three periods: the period of initial low flow, the flood and the period of prolonged recession of flow from its maximum value.

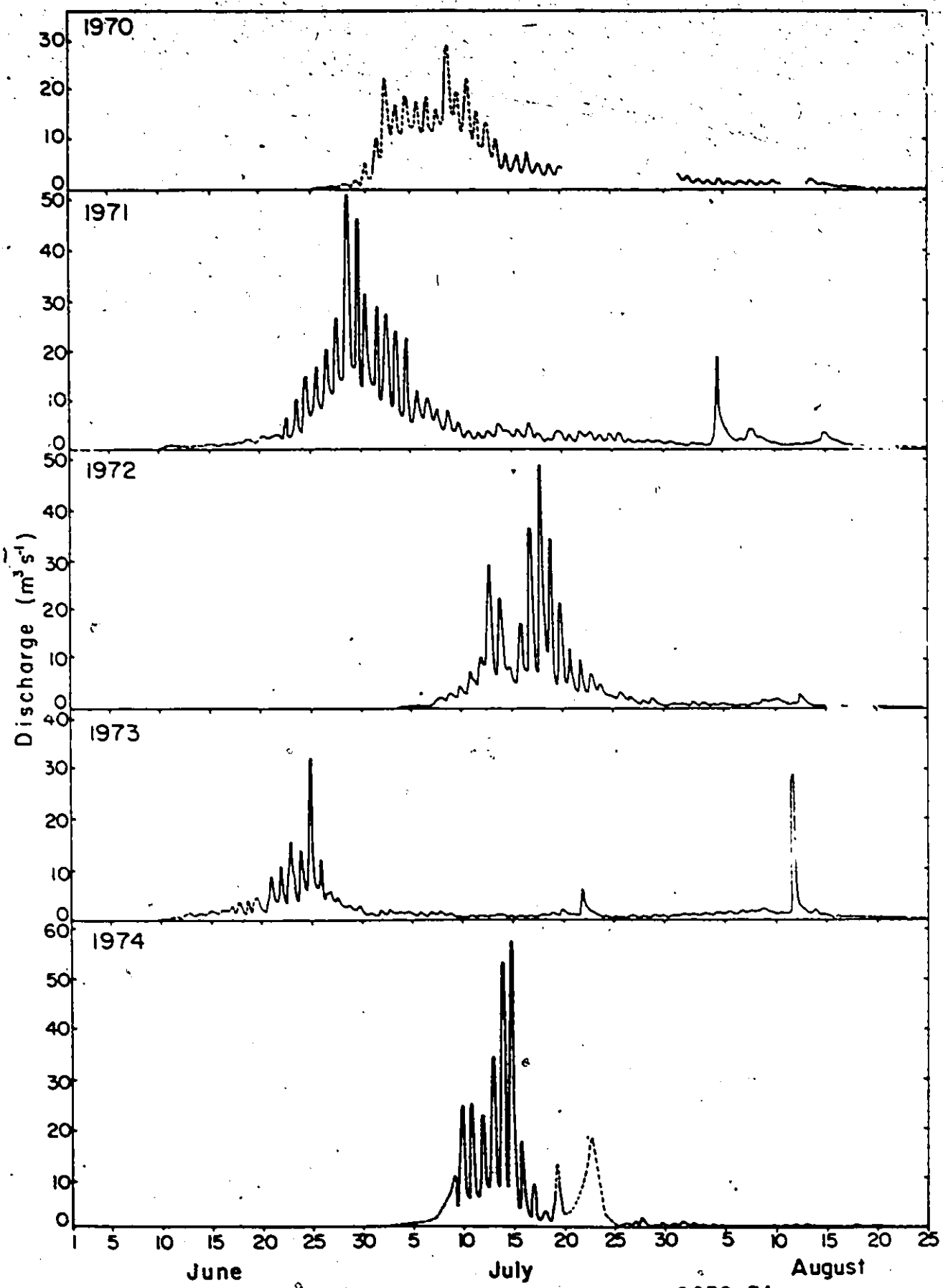


Figure 2:1. Discharge of the Mecham River, 1970-74.

At least three features of the annual hydrograph deserve notice. These are, first, the set of points which define the curve, among which some such as the dates of first and last flow are of particular interest; second, the collection of points which represents the season's maximum discharges; and third, the area under the curve of the hydrograph, which is equivalent to total annual runoff. This last feature is a focus of the section following this one, and is treated comparatively and at length there, but it will be useful to begin this section with a descriptive analysis of the more interesting and significant points shown on Fig. 2:1. Although the five hydrographs share a common basic form, the differences between them are striking and may be of great practical importance.

The most striking difference to be seen in Fig. 2:1 is in the timing of the annual flood. Twenty-three days separate the extreme recorded dates of annual maximum discharge, and roughly the same can be said for the extreme dates of first flow although this date is not known for 1971 and 1973. Table 2:1 summarizes the dates of major events during the five flow seasons of this study and during 1959, the year of Cook's observations (1967). Roughly speaking, the flow

TABLE 2:1

Summary of Discharge Events, 1970-1974

Year	Start of Flow	Date of Maximum Discharge	Maximum Discharge ($m^3 s^{-1}$)	Days to 75% of Total Flow[a]	Duration of Flow Season (days)
1970	25 June	8 July	26.6[b]	-	>55
1971	10 June[b]	28 June	51.2	35[b] (24)	>71
1972	4 July	17 July	48.2	15	>51
1973	10 June[b]	24 June	32.1	50[b] (42)	>74
1974	2 July	14 July	57.1	-	>50
1959[c]	20 June	2 July	52[b]	-	83

[a]The number of days from start of flow (start of measurement in parentheses) to the date at which cumulative runoff reached 75 per cent of total annual runoff.

[b]Estimated

[c]From Cook (1967).

season can be expected to begin at Resolute sometime between early June and early July, the mean date being in the last two weeks of June. In the years for which there is a reliable date of first flow, the number of days to the annual peak discharge has consistently been either thirteen or fourteen. On such limited empirical evidence it seems reasonable to suspect that this number is a conservative quantity from year to year for a given basin, and the estimated date of first flow for 1971 (10 June) may be a few days too early.

The instantaneous maximum discharge of a given year seems also to vary relatively little. The figure quoted in Table 2:1 for 1970 is probably an underestimate, but for the years (1971-1974) when maximum discharge is accurately known the mean is 47.2 $m^3 s^{-1}$ and the largest discharge is less than twice the smallest. If, as appears likely, the snowmelt flood is ordinarily the largest of the year, the prediction of damagingly high discharges may prove reducible to the prediction of unusually snowy winters. Naturally the progress of each year's flood is a function of its immediate meteorological antecedents, and these leave ample scope for variation. But on the evidence of Fig. 2:1 it appears that the amount of snow accumulated in winter has a considerable influence on the high discharges attained by the stream in spring.

The intensity of the annual flood varies somewhat from year to year. It is difficult to find a simple measure of this "intensity", which can be thought of roughly as the prominence of the spring peak with respect to discharges later in the season. The measure chosen, number of days needed to discharge 75 per cent of total annual runoff, is listed when possible in Table 2:1, but does not really indicate successfully what is meant by the concept of intensity. Fig. 2:1 displays the idea more clearly, however. For example, the year with the smallest reliably known maximum discharge, 1973, was a year in which the snowmelt flood as a whole was subdued; summer "baseflow" was lighter for a longer time than in other years, and it happens that two rainfalls provided prominent blips in the hydrograph. In 1974 conditions were different. The highest discharge yet recorded for the Mecham River occurred on 14 July, but the recession from this peak was more rapid and baseflow for the rest of the season was lower than in any previous year.

Differences in total annual runoff are to be discussed later, but it is appropriate to note here a number of other features of the annual flood. First, the course of the flood is only exceptionally a simple

increase to a maximum followed by a slower decrease. This is what happened in 1971, but in 1970, although the record is not continuous, observation showed that the increase to high discharges was abrupt and that one day of high flow (2 July) was followed by several days with lesser maxima before the annual peak was reached on 8 July. 2 July 1970 was a day on which a large but short-lived burst of water passed down the channel after the clearance of a dam of drifted snow blocking the mouth of the Mecham gorge. It is usual for snow to accumulate in the channel bed, and for flow in the early days of the season to occupy temporary channels with snow walls and floors. But 1970 is the only year on record in which the snow in the channel was so bulky that water arriving from the interior of the basin was delayed significantly by damming.

In each of the other years of record there was also no uniform increase to the flood peak. This is most visibly true of 1972 but can be seen in 1973 and 1974 as well. That immediate meteorological antecedents may control the course of the flood to this extent is worth noting, for although the characteristics of the arctic nival regime may be described straightforwardly the sequence of events which leads to the observed discharge fluctuations is by no means straightforward.

Diurnal fluctuations are an immediately noticeable feature of Fig. 2:1. During the greater part of any given year each day's maximum discharge may be expected to be at least double the same day's minimum discharge, whether or not the day itself is in the period of the spring flood. But the daily variation is naturally most significant during that period. On 14 July 1974, for example, a low of 8.41 $m^3 s^{-1}$ was reached at 1100 h as discharge receded from the maximum of the previous day; by 1900 h, eight hours later, discharge had increased by 686 per cent to 57.7 $m^3 s^{-1}$. This marked control of discharge by radiation usually persists through spring and summer until flow becomes so low that the fluctuations are too small to be monitored, or more often until they are damped out by rainwater runoff.

Rainstorms occur in most years, and the responses to them are strongly peaked. Most of the rainstorms, however, occur in middle or late summer, by which time even a small rainfall recession flow suffices to drown the still smaller production of meltwater. The storm responses are for the most part readily identifiable in Fig. 2:1, rising abruptly as they do out of the downward seasonal trend of snowmelt runoff. None were recorded in 1970, three in 1971 (4, 7, 13-14 August),

one in 1972 (12 August) and two in 1973 (22 July and 11 August); unfortunately the recorder pen was raised from the chart during the 22.4 mm precipitation event of 21-22 July 1974. The smaller storm of 1973 is interesting, not least because of its smallness, for the same weather system deposited over 50 mm of precipitation (a record-breaking amount) at Vendom Fiord, south central Ellesmere Island, the phenomenon is discussed in McCann et al. (1975) and by Cogley and McCann (in press).

2.1.3 The Arctic Nival Regime Fig. 2:1 shows clearly, and perhaps more comprehensively than has been done before, the characteristics of Church's arctic nival regime (1974). All five hydrographs are dominated by the spring snowmelt flood; this, and an absence of winter flow, are the two most important distinguishing features of the regime. The hydrograph for 1973, however, points to the possibility that rainstorm responses may infrequently exceed the snowmelt maximum of the year in which they occur.

The behaviour of the Mecham River seems in all important ways to typify the arctic nival regime, accepting Fig. 2:1 as a paraphrase of the published

documentation (sec. 1:5): because arctic nival basins are impermeable there is no contribution to flow from groundwater, and water only runs off basin surfaces when it melts from the snowpack or falls as rain; the ratio of rainwater to snowmelt runoff is relatively high, but snowmelt discharges, as far as is known, almost always exceed rainwater discharges; although Cook's date for 1959 (Table 2:1) is the only one known for the end of flow, the flow season is clearly short, and basin responses are usually rapid.

As studies progress, however, variations come to light. For example, in continuously vegetated basins in south central Ellesmere Island (McCann et al., 1975; Woo, 1975), which is drier than Resolute, it appears that flow becomes negligible or stops altogether much earlier in the season than at Resolute.

In further work more relatively minor differences such as this will no doubt be found, but with existing knowledge conditions are ripe for analytical and predictive procedures to be developed for handling the arctic nival regime. These procedures are outlined in the rest of this section, although much remains to be done on the problems which are involved.

2.2 The Prelude to Flow and Its Onset

It would be of considerable practical value to be able to chart in advance the progress of the snowpack towards ripeness and eventual overflow, the stage at which snowmelt runoff proper begins. In winter the snow covering High Arctic terrain is dry and often hard, so that with appropriate equipment mobility presents few serious problems. In early spring, however, the snow begins to melt and the pores of the snowpack to fill with liquid water. Travel on foot and by vehicle becomes difficult, watercourses in particular being dangerous because they are among the first places where the snowpack is saturated. When flow begins the danger increases, and at the height of flood runoff even rich and resourceful groups such as oil exploration companies may be hindered for a time in their movements about the terrain. In spring, the decision to switch to wheels from tracks or skis may, if judiciously timed, save some weeks of inconvenience.

Considerations of this sort give the impetus of usefulness to the study of a problem which is interesting in its own right. Since the sequence of events to be predicted always occurs in mid June or mid July the problem of long-term prediction is in that sense a

trivial one. But more precise short-range estimates demand attention to detail, and specifically to the details of the energy regime of the basin snowpack in the period before runoff begins. An example of a very detailed model of one part of the process, which is too complex to be rewritten for practical manipulation in the present state of the hydrologic art, is that of Male and Norum (1971). On the other hand, it is possible to model or rather mimic the processes of interest quite crudely, using for example the degree-day as an index of heat entering the snowpack. Since an understanding of the data presented and used in this section will be aided by an account of the physical processes they represent, such an account is given below. The aim of the account is to strike a balance between rigour and rule-of-thumb, so that the account itself might be usable at least in principle as a working model, and also so that the theoretical status of various rules of thumb can be appreciated more clearly.

2.2.1 The Energy Balance of the Snowpack. The thermal state of the snowpack is a function of the energy entering and leaving it. Within the snowpack itself, however, energy

gradients are common. Although most models of snowmelt incorporate very little information on the inner workings of the snow cover, it is important to be aware of the simplifications which are involved.

First, until the snowpack becomes "ripe", that is to say isothermal at 0°C, it is usual to observe considerable variation of temperature with depth. The snow can be thought of as a layer with energy exchanges going on at the top and bottom: the exchanges at the top are much greater than at the bottom because turbulent convection of energy is possible between the snow surface and the atmosphere; between the basal snow and the underlying ground most of the exchange is by conduction, and is much smaller. Temperatures at the surface, therefore, follow the fluctuations of atmospheric temperature with a certain time lag. When the snow is below the melting point some of the energy which is absorbed is used to change the temperature of the topmost layer, and some is transmitted into the snowpack. Neglecting transfers of sensible and latent heat, the most important energy source at the surface is solar radiation; although this is the force driving energy changes within the snow, the amount of insolation is not equal, or even approximately equal to, the flux of energy into the snow. The reason is that snow has a very high reflectivity or

albedo: snow surfaces commonly reflect 60 to 90 per cent of the incoming solar radiation. Not all of the received radiation is reflected at the surface itself, however, for some penetrates into the snowpack before it is reflected back. This phenomenon, the gradual extinction of radiation penetrating the snow cover, makes the solution of the energy balance more complicated, but the important general point is that most of the energy available to the snow from above is reflected without entering the energy balance.

The principal process by which energy is transferred within the snow is conduction. The maintenance of high temperature gradients is helped by the low thermal conductivity of snow; although it is a function of density, and more markedly of liquid water content, the thermal conductivity of snow is well below that of most soils, and also that of ice. Without direct measurements of snow temperature, therefore, it can be risky to estimate the disposition of energy between (to take this study for an example) the air below the height of atmospheric measurements, the soil above the height of soil temperature measurements, and the snow lying in between.

The problem would, however, not be unduly complicated if only conduction were the sole mechanism of heat transfer. There is evidence that in some circumstances,

mainly where the granular structure of the snow is loose, convection of heat by air moving through the pores may be a significant component of the energy balance. Sublimation and condensation within the snow cover may also result in appreciable transfers of energy through the depth profile, but the most important of these subsidiary transfer mechanisms, at least in the context of modelling snowmelt, seems to be advection by percolating meltwater.

Once the upper layers of the snowpack have reached the melting point, some of the continuing income of radiation is used to melt the snow. The water thus produced drains under gravity to the lower layers which are still below freezing point; on freezing, the water releases the latent heat which was consumed in melting it, and this heat warms the lower layers of the snowpack. By this process the temperature gradient is reduced and the ripening of the snowpack is accelerated; the accumulating meltwater itself has a higher thermal conductivity than the surrounding snow, and there is further acceleration for this reason.

There is thus some overlap between ripening and melting, but once the snowpack is fully ripe it begins to behave as a porous, granular medium. It has a capacity

to store water, and water may flow through it, but the situation is of course complicated because the medium is also the source of the moving fluid. The liquid water capacity of ripe snow is commonly quoted as 3 to 5 per cent by volume, but the paths taken by meltwater through the snow are often discrete: "channeling" over impermeable ice layers and through parts of the snowpack with more open granular structure leads to saturation in some places earlier than in others. Allowing for the corrugations of the terrain underlying the snow, it is natural to expect that the first parts of the snowpack to become saturated are those occupying watercourses and channel floors. At first, the overflowing water fills temporary ponds in hollows in the snow, or between snowdrifts and channel walls. The final stage leading to the onset of flow involves thermal and mechanical disintegration of the snow barriers between these growing ponds; this process normally requires from one to several days depending on the peculiarities of individual reaches of channel, and when it is complete the snowmelt flood has begun.

To predict accurately, then, the date on which flow begins would require ideally a knowledge of the detailed disposition of energy within the snowpack, and also of the movement of meltwater through the snow. In

practice such knowledge would be difficult to gather, and in fact the information which is gathered is scarcely adequate for the problem. Since advection of energy from one layer to another is such a pronounced feature of the energy balance, it would be desirable to have a model in which the energy available for melting could be distributed between layers. The energy transfers for each layer can be summarized in the following way (ignoring some terms which will be negligible):

<u>Radiative transfers</u>

- Incoming solar radiation (some transmitted to layers below the surface)
- Reflected solar radiation
- Net longwave radiation

<u>Convective transfers</u>

- Turbulent convection at surface and through air spaces
- Advected heat of precipitation (temperature differences between the snow and any precipitation which falls on it during melting represent a difference of energy, just as the precipitation itself represents a difference of mass)
- Advected heat of meltwater (the advection of latent heat by percolation, as described earlier)

<u>Conductive transfers</u>

- Gross molecular conduction between snow layers divisible into:
  - conduction through snow grains
  - conduction through air spaces
  - conduction through pore water
  - conduction through ice layers
- Conduction between basal snow and ground surface

Evaporative transfers

Latent heat of evaporation, sublimation and condensation.

2.2.2 Approaches to Energy Balance Modelling

2.2.2.1 A Single-layer Model Clearly a model incorporating expressions for all the terms summarized above would need a considerable amount of data for each layer in the snowpack, and it is impractical to think of implementing such a model. A simpler model based on an understanding of the energy balance might, however, still be possible if the snowpack were simplified so that it consisted of a single layer. In this form the energy balance would be written

$$M = R_n - H - \lambda E - S, \qquad (2:1)$$

where

M = energy available for heating or melting snow,

R_n = net radiation at the surface,

H = convective ("sensible") heat transfer at the surface,

λE = evaporative ("latent") heat transfer at the surface,

(λ = latent heat of vaporization,

E = mass of water vaporized),

S = conductive heat transfer at the base of the snowpack.

Eq. 2:1 is the basis for many analyses of energy transfer at natural surfaces; with M and S combined into the single term G, it is the same as eq. 3:10, which is used in a different context, namely an attempt to calcualte monthly evaporation at Resolute. In that context, however, G is estimated more crudely: it is not subdivided, and it is not the aim of the model as it is in this context. G is required in the expression used for evaporation in sec. 3.3, but a knowledge of evaporation is not sufficient in this section to determine M. Net radiation, R_n, can be obtained from published records, but a more sophisticated way of finding S, and differentiating it from M, may be required here than in sec. 3.3. Moreover, it is necessary here to have an independent measure of at least one of H or λE.

One of the standard approaches to solving eq. 2:1, well-exemplified by E. A. Anderson (1968), is the "Bowen ratio solution". The Bowen ratio

$$\beta = H/\lambda E \qquad (2:2)$$

can be approximated satisfactorily by

$$\beta = \gamma(\Delta T/\Delta q), \qquad (2:3)$$

where

γ = the psychrometric "constant" ≈ 57.3 N m^{-2} at 0°C,

$\Delta T, \Delta q$ = temperature (T) and specific humidity (q) differences between the surface and the height of measurement in the atmosphere.

If either H or λE can be found independently, the only remaining unknown in eq. 2:1 is S. It is usually easier to estimate λE: for example, Anderson chooses the expression

$$E = f_u \Delta e, \qquad (2:4)$$

in which f_u is a function of wind speed and Δe is the vapour pressure gradient between the atmosphere and the surface. The windspeed function can be found statistically from comparisons with direct measurements of evaporation by pan or lysimeter; atmospheric vapour pressure can be measured directly, and it is safe to assume that the snow surface is saturated and therefore has the saturation vapour pressure appropriate to its temperature. This simple procedure has drawbacks even when snow temperature is known - which is not so at Resolute - and when there are direct measurements of evaporation as controls - which is so at Resolute, although the measurements are

not trustworthy. Eq. 2:4 is therefore not applicable in the present study.

The conclusion to which this discussion tends is that at present a full energy balance solution is not possible for snowmelt in basins such as that of the Mecham River. At the minimum, the addition of snow temperature measurements to routine meteorological procedures would be required before energy balance methods could be applied in hydrologic analyses of the prelude to streamflow and its onset. But still simpler methods are available. Much work has been done, in practical situations, using so-called "index methods" in the simulation of snowmelt. The "indices" are quantities calculated from temperature or other measurements to give a measure or index of the energy entering the snowpack. These energy estimates can be converted to equivalents of meltwater or can be related statistically to direct observations of melt production or, if these are available, to observed snowmelt discharges in streams.

2.2.2.2 Indices of Ripening and Melting A number of such indices can be constructed from the data collected at Resolute. Although they cannot give a full account

of the disposition of energy during the period before flow begins, they may at least serve as guides in, for example, predicting the start of flow. Figure 2:2 illustrates the fluctuations of several important meteorological variables during the ripening and early melting of the snowpack of 1972. This year was chosen because it is the only one of the five with a discharge record for which the early days of flow can be charted with some confidence. In 1971 and 1973 the Mecham River was already flowing when the research season began; there is no continuous record for 1970 until well after the snowmelt peak, and the first few days of the record in 1974 are difficult to interpret in detail because there is no eye-witness account against which to check the chart fluctuations.

During spring 1972 the last substantial precipitation event to occur before the stream started flowing was on 14-15 June, when there was a snowfall of 6.1 mm water equivalent. Heat energy introduced to the ripening pack by fresh precipitation can therefore be neglected, as can major additions to the mass of the snowpack. The once-daily (morning) readings of snow depth show a steady decrease beginning the day after the last large snowfall; snow depth fell from 0.75 m until the last snow lying on the ground was recorded on

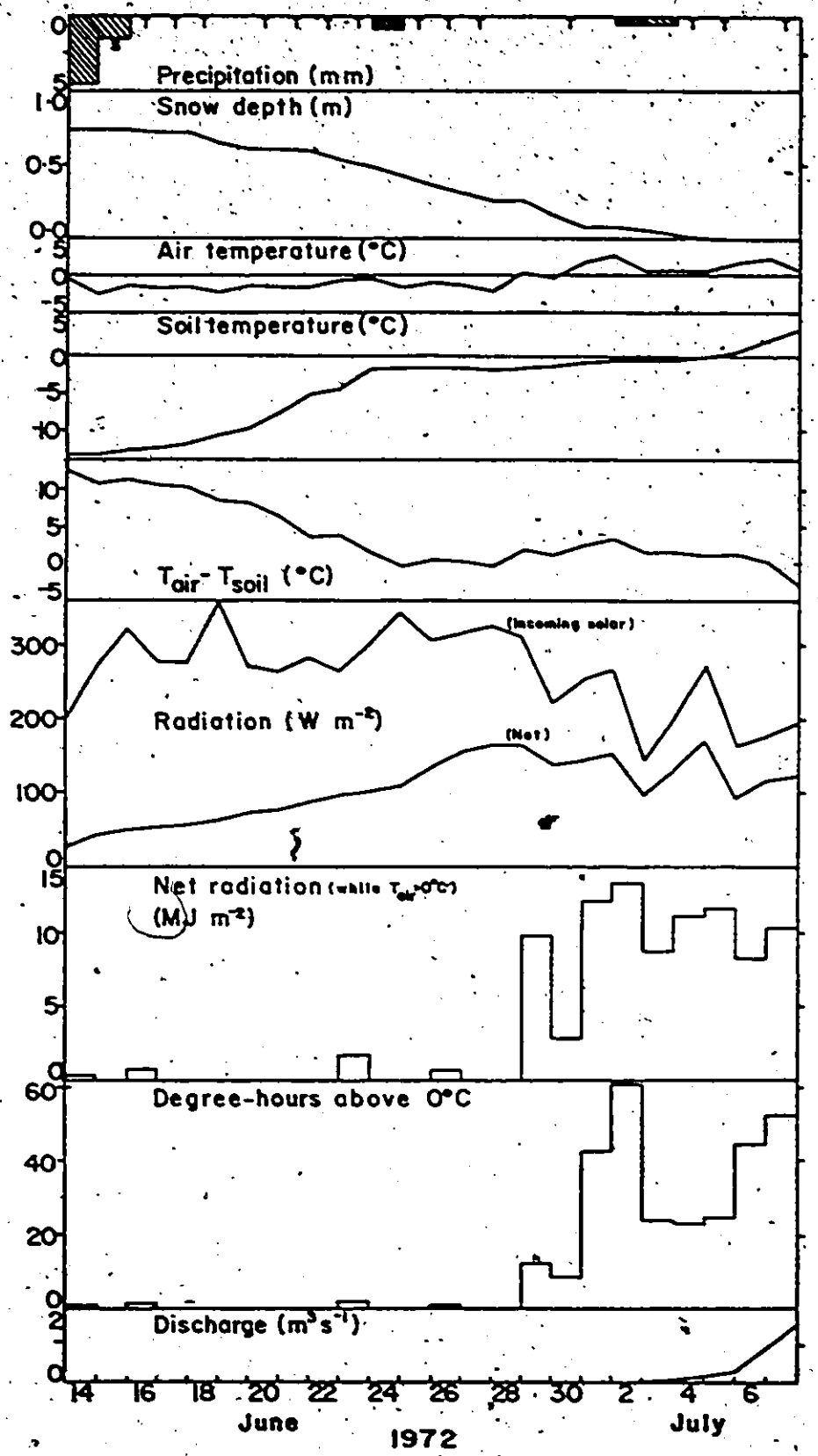


Figure 2:2. Meteorological variables before the start of flow in 1972.

4 July, the day on which water first flowed in the Mecham River.

Readings of snow depth are potentially of great usefulness, but for hydrologic purposes they are truly useful only when supplemented by readings of snow density, so that the water equivalent of the snowpack can be calculated. The decrease in snow depth shown in Fig. 2:2 is due in part to drifting, in part to rapid compaction of newly-fallen snow and less rapid compaction of older snow, but probably in the main to a combination of melting and evaporation. It is important to be able to separate these two processes, but since it is not even possible to state with accuracy the amount of water available for them, the prospects are not good for deciding how much of the observed decrease in depth is due to each of them. Measurements along a snow course of ten stakes are made regularly twice a month at Resolute, and the published snow course data include a figure for the water equivalent of the average snow depth. Unfortunately these regular measurements were suspended during June 1972, but substitutes are available: data on depth and density were collected within the basin of the Mecham River beginning on 11 June. These will be discussed below, but it should be noted here that the official record of snow depth is a poor guide to the progress of melting. By 29 June, the first day on

which the mean air temperature was above freezing, snow depth had decreased to 0.25 m, only one third of its value on 14 June, yet in the intervening days there were only five hours with above-freezing temperatures.

The day-to-day variation in air temperature is most instructive when compared with day-to-day variation in ground temperature at a depth of 0.05 m. The transition from freezing to above-freezing temperatures in the atmosphere was relatively abrupt, occurring on 29-30 June: once temperatures rose above 0°C they remained above almost without intermission. If the graph of air temperature is reliable as a guide, production of meltwater must have begun on or about 29-30 June and have continued uninterruptedly from then on, so that completion of the ripening process and the attainment of liquid water capacity required only five or six days. Temperatures in the uppermost layer of the ground stayed below freezing until 5 July, the day after the last snow disappeared from the station. They were, however, within two degrees of the melting point from 24 June onward, having increased steadily from -13.2°C on 14 June. The early steep portion of the soil temperature graph, considered in comparison with the air temperature

graph which remains between 0 and -3°C, suggests warming of the snowpack to a state of near-ripeness by about 28 June.

This suggestion is reinforced by the graph of temperature difference between ground and air, which narrowed until there was a gradient of less than 1°C from 25-28 June. It is reasonable to assume on this basis that the temperature of the snow was close to the mean of ground and air temperatures at this time: -20°C, for example, on 28 June. If snow density is taken arbitrarily to be (say) 400 kg m^{-3}, then the energy required to ripen the 0.25 m-deep snowpack of 28 June would be

$$M_r = -\rho C d T, \qquad (2:5)$$

where ρ = snow density (kg m^{-3}),

C = specific heat of ice = 2.09 J kg^{-1} K^{-1},

d = snow depth (m),

T = snow temperature (K).

The solution of eq. 2:5 is 424 kJ m^{-2}. The energy required to melt the ripened snow would be

$$M_m = \rho d \mu \qquad (2:6)$$

where μ = latent heat of melting = 335 kJ kg^{-1} and M_m thus becomes 34 MJ m^{-2}.

M_r is thus much smaller than M_m. It would be

satisfied by an energy flux of only 5 W m^{-2} for one day, and to judge from the incoming solar radiation and net radiation plotted in Fig. 2:2 it can probably be neglected. Whether this holds for other years besides 1972 can not yet be decided. It depends on the recurrence of a weather pattern in which air temperature persists just below freezing for long enough so that the whole snowpack reaches equilibrium with the atmosphere above it.

One point which must be raised in connection with the data on radiation concerns their representativeness. The disparity between the curves for incoming and net radiation arises from a progressive decrease in albedo of the surface beneath the radiation instruments, from 0.64 after the snowfall of 14-15 June to an average of roughly 0.26 from 29-30 June onwards. The latter figure is indeed the average for exposed, snow-free ground at Resolute, and gives grounds for a suspicion that the radiation measurement site lost its cover of snow some days before the snow depth and temperature measurement site, from which it is several hundred metres removed. It is natural to expect a decrease in albedo as a fresh snowfall ages, but a decrease to 0.26 is hard to credit unless the snow became

very dirty. More broadly, it is impossible to generalize about the snow cover of the Mecham River basin as a whole from data collected at at least two different sites, which are themselves outside the basin. Some portion of the basin area remains free of snow most of the time, and the snow covering the remainder melts over a period of weeks, so that a radiation budget for the continually-decreasing snow-covered fraction of the basin cannot be worked out from the radiation data collected at a single site.

Whatever the shortcomings of the data record, it can be shown to be unlikely that there was a substantial flux of energy capable of melting snow before 29 June. The bar graph beneath the radiation plot shows the daily total (rather than the flux density) of net radiant energy received while air temperature was above 0°C; if the summation is begun on 29 June, the total of such radiation received by the time the last snow was recorded on the ground is 69 MJ m^{-2}, or twice the amount required (from eq. 2:6) to melt the snow. Before 29 June, however, very little energy was available for melting because temperature was generally below 0°C. It would seem, then, that about half of the net radiation

at this time was used to melt snow at the weather station, the residue being used either to heat the atmosphere or for evaporation. A more definite statement, however, is not possible.

2.2.2.3 Snow Measurements in the Mecham River Basin, 1972 Corroboration of the above tentative conclusions can be sought in the record of snowpack properties compiled within the Mecham River basin. The programme included three series of snow depth measurements along the snow course (Fig. 1:13) and several detailed samplings of density and temperature in a pit dug into a snowbank near the base camp. The pit and the snow course, though 1-2 km apart, may be considered comparable in a general sense for present purposes. Measurements spanned the period 11 June to 8 July, and Table 2:2 summarizes the results; Figure 2:3 shows the temperature profiles obtained in the wall of the pit. Average snow course depth is initially less than that reported at Resolute, but on the last day with snow lying at the weather station the snow on the course was still 0.29 m deep. Snow depth at the pit began decreasing between 23 and 25 June, the earliest of these dates being that on which the first non-negative temperature was recorded below the surface, but the corresponding water equivalent

TABLE 2:2

Snowpack Properties, Spring 1972

	Snow Course			Snow Pit		
Date	Av. Depth (m)	Wtr. Equiv. (m)	Av. Density (kg m^{-3})	Depth (m)	Wtr. Equiv. (m)	Remarks
11 June	0.47	-				
17 June			460	2.45	1.13	snow dry and powdery
21 June			470	2.45	1.15	
23 June	0.45	0.22	480	2.45	1.18	
25 June			500	2.37	1.19	several ice layers especially in top 1.25 m
28 June			540	2.32	1.26	appearance of basal ice layer, 0.04 m thick
1 July			550	2.14	1.18	basal ice 0.10 m thick
4 July	0.29	0.16				
8 July			-	1.70	-	pit flooded with 0.5 m water

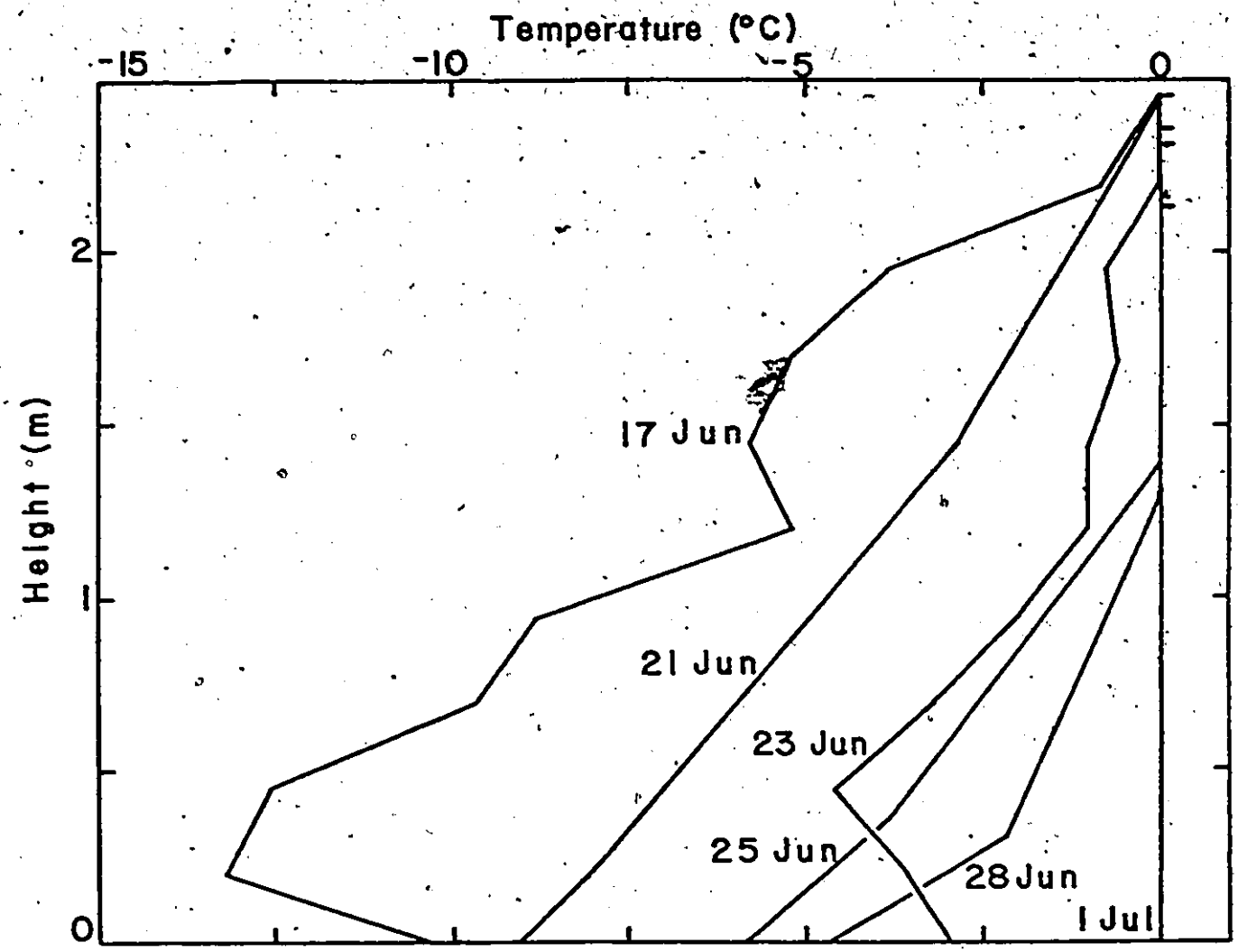


Figure 2:3. Snowpack temperature profiles, spring 1972.

increased until 28 June - 1 July, owing probably to an increase in the frequency of denser ice lenses. The appearance of these lenses, and particularly of a basal ice layer, suggests that melt water from higher parts of the snowbank was refreezing at the lower location of the pit. The lenses first appeared on 25 June, by which time the upper half of the snowpack had reached melting point; the snowpack at the pit was fully ripe and had begun to lose mass by 1 July. The date for the start of ripening at the weather station, 29 June, was thus six days later than the start of ripening at the snow pit, a finding which may not be unreasonable considering that the snowbank into which the pit was dug lay on a south-facing slope.

The natural variability of the state of the snowpack in an area of 100 km^2 should encompass easily a discrepancy of six days in the dates for the start of ripening at two different kinds of site, especially since the scale and type of the analyses used is different at each site. It remains to be seen whether the broad-scale examination of variables measured routinely at the weather station has any predictive or diagnostic value, but a final word should be said about interpretation of the water equivalents quoted for the snow course.

These equivalents were calculated by assuming that the average snow density in the pit could be extrapolated to the depth measurements taken on the course. That for 23 July is the more reliable because the depth and density were measured on the same day, and it is especially interesting because it represents the winter snowpack as equivalent to an input of 220 mm to the basin water balance for 1971-72. This will give occasion for further comment at the end of section 3. Suffice it here to say that the total measured precipitation for the water year (up to 23 June) was only 55.5 mm, and that 220/55.5 = 3.96.

2.2.2.4 Comparisons Further work is required before the beginning of flow in the Mecham River can be predicted reliably from meteorological data. The weather pattern observed in late June 1972 was observed also in June 1970 (Figure 2:4), in so far as the temperature gradient between air and soil was almost eliminated several days before flow started, the snowpack thus being brought to a state of near-ripeness. After 22 June 1970 a small rise in air temperature sufficed for melting to commence, and the river began to flow four days after the first non-negative daily mean temperature was noted.

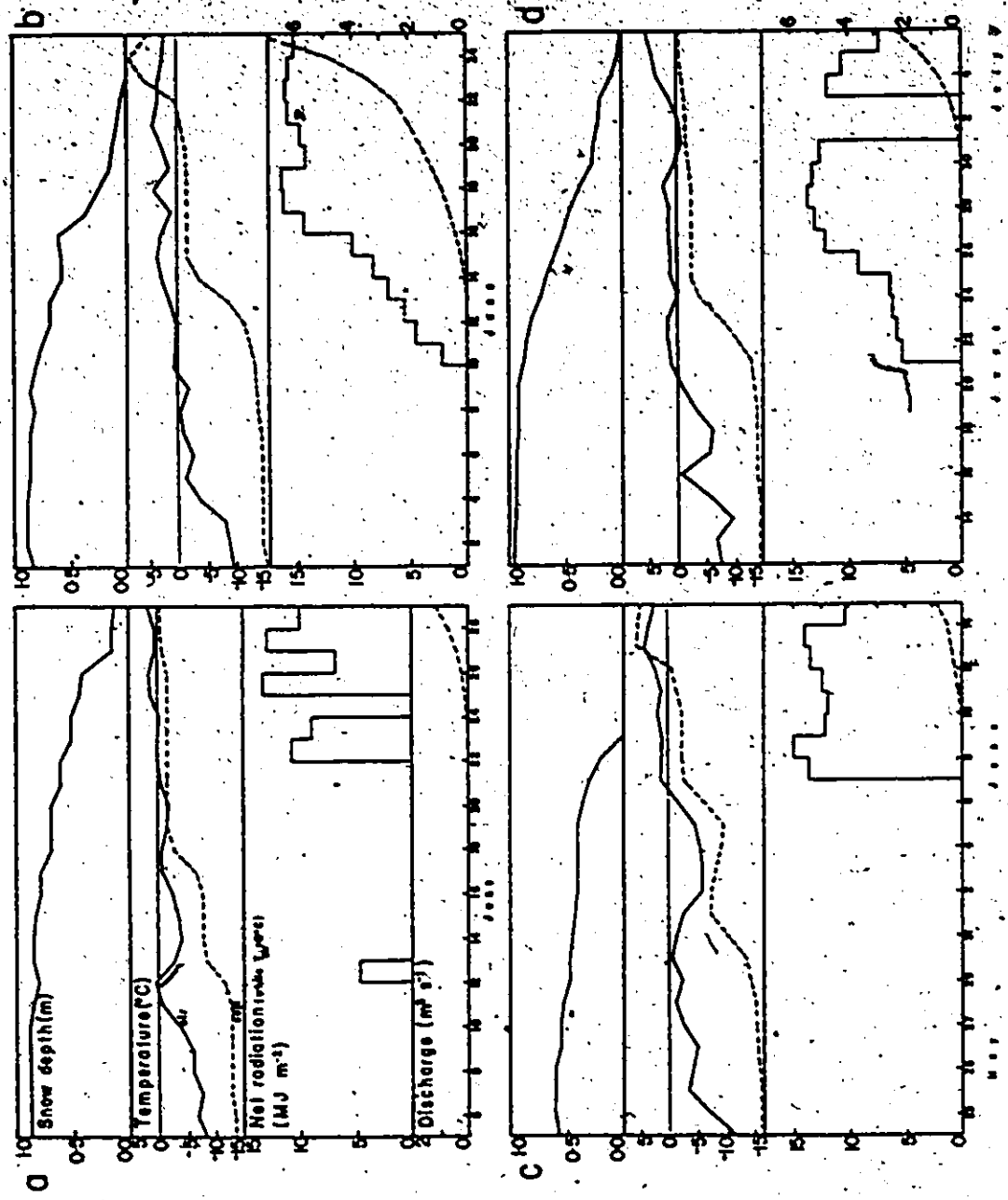


Figure 2:4. Meteorological variables in spring.
a) 1970; b) 1971; c) 1973; d) 1974.

The start of flow cannot be fixed in 1971: the date conjectured in Fig. 2:1, 10 June, is probably a few days too early, and by the rule of thumb that the year's first flow occurs fourteen days before the maximum flow (sec. 2.1.2) a more probable date is 14 June. This date is five days after the first day with a mean temperature above freezing, but since the soil was much colder than the air until about 15-16 June it is difficult to say anything about the rate at which the snowpack was heated to the melting point. In 1973 the date chosen for the start of flow is also conjectural (being estimated by backwards extrapolation of the discharge curve taken from the continuous stage record), but it too suggests that the first mean daily temperature above freezing is a good guide to the date of first flow. The lapse of time in this case was four days.

Hope that the start of flow can be predicted by counting four or five days from the rise of air temperature above freezing is, however, premature. In 1974, a full twelve days elapsed between the two happenings, and this in a year when the date of first flow is reliably known. Since it is not at present possible to derive even a simple index such as this, it would be out of place to seek a more elaborate one, given that the representativeness of the available data is less than ideal and

and that the amount of energy entering the snowpack cannot be calculated accurately.

2.3 The Snowmelt Flood

2.3.1 Ways of Modelling the Snowmelt Flood Ideally, the basis for a model of the progress of discharge once it has begun, would be the same principles of the surface energy balance as were outlined in the preceding section. However, as was seen in that section, the energy balance cannot be calculated with sufficient accuracy for these purposes. Once the snowpack is ripe, crude approximations to the amounts of daily snowmelt are the best that can be hoped for. Where the resulting discharge is concerned, an added complication is that not all daily melt within the basin flows out of the basin on the day it is produced. A variety of processes impose lags of differing durations on the passage of meltwater through the snowpack, into tributary channels and thence into the trunk stream, so that some means should be devised for routing the water from its origin to the mouth of the basin.

This problem, the successful translation of basin inputs to basin outputs, is a central concern of

much hydrologic research; the problem is too large to broach in this context, but an example of a reasonably successful model of the process is that by Derikx and Loijens (1971). This model is designed for glacier melt and incorporates a treatment of the energy balance problems mentioned above, but like most runoff response models published so far it treats its hydrologic system as a linear one. Recently, however, non-linear equations have been introduced in efforts to improve further the accuracy of simulation and prediction. The contribution of Chiu and Huang (1970) represents one such effort, and an important pair of papers by Freeze (1972a,b) arrives at the conclusion that drainage basins are intractably non-linear systems: intractable, that is, in the sense that existing linear models can hardly be improved upon unless non-linear components are introduced.

In this section the results are reported of some simple tests of regression models of the snowmelt flood in the Mecham River basin. All the tests were performed on daily data for 1972 for the reason given in the last section, namely that in the other years of record the early days of flow are not so well documented

as in 1972. The rationale of these models is that, while the full energy balance cannot be solved, it is at least possible to "mimic" the snowmelt process statistically, by regressing the observed discharges against the variables which form part of the full energy balance equation.

2.3.2 Statistical Simulation of Meltwater Production

The identification of these variables can best be proceeded with by recalling eq. 2:1

$$M = R_n - H - \lambda E - S.$$

The graphs of Figs. 2:2 and 2:4 suggest that the temperature gradient beneath the ripe snowpack at Resolute is generally very small, since the soil reaches a temperature very close to 0°C, which is by definition the temperature of the snow. It is fair, then, to neglect S, the heat flux at the base of the snowpack. Of the remaining terms on the right-hand side, net radiation may be taken as measured; in fact, however, tests show that incoming solar radiation gives slightly better results, probably because the net radiation data are for a snow-free surface.

The sensible heat flux is given (Sellers, 1965) by the expression

$$H_s = \rho C_p K_h \frac{\partial T}{\partial z} , \qquad (2:7)$$

in which ρ and C_p, respectively the density and specific heat of air, may be considered constant. The temperature gradient $\partial T/\partial z$ between the atmosphere and the surface is a function of the air temperature, since the surface temperature may be assumed constant at 0°C. The term K_h is a turbulent transfer coefficient; for present purposes it is sufficiently accurate to assume that heat transfer is dynamically similar to momentum transfer, in which case the transfer coefficient becomes a function of the wind speed. For the regression model, then, the sensible heat flux can be allowed for by including air temperature and windspeed.

The terms which comprise the expression for the latent heat flux (eq. 3:11) are discussed in more detail in section 3. Here it is sufficient to say that the relevant variables on the right-hand side of eq. 3:11, whether they appear explicitly or implicitly, are net radiation, wet-bulb temperature, windspeed and (assuming that the snow surface is saturated) the wet-bulb depression of the atmosphere, which is the difference between dry-bulb and wet-bulb temperatures. The

regression model should therefore include wet-bulb depression and wet-bulb temperature as independent variables. The latter, however, is redundant given both the former and the dry-bulb temperature. In the tests conducted, wet-bulb temperature was omitted but the relative humidity was substituted.

In keeping with the simplicity of the model, the problem of routing meltwater to the mouth of the basin is handled in a straightforward fashion. Two variables are added to allow for the influence of the previous day on the current day: the previous day's incoming solar radiation and air (dry-bulb) temperature. To simulate the general shape of the hydrograph it is necessary to have a function which expresses the tendency of discharge to rise rapidly and fall slowly. The rapid rise is due, in the most general terms, to variation in the time it takes for water originating (in this context, melting) at many points to flow to a single point, the measurement station. If the amount of snow available for melting were unlimited, the increase at the point of measurement would continue until meltwater had begun to arrive from all the contributing points. The reservoir of snow on which the stream draws is, however, finite, and its depletion is marked by a decrease in contributing area as more and more of the

basin surface loses its snow cover. A function which allows for both of these tendencies is one with the form

$$t^{\nu}\exp(-\xi t), \qquad (2:8)$$

in which t is time and ν and ξ are coefficients to be estimated.

The regression models presented here are all of the form

$$y = a\, x_1^{\,b}\, x_2^{\,c} \ldots x_n^{\,d}$$

As mentioned above, slightly better results were obtained with incoming solar radiation than with net radiation, and the models were based on the expression

$$Q(t) = \alpha . K\downarrow(t)^{\beta} . T_a(t)^{\gamma} . D_z(t)^{\delta} . RH(t)^{\varepsilon} . u(t)^{\zeta} . K\downarrow(t-1)^{\eta} . T_a(t-1)^{\theta} . t^{\nu} . \exp(-\xi t), \qquad (2:9)$$

where Q = daily mean discharge ($m^3 s^{-1}$),

K↓ = daily incoming solar radiation ($W\ m^{-2}$),

T_a = daily mean air temperature (°C),

D_z = daily mean wet-bulb depression (°C),

RH = daily mean relative humidity (as a fraction),

u = daily mean windspeed ($m\ s^{-1}$),

and the Greek letters are regression coefficients. All the variables were positive throughout the period of analysis, with t set to zero on the day before flow began. Some

computer runs were made with wet-bulb depression, relative humidity and windspeed omitted. Statistically and graphically the performance of the model thus truncated was slightly but not significantly worse than that of eq. 2:9; it is not discussed further here.

The best fit of eq. 2:9 to daily mean discharge for 4 July-16 August 1972 is shown in the upper part of Figure 2:5, and is given by

$$Q(t) = 2292K\downarrow(t)^{0.94}\ T_a(t)^{-0.25}\ D_z(t)^{1.07}\ RH(t)^{2.57}\ u(t)^{-0.56}\ K\downarrow(t-1)^{-1.02}\ T_a(t-1)^{0.36}\ t^{3.77}\ \exp(-0.31t) \qquad (2:10)$$

The (uncorrected) multiple correlation coefficient $r = 0.918$ ($N = 45$). The high discharges of the flood peak are considerably underestimated: although the predicted maximum is within 30 per cent of the observed, it comes four days too early and there is no suggestion in the predicted hydrograph of the well-marked double peak which was observed in reality. The high discharge predicted for 21 July is not a tardy version of the actual peak, but seems rather to be related to the continuation of a period of high temperatures and radiant fluxes. Discharge is overestimated from 23 July to 8 August, after which date it is once again

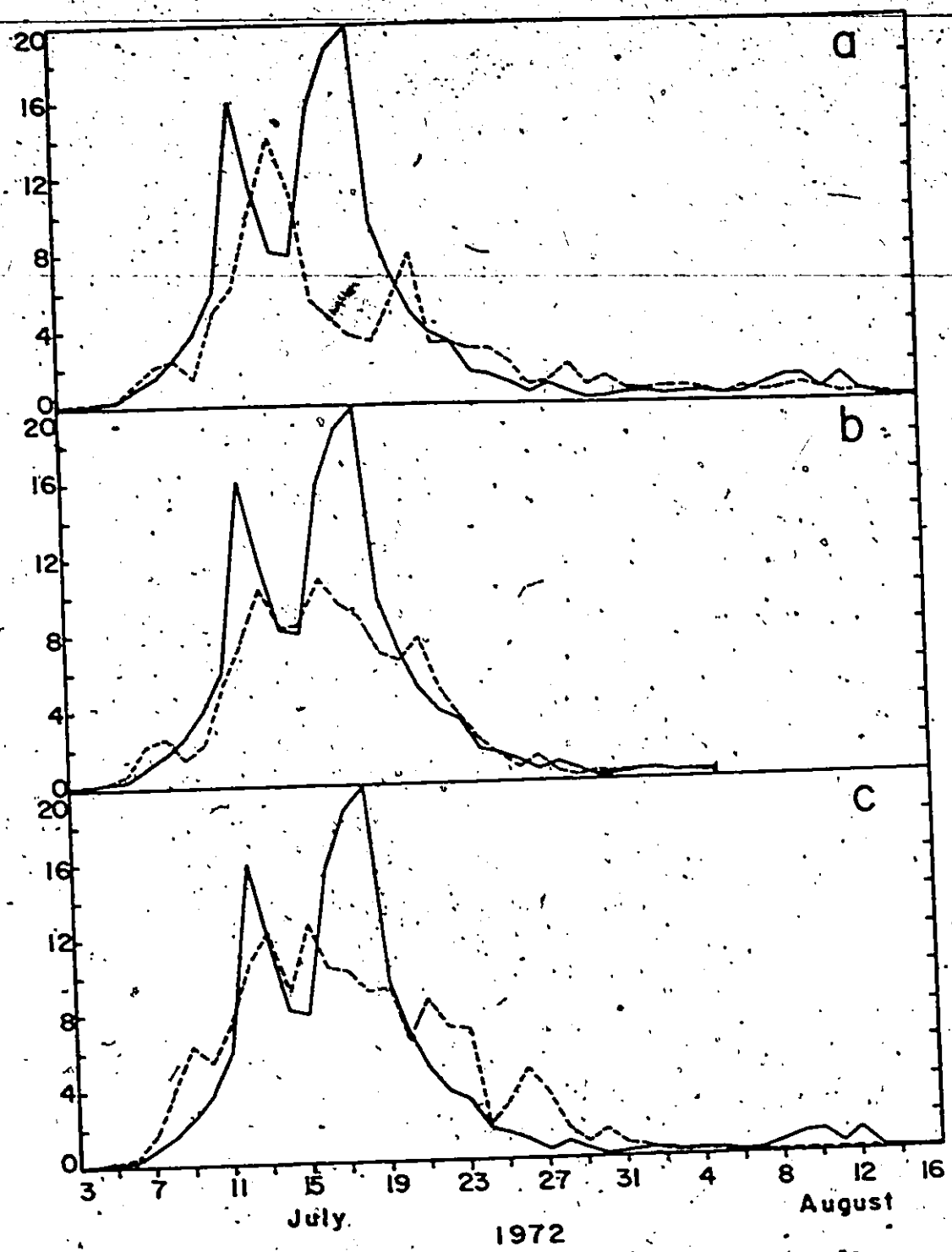


Figure 2:5. Models of the course of the snowmelt flood, 1972. a) eq. 2:10; b) eq. 2:11; c) eq. 2:15.

underestimated, although this is due partly to superadditions of rainwater runoff to the snowmelt baseflow. The fit of the model to reality is moderately good, but can hardly be considered adequate.

The middle part of Fig. 2:5 is a graph of the same model fitted to the data for 4 July to 6 August, thus excluding the days with rainfall after the latter date. The equation describing the regression is

$$Q(t) = 0.073K\downarrow(t)^{-0.12}\ T_a(t)^{0.43}\ D_z(t)^{-0.30}\ RH(t)^{-1.09}\ u(t)^{-0.10}\ K\downarrow(t-1)^{-0.39}\ T_a(t-1)^{0.60}\ t^{4.50}\ \exp(-0.40t) \qquad (2:11)$$

where r = 0.968 and N = 35. The further reduction of variance is appreciable, and so is the improvement in the simulation of the hydrograph shape. Predicted discharge is still too high on 21-22 July, but there is now a clear separation of the main snowmelt maximum into two discrete peaks, although the first of these is a day too late and the second two days too early. The principal defect of the first attempt remains, however: the predicted maxima are much too small.

Evidently the parameters of the model are not well chosen with regard to simulation of the drainage basin's dynamic characteristics. The model fails to bring sufficiently large quantities of meltwater to the mouth

of the basin in time for the spring peak, and improvement should be sought in the treatment of meltwater routing. Possibly the simplest way to elaborate this part of the model would be to introduce sub-basins with different lag times, by subdividing the basin either into its tributary catchments or into zones separated by isochrones of travel time.

Part of this problem is related to the statistical approach used. The method of least squares produces as good a fit as possible over the whole range of the observed data, and poor performance at high discharges is aggravated because of the model's logarithmic form. It is necessary to minimize logarithmic deviations at very small discharges as well as very large, but when transformed to real numbers the deviations for small discharges become negligible while the discrepancy during the flood is accentuated. If the conventional criterion, minimization of sums of squares, were abandoned or modified in favour of (say) a constraint to produce predictions of discharge within 20 per cent of observed values, it might be possible to place more confidence in the regression parameters and to seek physical interpretations of them. These interpretations, if meaningful, would give more insight into the behaviour of the drainage basin, which would in turn lead to further improvement

of the model.

The lower part of Fig. 2:5 shows the results of a test in which an attempt was made to calculate meltwater production as a function of net radiation. The rise and recession coefficients of the hydrograph were simulated, as above, with the expression given in 2:8, and the only other independent variable was an estimate of the total net radiation entering the snowpack. To get this estimate from the radiation data at the weather station, which was free of snow, it was necessary to breakdown the radiation balance into its components. The net longwave radiation

$$L_* = R_n - (K\downarrow - K\uparrow) \qquad (2:12)$$

is different over snow and bare ground surfaces, but only in so far as the ground is warmer than the snow and therefore emits more intense radiation. At 6°C a bare ground surface would emit about 29 W m^{-2} more than a snow surface at 0°C, but the longwave balance depends not only on the outgoing radiation but also on the incoming radiation, and hence on vapour pressure and cloud cover. This error in the calculated net longwave radiation over snow must therefore be accepted. The net solar radiation over snow can be approximated by taking the observed incoming radiation and multiplying by $(1 - \alpha_s)$,

where α_s is an estimate of the albedo of snow and is assumed to be 0.6. To convert the radiant fluxes (in $W\ m^{-2}$) to the same units as discharge ($m^3\ s^{-1}$), they must be multiplied by basin area ($A = 97{,}700{,}000\ m^2$) and divided by the latent heat of fusion ($\mu = 335\ MJ\ t^{-1}$) giving

$$R_s = \frac{A}{\mu}[L_* + (1 - \alpha_s)K\downarrow] \qquad (2:13)$$

as the energy available (in principle) for snowmelt.

The dashed line in the lower part of Fig. 2:5 represents the estimate by least squares of

$$Q(t) = \phi R_s(t) t^{\nu} \exp(-\xi t) \ , \qquad (2:14)$$

where ϕ, ν and ξ are coefficients to be estimated. The best fit is

$$Q(t) = 0.001 R_s(t) t^{4.6} \exp(-0.38t). \qquad (2:15)$$

Although the predicted snowmelt flood is double-peaked, and the estimates of maxima are reasonable in comparison with the upper two graphs, eq. 2:15 still gives poor results when judged on the criterion of accurate reproduction of peak discharges. The overestimates of discharge in the period just after the passage of the flood are particularly large, and as with the other models this defect can probably be ascribed to poor representation of drainage basin dynamics. Whether or not eq. 2:13

gives a better estimate of the available energy than the earlier equations is a separate question, but one which at the present stage of analysis ought to take second place behind efforts to improve the simulation of flow characteristics.

2.4 Diurnal Variability of Discharge

It is a peculiarity of streams nourished by the melting of snow or ice that their discharge changes through the day in response to changes in the meteorological variables which melt the snow or ice. Clearly, radiation and temperature increase or decrease as the sun is rising or falling. An illustrative example of this diurnal fluctuation in the Mecham River, involving a 686 per cent increase in discharge in eight hours, was quoted in section 2:1. It was also noted there that the diurnal variability often becomes damped out, at later stages of the season, by the recession flow from rainfall responses. While they persist, however, the daily changes are not without importance. During the High Arctic spring it is much easier, for example, to cross streams in the morning than when flow is much higher later in the day, after the passage of the diurnal radiation maximum. Suspended sediment

concentrations in streams are also higher later in the day, and solute concentrations, since they generaly vary inversely with discharge, are usually higher in the morning and lower in the evening.

Figure 2:6 shows the hourly discharge of the Mecham River for the period 8-29 July 1971. Each day, a roughly sinusoidal input of energy to the snowpack generates an imperfectly sinusoidal output hydrograph with a steeper rising limb and a more gentle recession limb. These sinusoids are superimposed on the downward trend from the spring peak, although this seasonal recession is not perfectly smooth from day to day. In Table 2:3 the hours of maximum net radiation and maximum discharge, with the lag time between them, are given for the period of Fig. 2:6. Excluding 19 July, when rain early in the day may have brought forward the hour of maximum discharge, the mean lag time was 8.0 hours.

One way to quantify the intensity of diurnal oscillation is to calculate the ratio of the day's peak discharge to its immediately preceding minimum. These ratios, expressed as percentages, are listed in Table 2.4. While none of them approach the 686 per cent of 14 July 1974, on only four of the days did discharge increase by less than half, and on nine days discharge more than doubled.

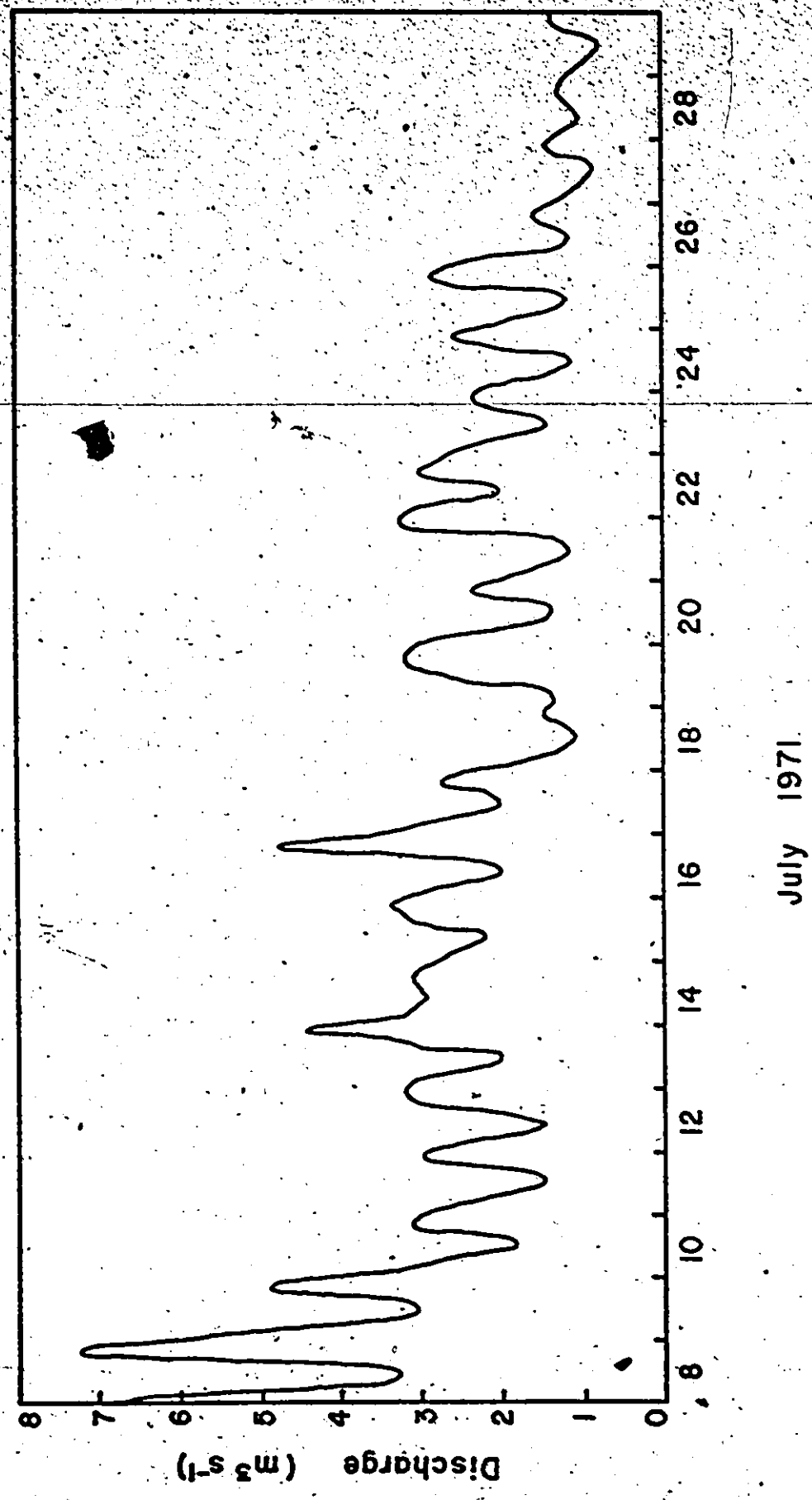


Figure 2:6. Hourly discharge of the Mecham River, 8-29 July, 1971.

TABLE 2:3

Lag of Discharge behind Net Radiation during July 1971

Date	Radiation Peak(E.S.T.)	Discharge Peak(E.S.T.)	Lag (hours)	Date	Radiation Peak(E.S.T.)	Discharge Peak(E.S.T.)	Lag (hours)
8	1230	2000	7.5	19	1200	1800	6
9	1200	2030	8.5	20	1300	2100	7
10	1000	2030	10.5	21	1200	2100	9
11	1400	2100	7	22	1230	1900	6.5
12	1100	2000	9	23	1200	2200	10
13	1300	2130	8.5	24	1230	2030	8
14	1200	1800	6	25	1300	2000	7
15	1200	2000	8	26	1400	1930	5.5
16	1200	2000	8	27	1100	2200	11
17	1400	1900	5	28	1000	2000	10
18	1200	2100	9	29	1300	2030	7.5

TABLE 2:4

Daily Discharge Ratios during July 1971

Date	Maximum/minimum Discharge (per cent)	Date	Maximum/minimum Discharge (per cent)
8	233	19	245
9	165	20	172
10	178	21	295
11	204	22	161
12	219	23	178
13	223	24	235
14	110	25	248
15	162	26	145
16	250	27	179
17	152	28	129
18	141	29	181

2.5 Responses to Rainfall

It is unlikely that rainstorm floods often or ever exceed the discharges attained during spring, the way in which the Metcham River basin responds to rainfall is nevertheless worthy of attention. No detailed analysis of hydrographs has been done to separate rainwater runoff from snowmelt runoff, but there are indications, for example in Fig. 2:1, that in at least some years most of the late-season flow is due to rainfall. The magnitude and frequency of rainstorms at Resolute is discussed in the following section; in this section some representative storm hydrographs are discussed, to illustrate the rainfall response characteristics of the basin.

Figure 2:7 shows the hydrograph of 3-6 August 1971, dominated by a flood which reached 18.1 $m^3 s^{-1}$ at 1300 h on 4 August. It is the only large rainfall response which can be coupled with hourly rainfall data: these were collected at the base camp and permit a calculation of lag time. Maximum discharge occurred slightly less than 9 hours after half of the rain had fallen, thus helping to remove any ambiguity which might remain from section 2.4 between lags of about 8 hours and lags of about 32 hours. The hourly data reveal three peaks of rainfall intensity, and these are reflected in

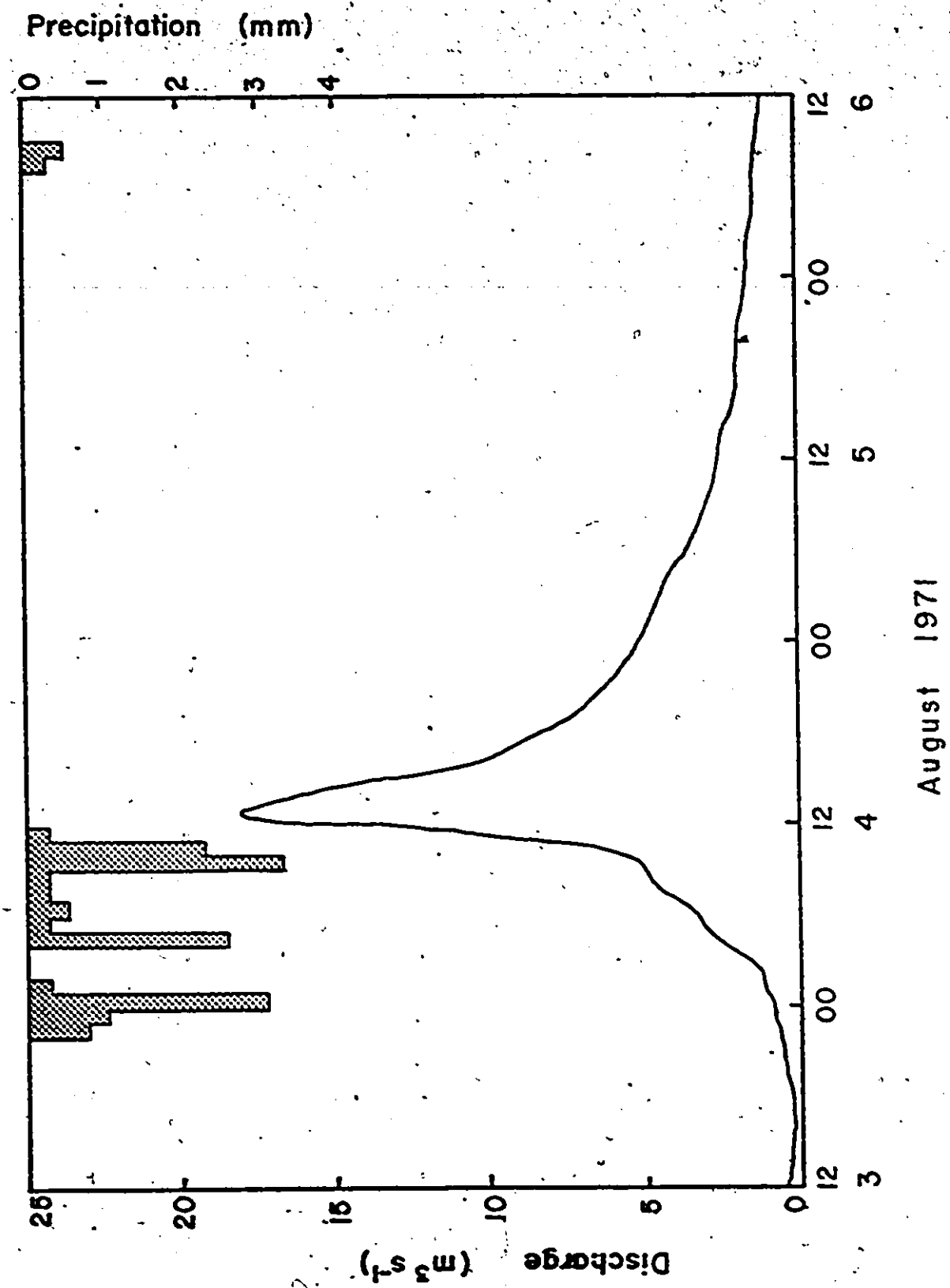


Figure 2:7. The flood of August 1971.

the hydrograph, the first two as subdued bumps on the rising limb, the third as the culmination of the flood. It would be unwise, however, to consider the three peaks as entirely separate events, for each served to increase the moisture content of the active layers and so to create new conditions for the reception of its successor. More significantly, the probable spatial variation of precipitation over the basin makes it difficult to generalize from data collected at one place near its mouth. The rainfall at the weather station 5 km away, although its time distribution is less well known, totalled 22.4 mm on 3-4 August in comparison with 14.7 mm at the base camp.

For the floods of Figures 2:8 and 2:9 the only available rainfall information comes from the weather station. Neither, however, shows any suggestion of multiple peaks. The flood of 12 August 1972 was much smaller than that of 4 August 1971, and in turn that of 11 August 1973 was larger than both of the others. The size of the discharge response bears no relation whatsoever to the amount of the rainfall: 8.9 mm of rainfall on 11-12 August 1972 led to a maximum instantaneous discharge of 2.21 $m^3 s^{-1}$, while 11.4 mm on 11 August 1973 produced a maximum discharge of 28.7 $m^3 s^{-1}$. The 1973

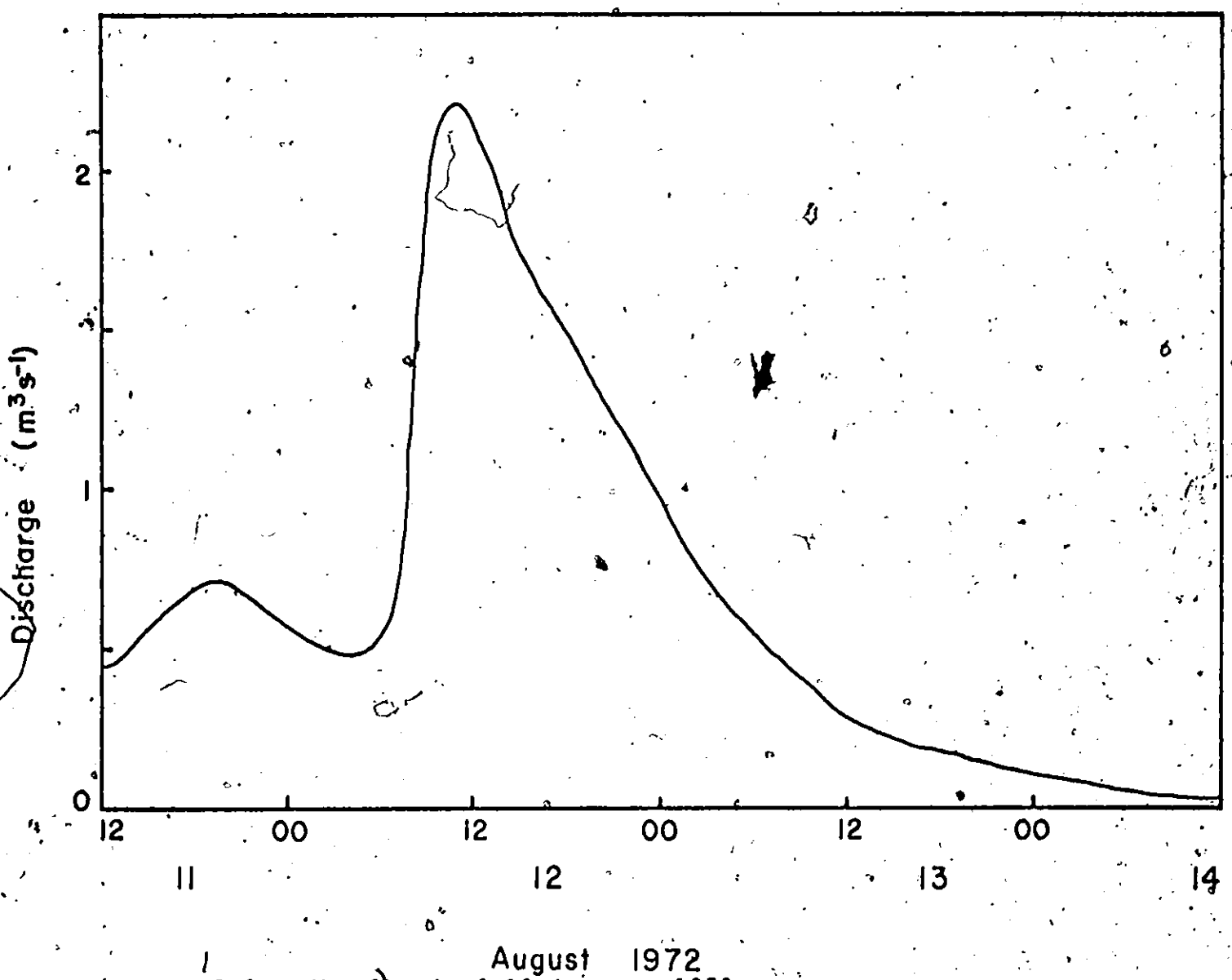


Figure 2:8. The flood of 12 August, 1972.

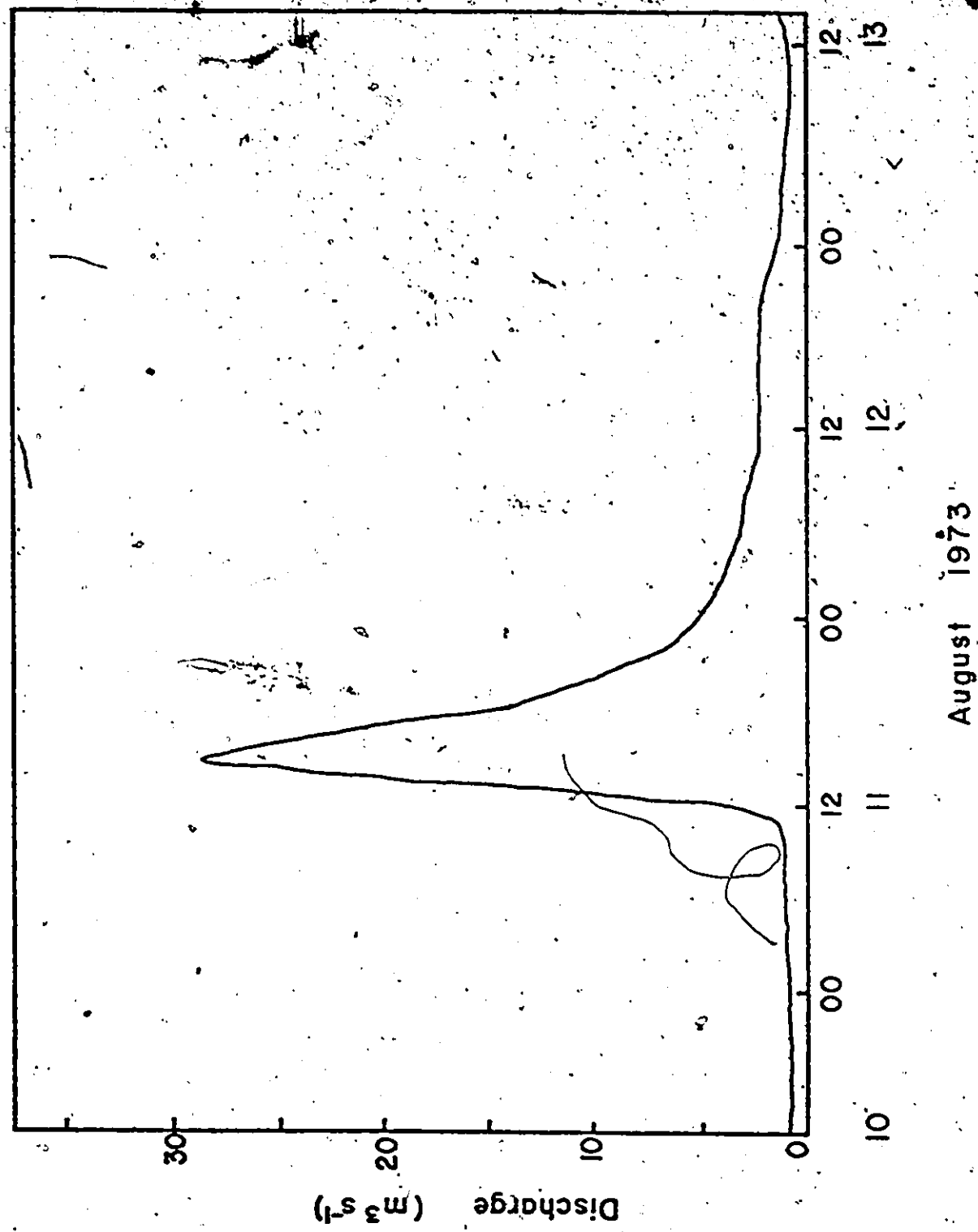


Figure 2:9. The flood of 11 August 1973.

flood thus exceeded the 1971 flood of Fig. 2:7 although it was preceded by only half the amount of rain which fell in 1971.

Probably the most distinctive feature of these three hydrographs is the rapidity of response and recession, which is a function of the shallowness of the active layer, the barrenness of the surface and the lack of lakes and other reservoirs to detain water on its way out of the basin. The rising limbs of the 1972 and 1973 floods occupied only 8 and 4 hours respectively; the storm of 1971 was evidently more protracted for the rise took place over more than 13 hours, but the response to the third of the three bursts of rainfall was rapid, only 4 hours being required for peak discharge to be reached. Table 2:5 lists the results of fitting the exponential recession

$$Q = Q_{max}\exp(-t/t_*) \qquad (2:16)$$

to the receding limbs of the three floods; Q is discharge ($m^3\ s^{-1}$), t is time in hours, the hour of peak discharge Q_{max} being numbered 1, and t_* is a recession coefficient.

The values found for t_*, ranging from 11 to 19 hours, are smaller than many recession coefficients for streams in more temperate latitudes, and also in lower latitudes of the Arctic. Church (1974), for example,

TABLE 2:5

Flood Recessions of the Mecham River

Date	Q_{max} (m^3s^{-1})	t_* (hours)	$\frac{1}{t_*}$	std.error of-($1/t_*$)	r^1	N^1
040871	18.1	19.0	-0.053	0.008	-0.971	47
120872	2.22	11.7	-0.086	0.005	-0.999	47
110873	28.7	14.1	-0.071	0.012	-0.944	41

1: r = coefficient of correlation, N = sample size

has collated information for streams in Alaska and elsewhere: a small marsh tundra basin with many ponds (Brown, Dingman and Lewellen, 1968) had t_* between 50 and 160 hours; for a forested basin with discontinuous permafrost and a thick mat of mosses (Dingman, 1973b), t_* ranged from 19 to 77 hours, averaging 39 hours; while for a mixed forest-lichen heath basin (Anderson, 1973) the range was 61 to 105 hours. The greater harshness to vegetation of the climate at Resolute is apparently of some significance in determining hydrologic responses. It must be remembered, however, that the barrenness of the Mecham River basin is due in no small measure to its limestone and dolomite bedrock.

2.6 Variations in the Long Term

A besetting problem of all investigations in the earth sciences has to do with the scale of time. Empirical research into such things as the rates of geomorphic processes, the extent of discrepancies in the measured water balance, or the occurrence of floods, must, when it does not rest on theory of proven generality, be accompanied by an answer to the question, How representative are the results? It does not matter greatly if a year of observation is an extreme year or an average year, so long as it can be placed in perspective with reference to all the other years which pass unobserved or unanalyzed. Nor does it make a difference if the period of observation be one year or five: five years is a very short time.

In this subsection the data presented earlier, and those which are the raw material for the following sections, are considered in the context of the 27-year meteorological record from the Resolute weather station. Even 27 years is a short time, but it approaches, for example, the lifetimes of many engineering structures built to withstand floods on streams like the Mecham River.

2.6.1 Annual Patterns Figure 2:10 is a graph of precipitation for the water years 1948-9 to 1973-4. The years covered by this study are the last five in Fig. 2:10. By inspection it can be seen that the study period has included two hyetographically average years (1969-70, 1970-1), the two driest years on record (1971-2, 1972-3) and the second wettest (1973-4), thus giving a satisfactory view of the climatic range so far measured at Resolute. Most of the geomorphic research for section 4 was done in summer 1971, during the "most average" of the five water years and one which saw a single summer rainstorm of moderately low frequency: at this simple comparative level, then, there is no evidence to suggest that the geomorphic results of section 4 are in any way unusual.

There is a moderately good correlation at the annual scale between meteorologic inputs and hydrologic outputs. Figure 2:11 is a plot of total annual runoff against total annual precipitation. Clearly, precipitation is of little value as a numerical index of concomitant runoff, but the correlation between the two is good enough for Fig. 2:10 to stand as a general indicator of hydrologic happenings in recent years.

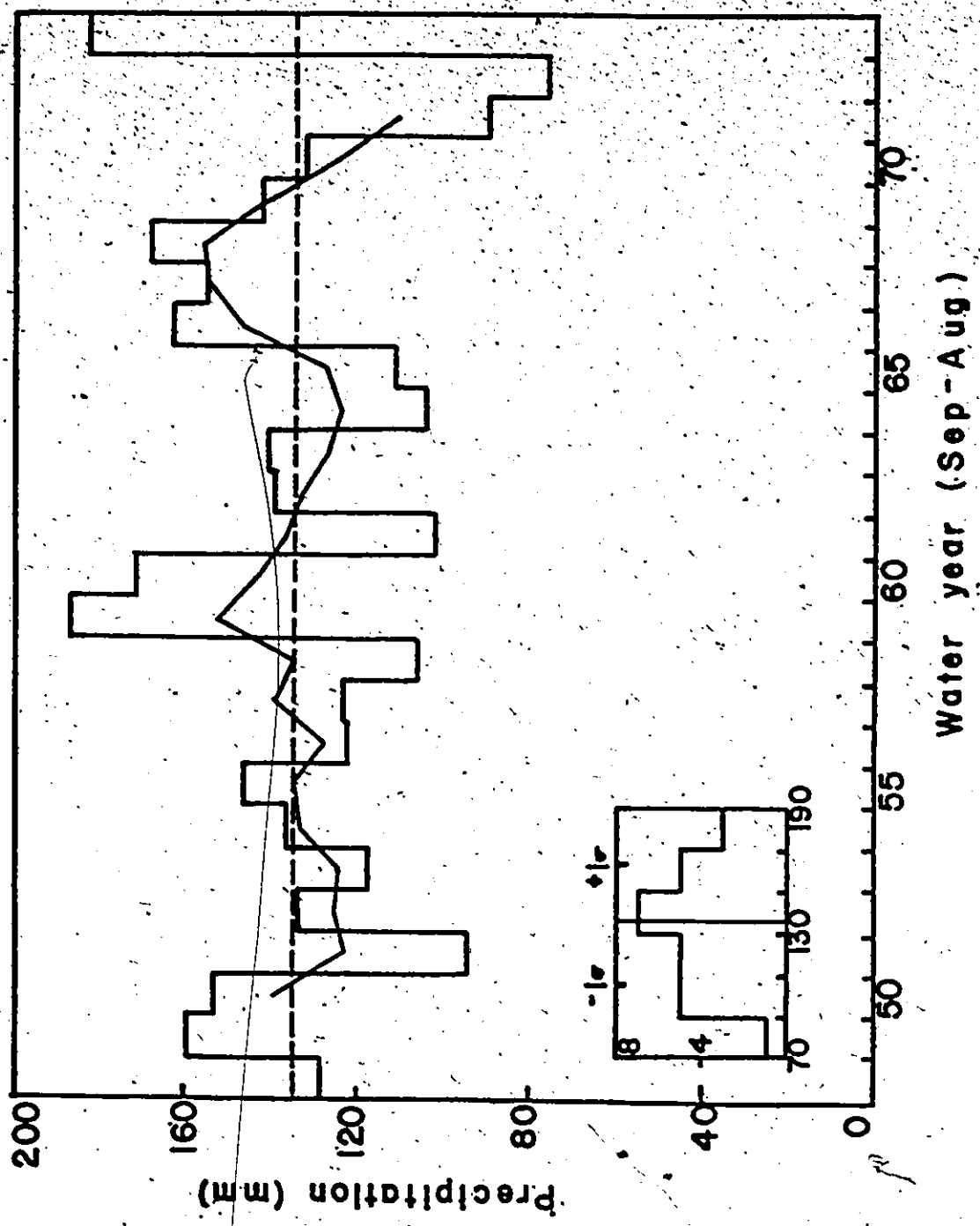


Figure 2:10. Annual precipitation, 1948-49 to 1973-74.

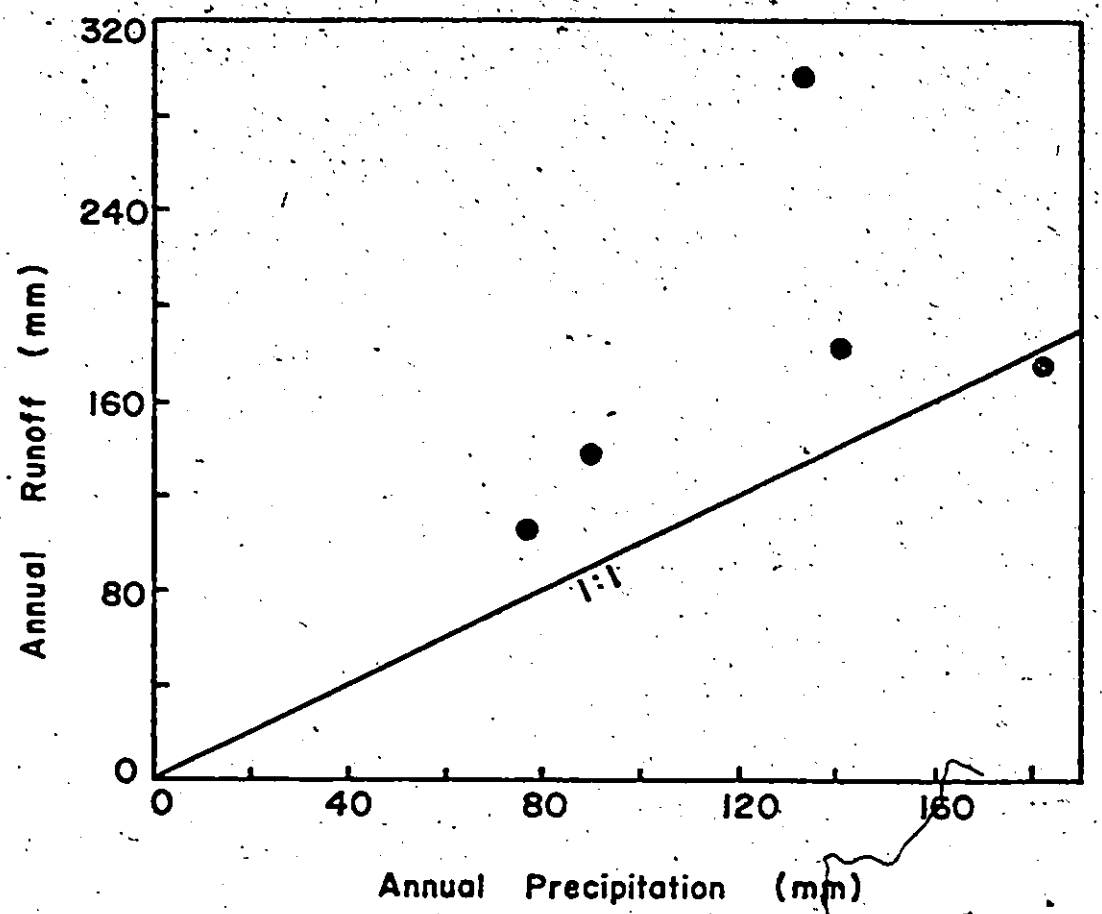


Figure 2:11. Annual precipitation and runoff 1969-70 to 1973-74.

The mean of the 26 yearly precipitation figures is 134.7 mm, with a standard error of 5.6 mm and a standard deviation of 28.8 mm (cf. inset graph, Fig. 2:10). The officially published mean total precipitation (Atmospheric Environment Service, 1971) refers to the calendar years 1941-70; since the dry years 1972 and 1973 are excluded, the official figure of 136.4 mm is slightly greater than that quoted here. Little can be said, on the basis of so short a record, about the statistical distribution of annual precipitation. The extreme years are both within 50 per cent (+31 per cent, -43 per cent) of the mean, and only one figure lies beyond the second standard deviation from the mean (76.9 mm in 1972-3 is less than the mean by 2.01 standard deviations). Either exceptionally wet years do not occur at Resolute because the large-scale behaviour of the atmsophere is stable at these high latitudes, or such years do occur but are so exceptionally infrequent as not to have been recorded yet.

If the range of precipitation so far observed in the total record is moderate, what of sequential variation through the record? Interest is growing in the study of recent trends in global and regional climates, especially those of high latitudes where relatively little climatic

deterioration may affect largely the prospects for a new ice age. Any pronounced trends in the weather record at Resolute would naturally require comment for the purposes of this study, but it is interesting to consider Fig. 2:10 in the light of recent work on the climate of Baffin Island by Bradley (1973). This work is part of a larger research effort, the conclusions of which (Andrews and Barry, eds., 1972; Bradley and Miller, 1972) are that a slight worsening of the climate over Baffin Island might induce a rapid growth of the residual ice caps there. Lowering of the snowline by about 200 m would suffice for this to come about, and the process of accretion might well become self-propagating such that a new continental ice sheet could enlarge itself by altering the climate above it. The analysis of recent records by Bradley is a contribution which suggests that the trend is indeed towards deterioration, although the results are not unequivocal. However, the global picture which has emerged from current research on climatic change is that, since about 1940, the atmosphere has cooled somewhat, in the Northern Hemisphere at least; in consequence, the ratio of snowfall to rainfall has increased and winters have become longer on average.

Baffin Island seems thus to exemplify a global pattern, but it seems also to be a particularly critical region because of the potentially catastrophic feedback which may come from an enlargement of its ice caps and snowfields. Resolute and its vicinity may be less important for the climatic future, but whatever the actual situation it is impossible to extract any secular trend from the data of Fig. 2:10, as is shown by the course of the line representing 5-year weighted running means of the annual values. (The weighted means are given by

$$\overline{P}_i = \frac{P_{i-2} + 2P_{i-1} + 4P_i + 2P_{i+1} + P_{i+2}}{10} \qquad (2:17)$$

where $\overline{P}_i$ is the weighted mean for year i, P_{i-2} the total precipitation for year i-2, and so on.) The years from 1951-2 to 1958-9 were, for example, on the whole dry, while in contrast 1966-7 to 1969-70 was a wet period - but the line trends neither up nor down. This has very little meaning in the context of large-scale climatic change, although it does illustrate the possibility of exceptions to a rule which is only a rough generalization. For present purposes, gross (or even minor) secular change in the climate over Resolute and the Mecham River basin can be disregarded.

An important point in the analyses of Bradley, however, was that studies of annual climate could easily mislead if the nub of the problem were an increase in seasonal extremeness. Accordingly, he turned his attention to change in the climates of the accumulation and ablation seasons considered separately. For convenience he defined these seasons as September-May (accumulation) and June-August (ablation), and this convenience is retained in Figure 2:12, a graph of the ratio of summer precipitation to winter precipitation. Roughly speaking this is the same as the ratio of snowfall to rainfall, and as mentioned earlier (sec. 1.1.4.1) snowfall exceeds rainfall at Resolute. However, because the two ratios are not exactly the same, the average of that in Fig. 2:12 is 1.14. So much precipitation is concentrated into the three summer months that the average precipitation of the nine remaining months of the year is less.

Like Fig. 2:10, Fig. 2:12 shows no secular trend. The period 1969-70 to 1973-4 includes two of the snowiest years on record, two relatively rainy years, and the "most relatively rainy" year of all, though this last was also the driest year recorded. The years of this study thus satisfy the criterion of representativeness

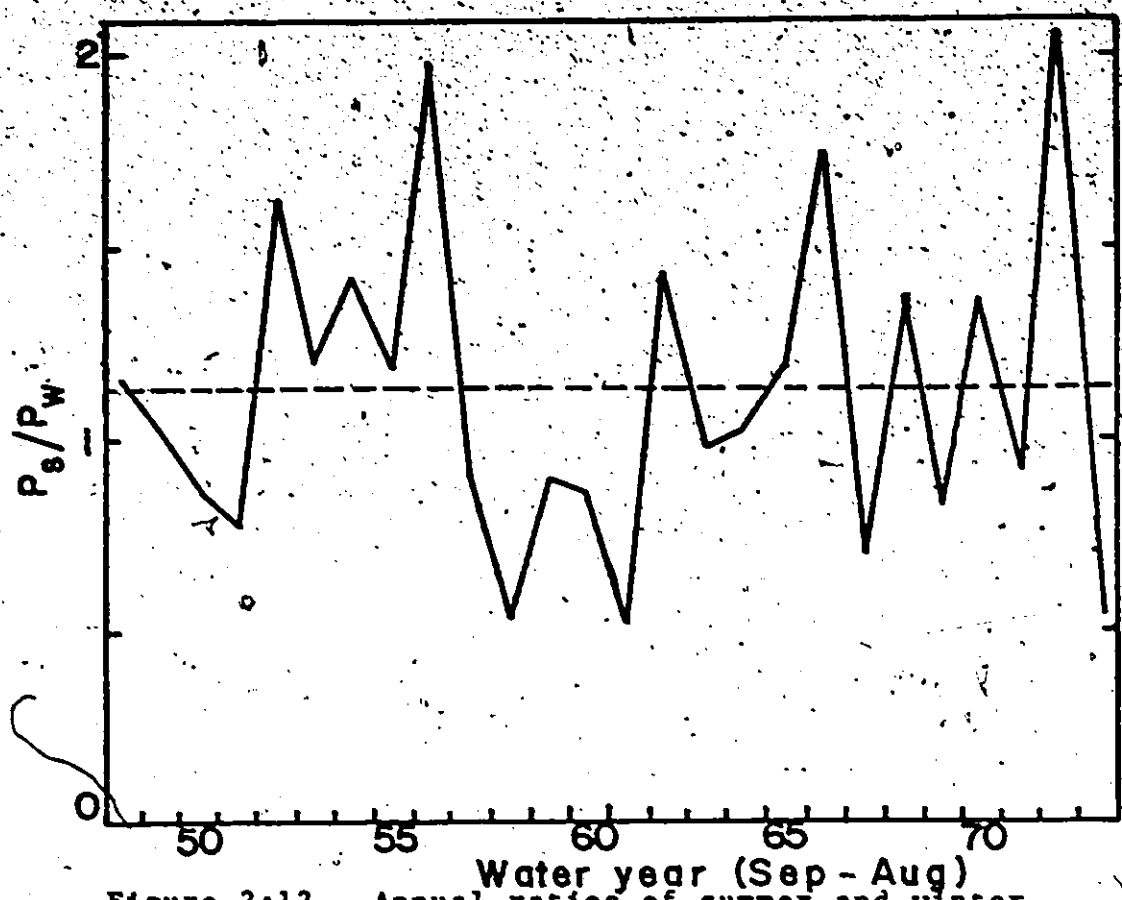


Figure 2:12. Annual ratios of summer and winter precipitation.

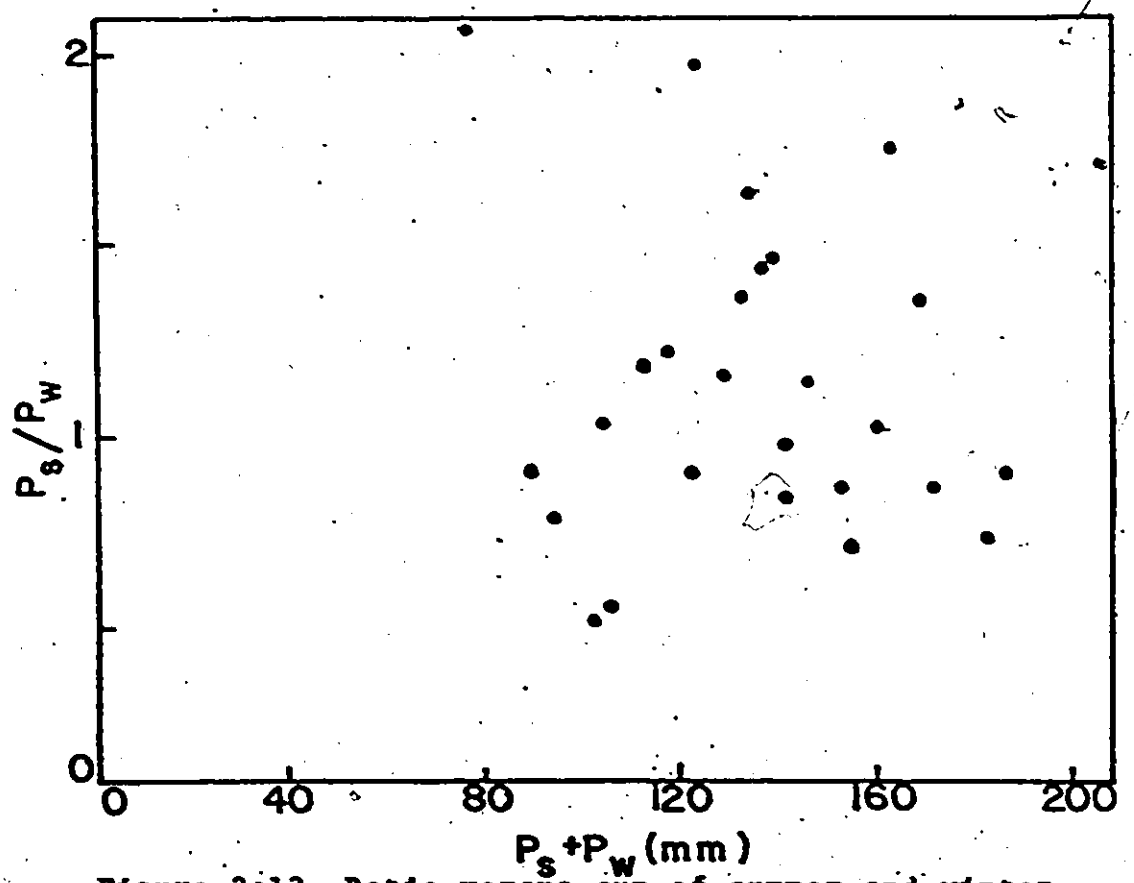


Figure 2:13. Ratio versus sum of summer and winter precipitation.

on this score as well. The early and middle fifties were rainy years, the late fifties and early sixties snowy, but to repeat: the graph shows no tendency for the climate at Resolute to grow either more snowy or more rainy with time. The numbers in Table 2:6 serve to amplify the evidence for this comment, but since Fig. 2:12 is a graph of ratios and not totals, Fig. 2:13 is offered to indicate that there is no correlation of relative snowiness or raininess with wetness. In other words, the ratio of summer to winter precipitation is not a function of their sum: there is, for example, no good reason to expect that years with major rainstorms will be especially rainy years, for it is just as probable that the rainstorm will be counterbalanced by a snowy winter as that it will not.

It is appropriate here, however, to consider the possibility of "persistence" at the annual scale. If there is no tendency for rainy (or dry) summers to follow snowy winters, is there a tendency for wet (or dry) years to follow wet years? The periods of heavy and light precipitation, in the fifties and early sixties respectively, have already been mentioned. To what extent can their occurrence be regarded as a random phenomenon?

TABLE 2:6

Winter (Sep-May) and Summer (Jun-Aug) Precipitation at Resolute (mm)

Year	Winter	Summer	Year	Winter	Summer
1948-49	59.8	69.1	1961-62	67.7	35.5
1949-50	79.4	80.6	1962-63	56.9	82.9
1950-51	81.9	71.6	1963-64	71.9	69.9
1951-52	53.6	41.6	1964-65	51.5	53.6
1952-53	51.1	83.1	1965-66	51.3	60.9
1953-54	53.1	65.0	1966-67	59.4	104.2
1954-55	56.2	80.7	1967-68	91.7	64.0
1955-56	67.0	79.6	1968-79	71.9	97.0
1956-57	41.4	82.0	1969-70	78.2	64.7
1957-58	65.2	58.6	1970-71	55.9	76.7
1958-59	69.0	37.9	1971-72	47.0	43.4
1959-60	98.6	88.4	1972-73	25.1	51.8
1960-61	92.5	79.5	1973-74	105.6	77.5

This question can only be answered with a more sophisticated analysis than is possible here, and ideally with a longer data series from which to draw inferences. But a presentation of descriptive indicators is not out of place, and Figure 2:14 shows one way of illustrating the phenomenon of persistence. It is a graph of winter-to-winter change in precipitation, plotted against the total precipitation of the later winter involved in the change. The stronger the statistical correlation between these quantities, the less the persistence in the record, for persistence implies that year-to-year changes are small and not dependent (or at least not strongly dependent) on the magnitude of the precipitation. In a record with strong persistence, year-to-year change should be normally distributed about the value zero with a relatively small variance. In contrast, a very high correlation would indicate a tendency for extreme years to stand out in the record, their immediate predecessors being much wetter or dryer. This interpretation is not statistically rigorous, but in any case Fig. 2:14 is inconclusive. The regression equation describing the relationship is

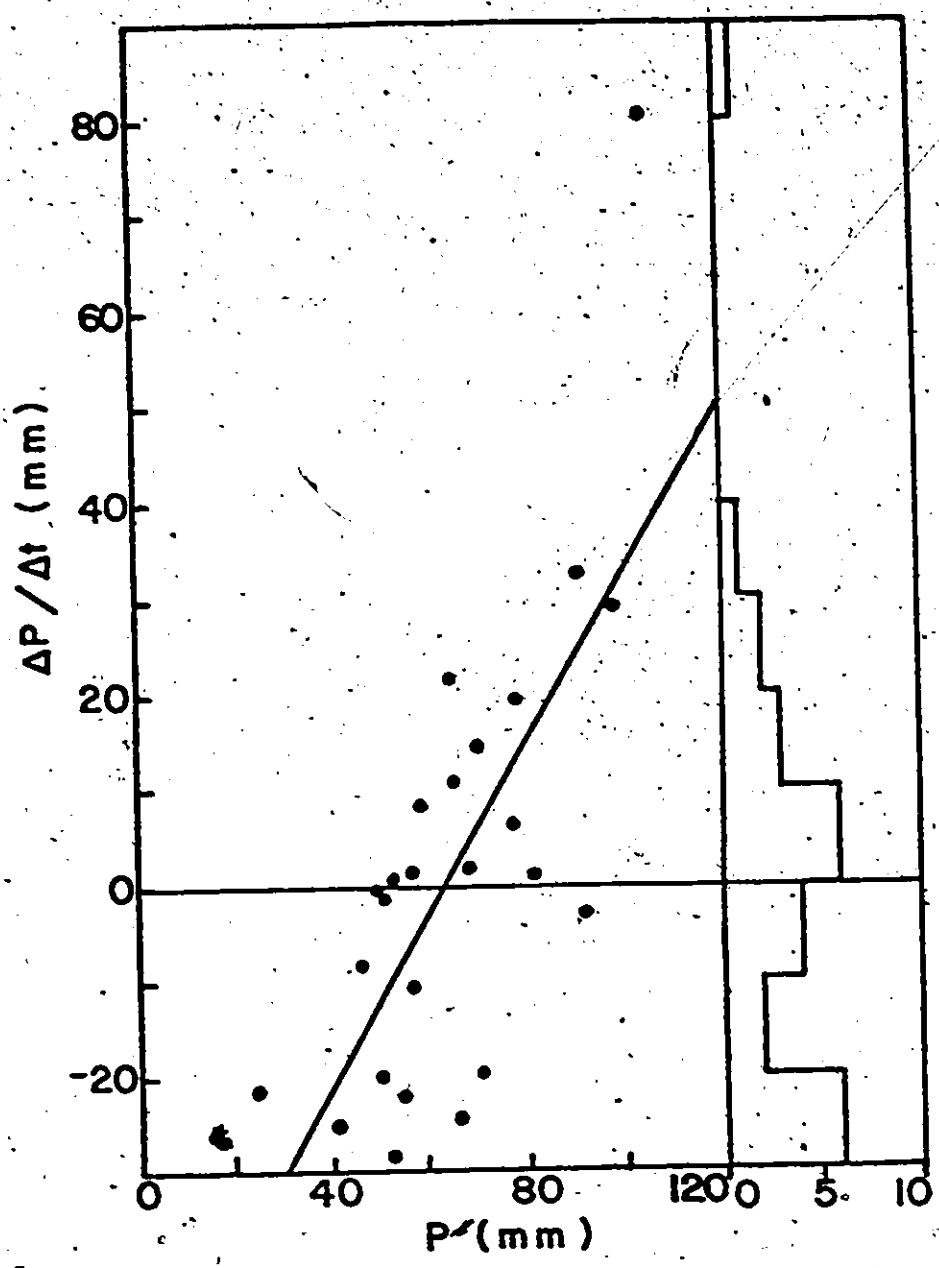


Figure 2:14. Possible evidence for or against persistence in the record of annual precipitation, 1949-50 to 1973-74.

$$\Delta P/\Delta t = 0.9P - 57.4 \quad , \qquad (2:18)$$

where the sample size is 25 and the correlation coefficient is 0.71. Eq. 2:18 is statistically significant at the 0.05 confidence level, but it is very difficult to draw any conclusions from it, or from the bar graph at the right-hand side of Fig. 2:14. The bar graph shows the number frequency of winter-to-winter changes grouped in 10 mm classes, and suggests a normal distribution skewed somewhat towards high positive changes. This, however, is misleading because winter precipitation can never be less than zero, and there is therefore a lower bound limiting the magnitude of negative changes. The phenomenon of persistence deserves closer attention than it has received, but the record at Resolute is inadequate for the purpose of determining its existence or non-existence.

2.6.2 Extreme Events One remaining long-term property of the weather record which is hydrologically and geomorphically pertinent is the incidence of extreme events. The rainstorm responses seen in Fig. 2:1 are the extreme recorded members of a hydrologic population which is closely related to the meteorologic population of discrete rainfalls, and the 27-year sample of rainfalls is much larger than the 5-year sample of floods.

In more southerly latitudes than that of the Mecham River the spring snowmelt flood becomes a less important feature of the annual hydrograph. In the temperate latitudes of eastern North America, for example, rainstorm floods are commonly the largest experienced, and the "design flood" for a given stream is often tied to the past occurrence of a severe hurricane. No rainstorm response exceeding the maximum snowmelt discharge of the same year has yet been recorded on the Mecham River, but it is reasonable to enquire about the likelihood of such a thing happening. While it is dangerous to make a simple translation of rainfall amounts into corresponding flood discharges, an examination of the rainfall record is still instructive.

First, something can be learned of the time distribution of rainstorms by increasing the resolution of the hyetograph in Fig. 1:7, so that the mean precipitation for each day of the year can be seen (Fig. 2:15). A clear cyclical pattern emerges, although there is much noise in the graph because exceptional storms have raised greatly the mean for some dates, while others have chanced so far to be dry in most years. It should be noted that almost all of the wetter days are days on which heavy rain has fallen. Snow falls in much

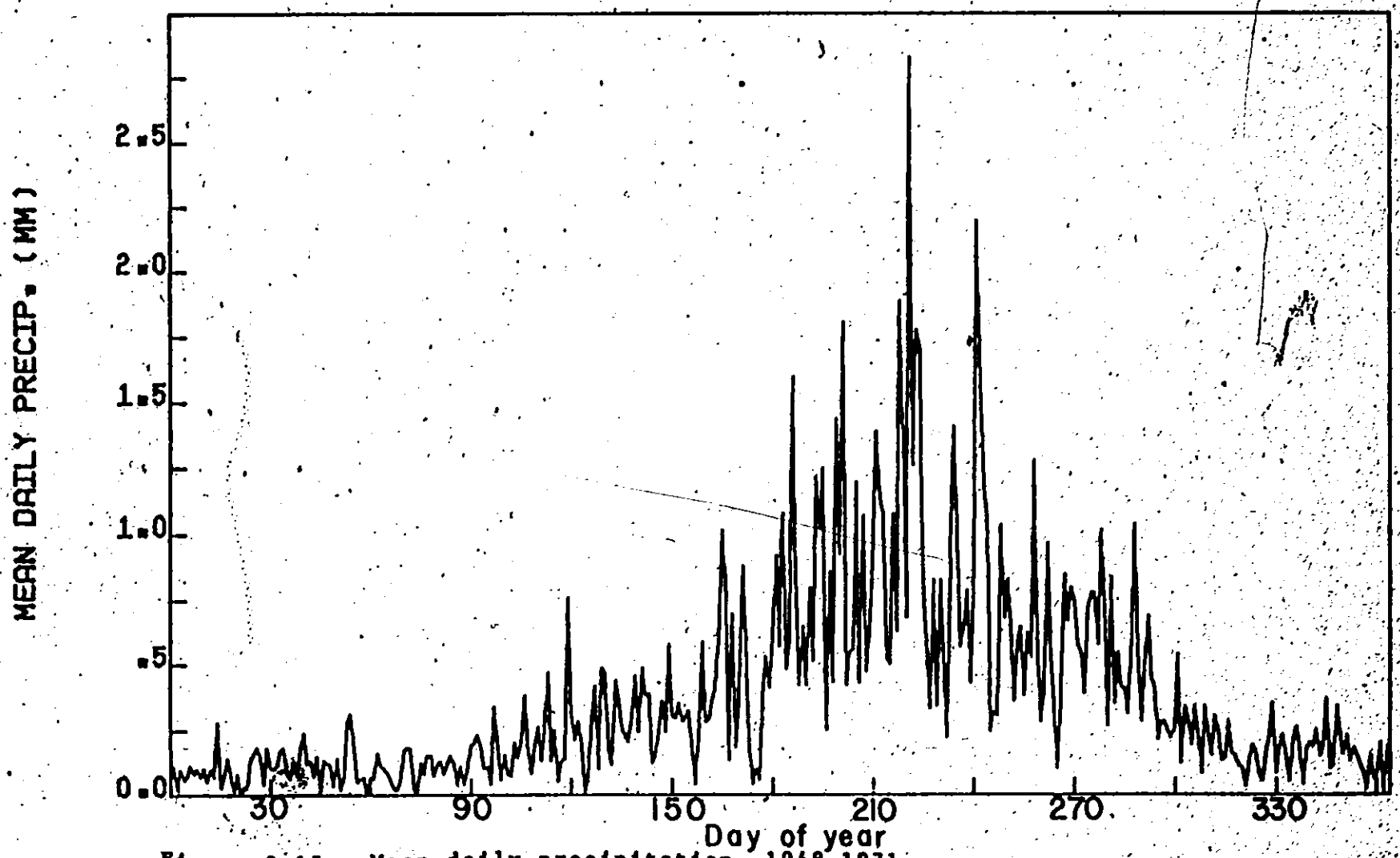


Figure 2:15. Mean daily precipitation, 1948-1971.

smaller individual amounts than rain at Resolute, and no "large" snowfalls have been measured at all in July or August (though light falls of snow are not unknown for either of these summer months).

A better organized view of the variation of precipitation with time is obtained from Figure 2:16, in which each bar represents ten days. Again, all of the taller bars stand for rainfall, and the two tallest represent the period 30 July to 18 August. The "most probable date" for heavy rainfalls is then in early August, about 8-10 August. The third tallest bar represents 29 August to 7 September; its prominence is due in considerable measure to the occurrence of 24.6 mm of rain on 29 August 1954 and 18.5 mm on 29 August 1957, and to a fall of 13.5 mm on 30 August 1970. These three dates were each the days of heaviest precipitation during the years in which they occurred, and although on average early August is the wettest time of year the date of heaviest daily precipitation has ranged from 14 June (in 1951) to 19 September (in 1960).

The greatest daily precipitation of each year is graphed in Figure 2:17. In 4 of the 27 years of record falls exceeding 20 mm have occurred, but on the

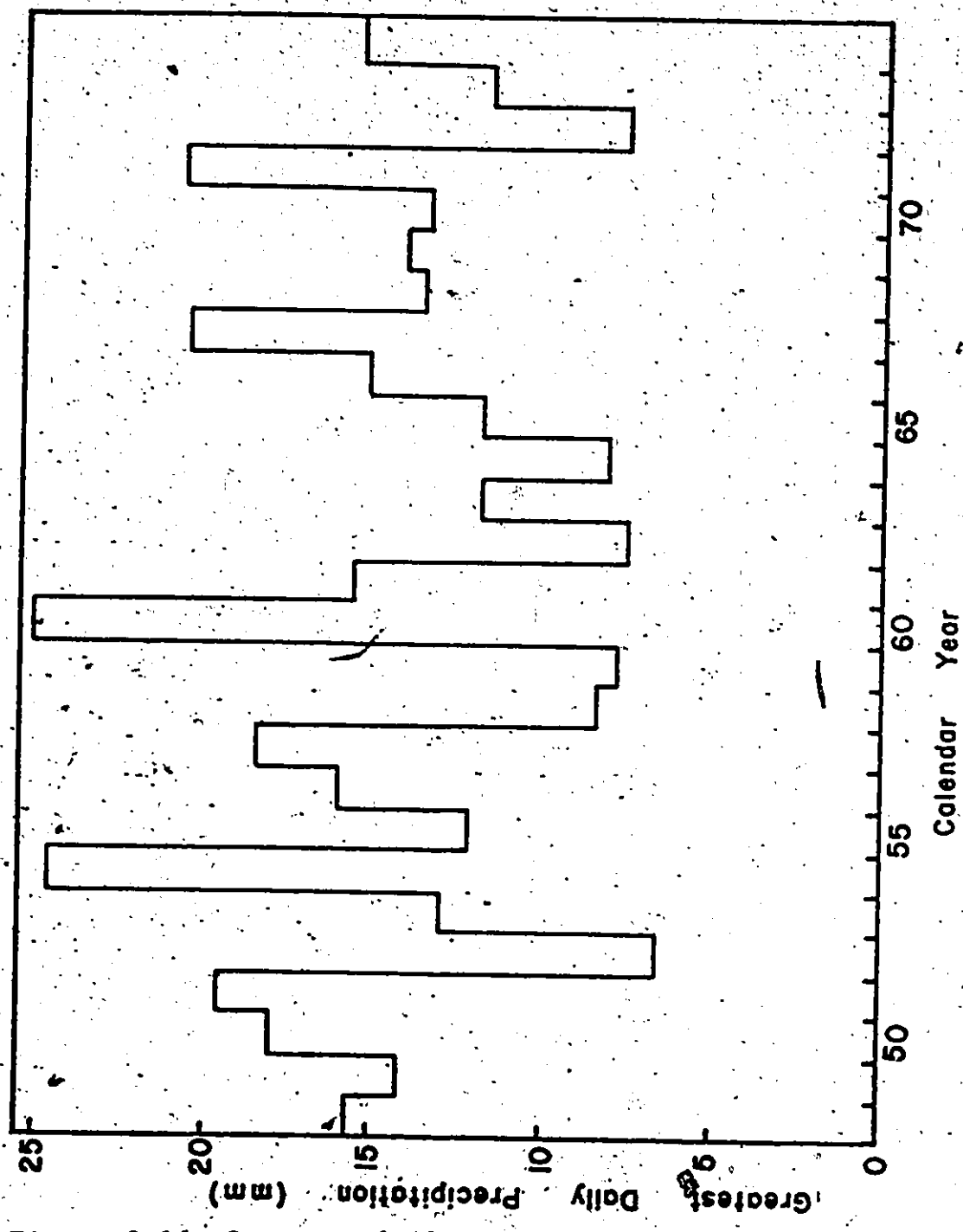


Figure 2:16. Greatest daily precipitation, 1948-1974.

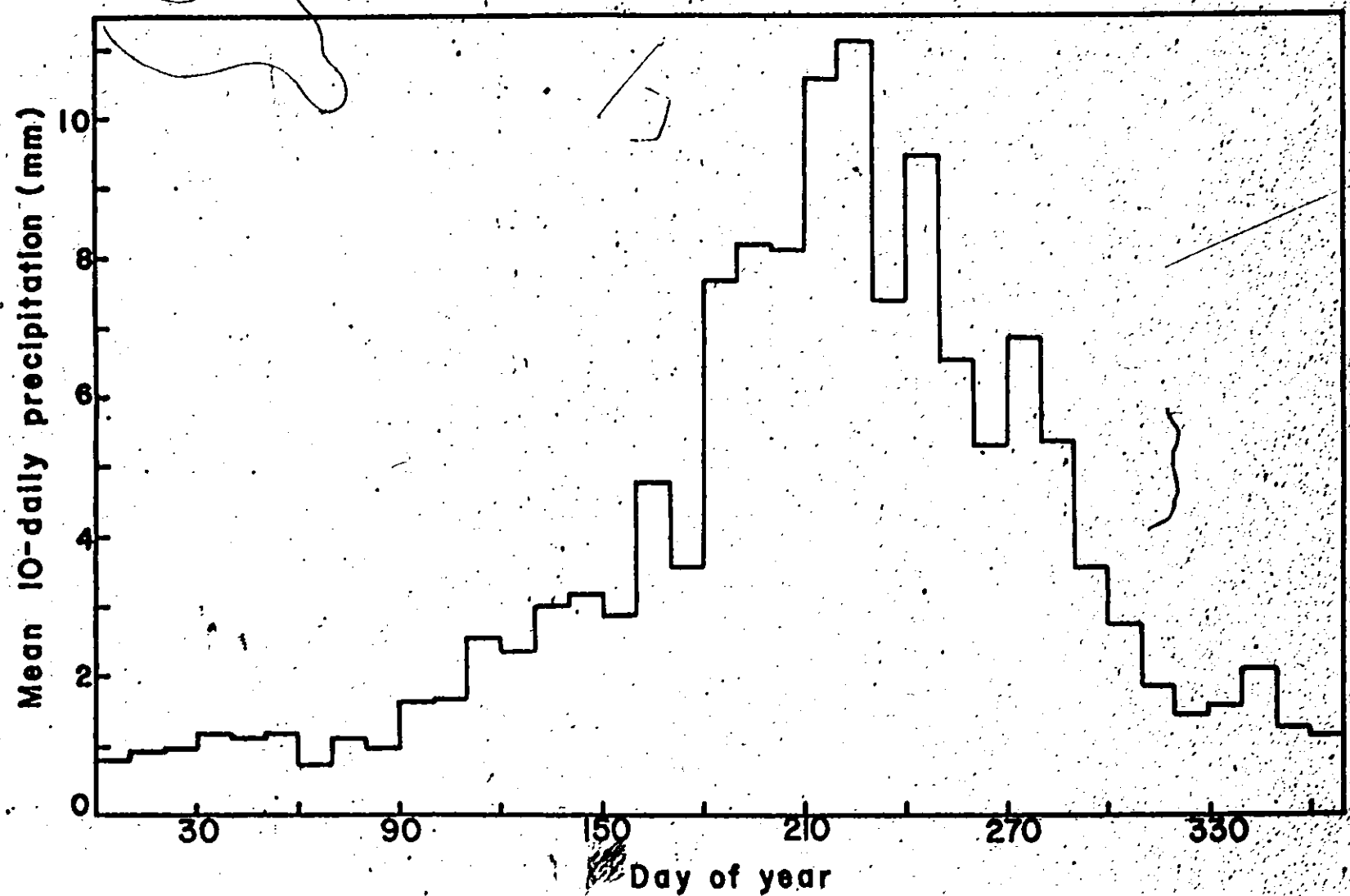


Figure 2:17. Mean 10-daily precipitation, 1948-1971.

other hand there have been six years without a daily fall greater than 10 mm. Even 20 mm is a modest amount of rain, and in fact no daily precipitation greater than one inch (25.4 mm) has ever been recorded. The mean of the values in Fig. 2:17 is 14.4 mm, the standard deviation 5.0 mm. Although a rainfall of one inch would undoubtedly produce a substantial flood on the Mecham River, its effects would probably not be devastating, as is the case with responses to the heaviest rainfalls in temperate latitudes.

The possibility of damaging rainstorm floods can not, however, be discounted altogether in the High Arctic. The data of Fig. 2:17 suggest a return period of 6-7 years for a rainfall exceeding 20 mm, and an extreme-value analysis of the data using the methods of Gray (1970), after Gumbel (1958), gives 45.2 mm as the amount of the annual maximum one-day rainfall to be expected on average once every 500 years. This amount, if it were to be discharged steadily over one day, would give a discharge of 51.1 $m^3 s^{-1}$ from the Mecham River, which is coincidentally close to the mean of the annual maximum discharges observed during this study. The instantaneous maximum discharge resulting from such an input to the basin would certainly be

much greater than 51 m^3 a^{-1}, but it is well here to remember the caution given earlier, that it is dangerous to translate rainfall into runoff in this crude fashion.

In any case, a more important point is that 45.2 mm is the 500-year one-day rainfall only at the Resolute weather station. The regional as opposed to the station return period of such a fall can be shown to be much shorter. Thomas and Thompson (1962) reported on widespread heavy rainfall over the High Arctic during August 1960, noting that a record daily rainfall of 47.8 mm fell at Mould Bay; at Eureka and Pond Inlet the record daily rainfalls are 41.7 mm and 58.4 mm respectively, and Cogley and McCann (in press) have described in some detail the hydrologic and geomorphic effects of a storm over south central Ellesmere Island which left 49.4 mm of precipitation at Vendom Fiord in one day. These four figures refer to four different rainstorms in the period 1948-1973, and although they represent local maxima within a very large region they indicate that relatively heavy rains are not rare or insignificant in the hydrometeorology of the High Arctic.

For the Mecham River only one rainstorm response is well documented geomorphically, that which occurred on 3-4 August 1971 in response to a 22.4 mm rainfall. The bed of the Mecham River consists of coarse gravel, and although there were no dramatic changes in its configuration during the flood the bed material was mobilized for the 4-6 hours of greatest discharges. This was the only occasion in 1971, apart from the times of greatest discharge during the snowmelt flood, when bed material was in motion. The effects of some much larger flood, as yet unrecorded, can only be conjectured, and indeed a good estimate of the geomorphic significance of High Arctic rainstorms in general must await both a wider survey and more intensive analyses of more individual events.

2.7 Summary

The five years of discharge record for the Mecham River reveal considerable change from year to year in the timing of the onset and peak of the spring flood, which occurs in late June or July. The year's peak discharge fluctuates moderately, the observed range being 30-60 $m^3 s^{-1}$; after the snowmelt flood there is a gradual recession through the rest of the summer, which is interrupted by occasional rainstorm

responses. The Mecham River typifies the arctic nival regime, in which the snowmelt flood is the predominant feature of the hydrograph and there is no flow in winter.

Analysis has not yet reached the stage at which the energy balance of the ripening and melting snowpack can be modelled accurately. Difficulties are experienced with the question of areal representativeness of point data, and with the prediction of snow temperature. In most years a period of 4-6 days elapsed between the first day with mean air temperature above freezing and the first day of streamflow. The difficulties of energy balance modelling extend to the prediction of discharge once it has begun. The most promising approach to discharge prediction is at present a statistical simulation of the energy balance, using as independent variables the terms which enter the energy balance equations. Results, however, leave much room for improvement, and one approach which may lead to this is disaggregation of the drainage basin into spatial sub-units.

The diurnal fluctuations which are superimposed on the seasonal course of the flood peak and its recession are substantial. During the flood proper,

discharge may increase fivefold or more in a few hours in response to the daily income of energy to the snowpack, and later in the season it is common for the daily maximum to be at least double the daily minimum. Lag time of discharge behind net radiation averages 8 hours, and roughly the same lags occur in response to rainfall. Rainstorm floods rise and decay rapidly, since the water retention capacity of the basin is small.

An examination of the 27-year weather record at Resolute shows that the 5-year discharge record of the Mecham River, and the period of study generally, are reasonably representative of what is known of longer-term conditions. No trends or persistence can be detected in the weather record, at least by simple analysis, and the observed range of variation in precipitation is well covered in the study years 1969-70 to 1973-74. Single precipitation events are not large in comparison with the severe storms experienced in lower latitudes, but it would be unsafe to discount the possibility that very infrequent rainfalls may produce very damaging flood responses.

In the analysis of the prelude to flow in spring, the snow course measurements made in 1972 were

mentioned. Because of uncertainty in the calculation of available energy, and in the routing of meltwater to the mouth of the basin, the improved estimate of the snowpack water equivalent given by the snow course data did not lead to much improvement in the analysis of ripening and melting. But there was a curiously large disparity between the water equivalent of the snow on the course and that of the snowfall measured in the preceding winter. This disparity is the opening to a wide field of inquiry in which the scale of time resolution is generally much coarser than that of the present section, and in which a clearly defined problem must be solved. The impossibility of balancing the water budget of the Mecham River at the annual scale, with presently available data, is the subject of Section 3.

SECTION 3

THE WATER BALANCE EQUATION

3.1 Introduction

The continuity equation for water in an unglacierized Arctic drainage basin with continuous permafrost may be written, for one water year, as

$$S + R + B + M_p - E - Q - I_f - I_\ell = 0 \qquad (3:1)$$

where S = total snowfall,

R = total rainfall,

B = flux of wind-borne snow through the basin,

M_p = total permafrost water melt,

E = flux of water vapour,

Q = discharge from the outlet of the basin,

I_f = increment to the "reservoir" of frozen water in the basin,

I_ℓ = increment to the lake storage reservoir.

Eq. 3:1 is a particular case of the classical continuity equation, which states that precipitation is equivalent to the sum of evaporation and runoff and which sometimes includes a storage term. Of the eight terms in eq. 3:1 each may have a special significance in one hydrological context or another. Snowfall and rainfall are differ-

entiated partly because they fall in different seasons, partly because snowflakes and raindrops have different aerodynamic characteristics (which may be influential in more than one problem) and partly because the hydrological and geomorphic results of snowfall and rainfall are different. Drifting of snow by wind is undeniably a physically distinctive flux of water, but its magnitude is not immediately certain. Where the ground surface thaws to great depth, permafrost meltwater may be a large part of the water balance, but even where the active layer remains shallow, melting permafrost can in certain circumstances contribute modest amounts of water to the surface: in either kind of area, remarkable happenings may ensue if the advance of the thawing front brings massive segregations of ground ice back to the active hydrological cycle. Evaporation and discharge are commonly recognized elements of the water balance. So also is the storage of water, and in the Arctic there are two large reservoirs of fresh water: bodies of perennial snow and ice, and lake basins.

Not all of these terms, however, need be significant in any given context, and in the following discussion of eq. 3:1, with special reference to the

Mecham basin, the lesser terms are disposed of first. The thrust of the analysis is towards an accurate estimate of the annual water balance, but it is natural to evaluate most of the terms on a monthly basis and also to portray interrelationships at a monthly scale.

3.2 Lesser Terms

3.2.1 Permafrost Melt The annual permafrost water melt can be stated as

$$M_p = (W_1 \cdot m_{act}) - (W_2 \cdot f_{act}) , \qquad (3:2)$$

where m_{act} = depth of thawing during the warm season,
f_{act} = depth of freezing after the warm season,
W_1, W_2 = volume fraction moisture contents of the active layer.

f_{act} is almost invariably equal to m_{act}, and there is no evidence of great changes in m_{act} either from year to year or on a secular scale. Again, W_1 and W_2 probably do not differ very much from each other in any one year. Take m_{act} = 0.6 m and the extreme and improbable case where W_1 = 0.4 and W_2 = 0.0: then M_p = 0.24 m, which is greater than total measured precipitation (on average

about 0.14 m). However, this sort of hydrological behaviour would be most unusual, and while admitting that data are lacking on the subject it seems reasonable to assume that on the average the active layer is in a "steady" state, with W_1 always approximately equal to W_2. In this case, M_p can be neglected.

At a time scale of single summer months and days, it can be shown that permafrost melt is always small in comparison with the concurrent snow melt. A question arises about the place in eq. 3:1 of water contained in the void spaces of the active layer: strictly, M_p for a single day is water substance which is below the permafrost table at the start, and above it at the end of the day. The water in the void spaces is certainly not embraced by this definition of M_p, and should be regarded mainly as water in transit between S (or R) and Q (or E), travelling at a speed which makes it inappropriate for inclusion in an annual balance equation.

The thawing front advances at rates such that, even if the soil were saturated at the end of the previous summer, permafrost water would be released much more slowly than water melting out of the snowpack. Repeated measurement of ground temperature profiles

(e.g., Cogley, 1971, Figs. 4:11, 12) shows that surface temperatures are very close to the freezing point at the time of disappearance of the snow cover; exposure of the ground surface reduces albedo, and heat conduction lowers the frost table rapidly at first, with rates of lowering as great as 0.05 m day^{-1}; rates of lowering decrease gradually, however, as the frost table moves further and further from the surface. M_p should be directly proportional to the rate of lowering, and the rate of lowering decreases roughly synchronously with the recession of the snow-melt flood. Permafrost melt is thus directly proportional to snow melt also, and the maximum possible permafrost melt is much less than observed amounts of snow melt.

On a monthly basis, maximum possible permafrost melt is a function of frost table lowering during the month, and of the volumetric moisture content of the active layer at the end of the previous summer. Most summers end with a slow decrease of evaporation and discharge towards very low winter values, reflecting the drying up of the inputs to the water balance; in such summers the term W_2 should be very small. Significant changes in M_p can be expected only in summers

which end with a large input of water such as a rainstorm followed immediately by rapid freezing of the entire active layer. An event like this would be unusual but not necessarily rare: Holecek (1975) appears to refer to one in his study of basin water balances in northeast Devon Island (seen only in abstract), but it is not clear whether his discussion relates to an arctic nival catchment or to muskeg terrain. In all of the years examined in this study, however, late-summer rainfall has either been inconsiderable or been followed by a period of thaw conditions sufficiently long to allow most of the rainfall to run off directly and most of the residue to drain more slowly out of active layer storage.

3.2.2 Frozen Water Storage Steady state considerations similar to those applied to M_p may also be invoked for the change in frozen water storage, I_f. (This remark is intended only for unglacierized basins which may yet contain semi-permanent snowpatches in gullies and hillslope concavities, semi-permanent snowpatches being those patches which do not melt completely in at least some summers.) I_f may be written for the

present approximation as

$$I_f = \Phi P \quad (3:3)$$

where Φ lies between 1 and -1 and $P = S + R$. Remembering that semi-permanent snowpatches cover only a small fraction of the terrain in the Mecham basin, and that no dramatic growth or decay of snowpatches has been observed, it is safe to conclude that Φ is a very small fraction. From month to month during winter I_f is in effect the solution of the monthly water balance, but for the annual balance I_f is well within the limits of accuracy appropriate to eq. 3:1.

Naturally, I_f becomes a major term of the water balance in drainage basins which are wholly or extensively glacierized, and there are many glacierized basins in the High Arctic. In general, also, the amount as opposed to the increment of frozen water storage is of critical importance in global hydrology, meteorology and glaciology.

3.2.3 Lake Storage The magnitude of lake and surface depression storage may also be considered small:

$$I_\ell = (V_s - V_e)/A_t \quad (3:4)$$

where V = volume of lake storage,

s,e = subscripts referring to the start and end of the water year,

A_t = drainage basin area.

In basins draining glaciers, the quantity I_ℓ may be vitally important if there exist ice-dammed lakes with the habit of growing for long periods and discharging their contents in episodic and catastrophic floods. In a later discussion of sediment transport, results are presented from just such a basin. In ordinary lakes in the High Arctic, however, annual changes in volume are small, and I_ℓ is only considerable where the lake occupies most or all of the drainage basin. In particular, lake surface area is very small in the Mecham basin. The role of lakes as reservoirs is therefore ignored here, as is their interesting role in delaying runoff.

3.3 Greater Terms

In the basin of the Mecham River at least three of the eight terms in eq. 3:1 can be discarded, and the result is the new form

$$S + R + B - E - Q = 0 \qquad (3:1a)$$

The remaining terms are now treated roughly in increasing order of importance and ease of evaluation.

3.3.1 Blowing Snow The phrase "flux of wind-borne snow" refers to the remobilization of fallen snow at high wind speeds. It may be expressed as

$$B = b_{in} - b_{out} , \qquad (3:6)$$

where b_{in}, b_{out} are the inward and outward fluxes of blowing snow. B would be very difficult to evaluate accurately, even with an ambitious programme of measurement. Little work has been done in the Western countries on blowing snow, but several important papers from the Russian literature have been translated into English.

One is tempted to apply the same steady state considerations to B as to M_p, I_f and I_ℓ, arguing that $b_{in} = b_{out}$ and therefore $B = 0$. So little is known about the problem, however that an effort to evaluate B may prove worth while even if the results are very crude. The magnitudes of fluxes will probably have to be determined in the near future anyway, when and if overland transportation systems in the High Arctic are extended. The following discussion of blowing snow draws mainly on the work of Dyunin (1954 a, b, c, 1959), Dyunin and Komarov (1954) and Komarov (1954 a, b). It should be stressed that

none of the analysis for the Mecham basin is based on data gathered in the field.

Komarov (1954 a), working in the low-relief terrain of western Siberia, reported the important empirical fact that the distance over which snow is transferred by the wind during winter does not exceed 2-3 km. This was determined from analyses of the effectiveness of different kinds of forest shelter belts and snow fencing systems. There seems to be no good reason why this finding should not apply elsewhere in the world. Generally, in terrain more broken than that of Siberia, transfer distances should be less than 3 km. Komarov also reported transfer efficiencies of up to 0.8 (and greater where wind speeds were above average): that is, up to 80 per cent of fallen snow was removed from open terrain by the wind. Other results obtained by Dyunin and Komarov suggest that snow transfer begins at wind speeds (at 1 m above the surface) of 3-4 m s^{-1} and becomes considerable at 7-8 m s^{-1}, and that all of the snow transfer takes place within 2 m of the surface (with 95 per cent transferred in the lowest 0.2 m).

To analyze snow fluxes across the watershed of the Mecham River, a conceptual model defining source

areas, "sink" areas and snow traps was devised and applied to maps of the basin. Source areas lie outside the basin, with their outer boundaries not more than 3 km distant from the watershed, and the term b_{in} in eq. 3:5 originates in these areas. Similarly, sink areas lie within the basin, their inner boundaries being not more than 3 km from the watershed, and they provide the term b_{out} in eq. 3:5. Traps are of two sorts: gullies and other incised watercourses, which are regarded as barriers across which no snow is transferred; and steep slopes, which inhibit snow transfer in the upslope direction only. One source area and one sink area are associated with each wind direction, and the contribution to B from any wind direction is the sum of gains from source areas and losses from sink areas; for any one wind direction the term $(b_{in} - b_{out})$ may be positive or negative depending on whether the source area is larger or smaller than the sink area. Gullies and similar traps have the effect of reducing the source and sink areas for a basin as a whole, by collecting snow which might otherwise have entered or left the basin across the watershed. Steep slopes reduce source and sink areas preferentially for particular wind directions.

In general the presence of a surface gradient will reduce the amount of blowing snow transferred upslope, and since the drainage divide is always upslope from its surroundings the efficiency of snow transfer across it is less than across an arbitrary line drawn on a flat surface. However, only the steepest and highest slopes will be able to reduce snow transfer to nil.

Komarov's empirical finding that blowing snow is transferred over distances not exceeding 3 km is not related only to the relief of the terrain and the effective fetches available in different directions. It is related also to the magnitude and frequency of winds from different directions (and in particular from the prevailing wind direction), to the amount of snowfall, and to the duration of winter, defined as the period of uninterrupted snow accumulation. In the High Arctic winter is very long and wind speeds are high, but snowfall is light and long unbroken fetches are comparatively rare. Some simple geometric considerations help to indicate the nature of topographic controls on the magnitude of the blowing snow flux.

Working with the conceptual model outlined above, it can be shown that the importance of the term B in the total water balance decreases with basin size, and also that it increases with basin elongation and with orientation with respect to the dominant wind direction. Table 3:1 gives figures for basins of different shapes and sizes; the basins are assumed to be developed on a plane surface, and their areas are compared with the areas lying within 3 km of their perimeters, (A_p/A_t) being a measure of the relative size of the wind-borne snow flux. The significance of basin orientation, not brought out in Table 3:1, is that in a basin oriented long side on to the prevailing wind direction the terms b_{in} and b_{out} may be larger than in a basin oriented end on to the prevailing wind direction. In general it turns out that the term B in eq. 3:1 may be very difficult to detect in drainage basins which are (say) much larger than that of the Mecham River; however, wind-borne snow transfer may be sufficient to transform the water balance of the basin if other features of the basin, such as its orientation, are favourable. The "design flood" for a small High Arctic drainage basin, then, may not always be found accurately by

TABLE 3:1

Geometric Data on Imaginary Basins for Snow Flux Calculations

Basin Type	Dimensions	Perimetric length, P	Circularity*, C	Area, A_t	Near-perimeter area, A_p	$\frac{A_p}{A_t}$
Circular	radius 10	63	1.00	314	377	1.200
	radius 100	628	1.00	31416	3770	0.120
Rectangular, width constant	18 x 18	72	0.78	324	432	1.333
	18 x 27	90	0.75	486	540	1.111
	18 x 36	108	0.70	648	648	1.000
	18 x 54	144	0.59	972	864	0.888
Rectangular, shape varying	18 x 18	72	0.78	324	432	1.333
	12 x 27	78	0.67	324	468	1.444
	9 x 36	90	0.50	324	549	1.667
	6 x 54	120	0.28	324	720	2.222
Square, size	18 x 18	72	0.78	324	432	1.333
	54 x 54	216	0.78	2916	1296	0.444
	180 x 180	720	0.78	32400	4320	0.133

* $C = (4\pi A_t)/P^2 = A_t/A_c$, where A_c = area of circle with circumference P.

calculating its area and extrapolating to it results obtain in larger basins.

In general, also, the effect of increasing relief (i.e., increasing ruggedness) should be to reduce snow transfer between neighbouring basins. Relief, and average elevation, increase from west to east in the High Arctic. In very rugged terrain, however, orographic influences on the distribution of precipitation are likely to mask the effects of redistribution of snow by winds. Figure 3:1 is a vertical air photograph of an area near the head of Vendom Fiord, Ellesmere Island, taken on 12 June 1972 immediately before the spring thaw. Total relief within the area is about 400 m. The difficulty of fitting a model for between-basin snow transfer to terrain of this sort should be apparent. Many gullies, concave slope sections and other localities can be seen as snow traps, but patterns of redistribution of snow are evidently complex.

Results obtained with a deductive model should therefore be treated with great caution. Most of the physical processes governing snow transfer are understood in principle, but their influences on the final value of B must be lumped together and expressed in coefficients of the measured areas of snow sources and

Figure 3:1. Vertical view of an area near the head of Vendom Fiord, Ellesmere Island.

sinks. Moreover a coarse resolution is appropriate in practice for scales of both time and area. A possible operational form of the model for calculating the flux of blowing snow is

$$B = \eta\ S\ \Sigma[\alpha_i(A_{in} - A_{out})_i],\ i = 0, 45, 90, \ldots 270, 315, \qquad (3:6)$$

where B and S have been defined above, i is a counter in degrees for successive wind directions and the A_{in} and A_{out} are source and sink areas respectively for the different directions. If S is given as a length, eq. 3:6 yields B as a volume; to obtain B as a length, eq. 3:6 must be divided by A_t, total basin area.

The coefficients α incorporate the effects of magnitude and frequency of wind. For instance, α_i might have the form

$$\alpha_i = \frac{u_i t_i}{\Sigma(u\ t)}, \qquad (3:7)$$

where u_i = mean velocity of wind direction i,

t_i = duration of wind from direction i,

$\Sigma(u\ t)$ = product of wind velocity and time over period of analysis.

It would be easy to alter the α_i to allow for the ineffectiveness of snow transfer below a threshold wind velocity.

Other effects are incorporated in η, which represents overall transfer efficiency. The "other effects" include gradient, ground surface roughness, air temperature and possible changes in the snow surface such as crust formation and increase in snow grain size.

Source and sink areas for the Mecham River basin were plotted on base maps traced from the orthophotomap of Resolute. Detail on the orthophotomap was used to locate snow traps. Relevant information on the basin is given in Table 3:2, and Figure 3:2 shows the base map which was used to compile source and sink area maps. A working copy of one of these appears in Figure 3:3.

TABLE 3:2

Areal Data for Snow Flux Calculations

P (km)	C	A_t (km^2)	A_p (km^2)	A_p/A_t
50	0.49	97.7	185.5	1.899

Zones of potential transfer (km^2)

i) into the basin : 109.2

ii) out of the basin : 76.3

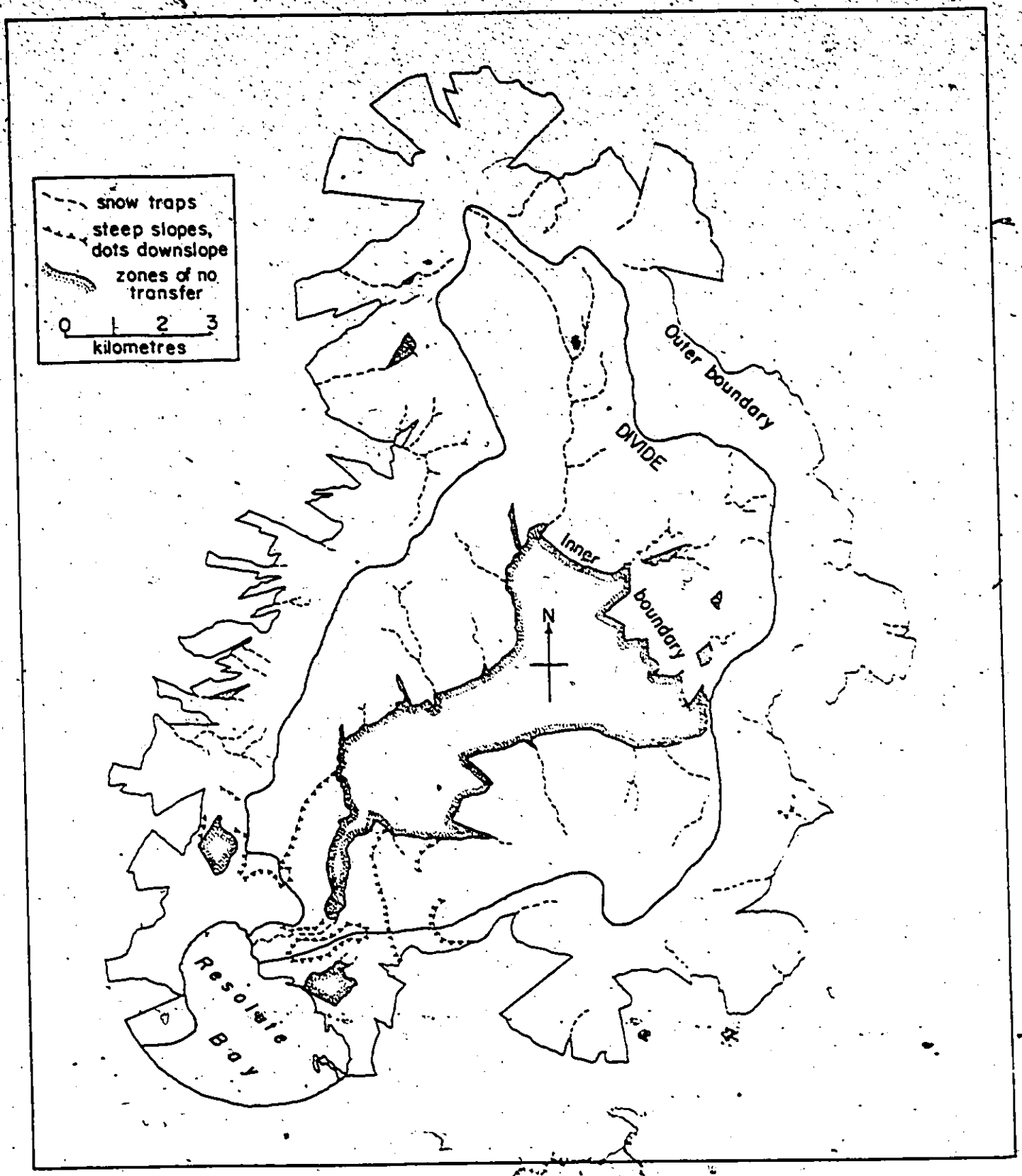


Figure 3:2. Base map for blowing snow measurements.

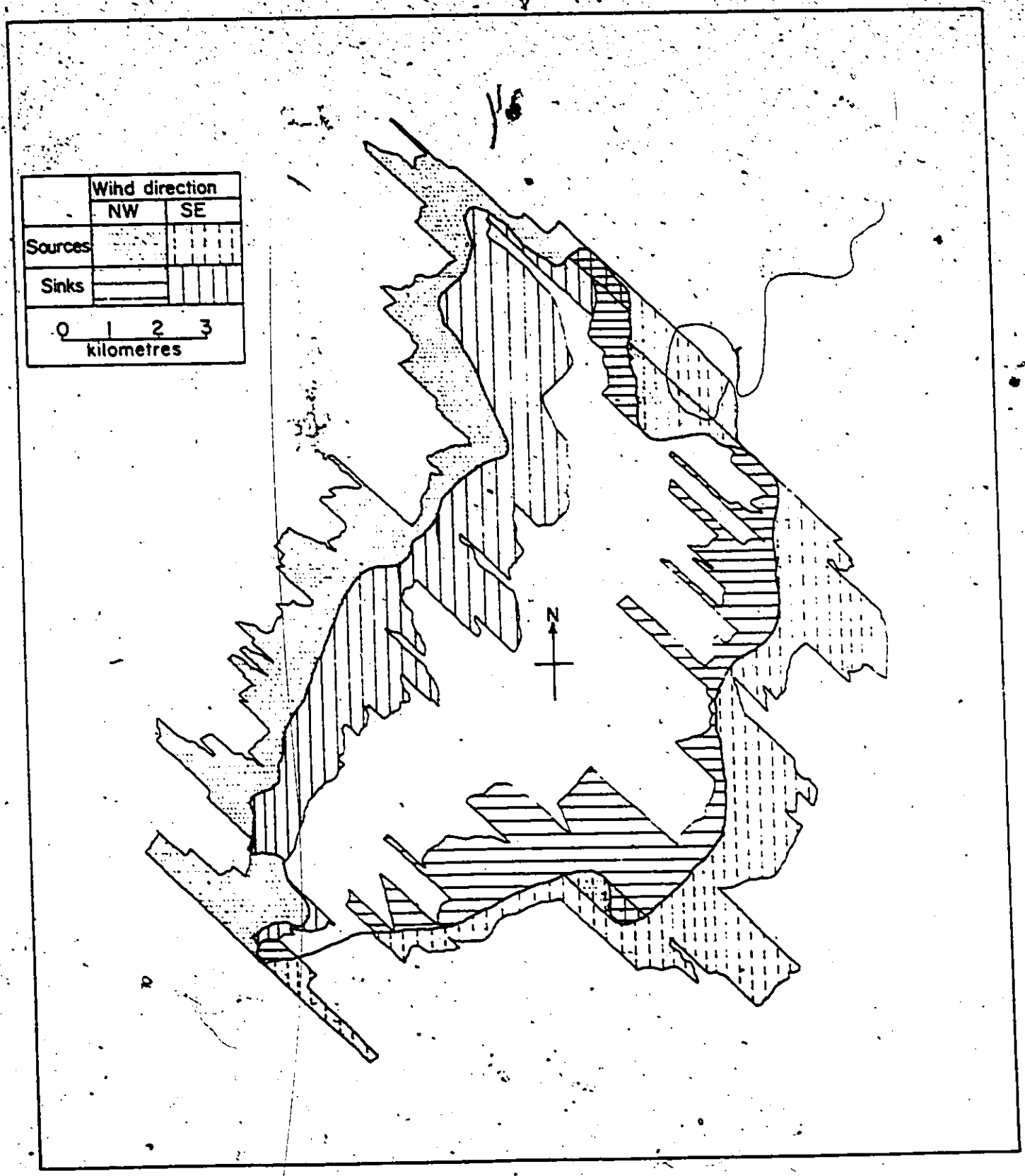


Figure 3:3. Snow transfer zones for NW and SE winds.

The snow flux model explained above may be criticized on a number of grounds. First, it is not supported by any accurate numerical information, and all of the terms on the right hand side of eq. 3:6 must be admitted to be crude approximations to what actually happens or is supposed to happen. The only term which is based on immediately relevant measurements is S, and its presence in both eqs. 3:1 and 3:6 means that at least one of the terms in eq. 3:1 is a function of one of the other terms; actual fluxes of wind-borne snow are probably proportional to snowfall, but not in the simple way of eq. 3:6. The measurement of A_{in} and A_{out} on maps is based on a very generalized assumption that transfer zones are 3 km wide, and the calculation of α_i according to eq. 3:7 implies firstly that any non-zero wind velocity will transfer snow, and secondly that the relationship between wind velocity and snow transfer is linear. In other words, eq. 3:7 gives too much weight to light winds which probably move no snow at all, and too little weight to high winds which move most of the snow.

Furthermore, it is not at all clear that eq. 3:6 is even of the right structure for the problem in hand. It neglects, for example, a potentially significant non-linearity in the process of snow

transfer across watersheds, by causing contributions from different wind directions to be added clockwise around the compass instead of in strict chronological order through the winter: as snow is transferred during one discrete event in which wind direction remains constant, so the "true" value approximated by S - the amount of snow lying on the surface of the terrain - changes somewhat for the next discrete event. This sequence of changes occurs concurrently with the seasonal increase in S due to accumulation, but both phenomena would be awkward to build into a simple model.

However, all of these grounds for criticism share one property, which is that neglecting them tends either to produce an overestimate of B or (at least on the basis of what is now known) to lead only to small changes in its final value. From the general geometric considerations given above, and the specific data presented for the Mecham River basin in Table 3:2, a small value of B may be anticipated. Table 3:3, therefore, may be consulted to show two things: first, that the net flux of blowing snow through the Mecham River basin is likely to be positive,

and second that, although sources or sinks for given wind directions reach about 7 per cent of total basin area, the flux is likely to be very small in an average year. The wind velocities and direction frequencies in Table 3:3 are means of published averages for the winter months September-May in the years 1955 to 1962. The results of the calculations in Table 3:3 is the equation

$$\overline{B} = 0.01121\,\eta\,\overline{S}\,, \qquad (3:8)$$

where the overbar indicates that the calculations are done with averages. Since $\overline{S}$ = 63.5 mm at Resolute for the nine winter months,

$$\overline{B} = 0.71\,\eta \text{ mm.} \qquad (3:9)$$

Even if Komarov's high value for η (0.8), obtained on the steppes of Siberia, is used in eq. 3:9, $\overline{B}$ is equal only to 0.57 mm, which is less than 1 per cent of S and a negligible quantity for water balance calculations. Even in exceptionally windy winters the flux of windborne snow will remain proportionally small, for high winds are usually accompanied by increased precipitation.

TABLE 3:3

Data on Snow Transfer Zones

	Wind Direction								
	N	NE	E	SE	S	SW	W	NW	TOTAL
A_{in}, km^2	26.0	23.0	29.0	26.4	17.0	21.6	24.0	24.4	191
A_{out}, km^2	23.5	28.7	23.1	28.0	23.9	18.7	20.0	20.6	186
$(A_{in}-A_{out})$, km^2	2.5	-5.7	5.9	-1.6	-6.9	2.9	4.0	3.8	4.9
u, m s^{-1}	5.34	7.02	6.91	6.50	5.20	4.11	3.41	4.90	5.42*
t, per cent	17	10	15	13	7	3	7	24	96
α	0.16	0.13	0.19	0.16	0.07	0.02	0.04	0.22	0.99
$\alpha(A_{in}-A_{out})/A_t$	0.004	-0.007	0.011	-0.003	-0.005	0.001	0.002	0.008	0.011

* mean wind velocity;

more significant numbers are underlined;

$\overline{B} = 0.01121 \quad \eta \; \overline{S}$.

Although the analysis leading to this result may resemble a numerical sledgehammer taken to crack a common sense nut, a cautionary note is provided by the statement of Cook (1967) that "the [Mecham] river's potential runoff from melting snow is appreciably higher than the average snowfall records would suggest". The context of this suggestion was an early attempt to show that runoff from Arctic drainage basins should not be neglected, but the suggestion was erroneous at least as far as it referred to extra contributions from blowing snow. Further, Black (1954), in an often quoted paper, says that "an unknown but appreciable amount of snow... falls in the [Barrow, Alaska] area but is lost by wind drift out to sea or inland". Here the suggestion is of a negative flux, and it seems that "snow is very rarely replenished by southerly winds". One suspects, however, that careful measurement would give the same result at Barrow as it would at Resolute: blowing snow may be neglected in the water balance, and eq. 3:1 may again be rewritten as

$$S + R - E - Q = 0 \qquad (3:1b)$$

3.3.2 Evaporation One might expect the smallest term in eq. 3:1b to be evaporation because it is relatively small when temperature and net radiation are low. Whether or not this is so will transpire in what follows.

The flux of water vapour at the ground surface of an Arctic drainage basin is the sum of three processes: evaporation, sublimation and condensation. These components are distinguished by the release or consumption of differing amounts of latent heat: for evaporation, $(2501 - 2.4T)$ kJ kg^{-1}, where T is the temperature of the reaction in degrees centigrade; for sublimation, $(2832 - 0.3T)$ kJ kg^{-1}; for condensation, these quantities are released rather than consumed. At the surface of the earth, the energy for evaporation or sublimation must normally be supplied from net radiation, and the surface energy balance can be written

$$R_n - G = H + \lambda E, \qquad (3:10)$$

where R_n = net radiation (W m^{-2}),

G = conductive heat transfer downward into the ground (W m^{-2}),

H = convective heat transfer upward to the atmosphere (W m^{-2}),

λ = latent heat of vapourization/sublimation ($J\ kg^{-1}$).

E = evaporation/sublimation ($kg\ m^{-2}\ s^{-1}$).

It can be shown (Monteith, 1964) that, of the energy available for evaporation or convection, the proportion used for evaporation is a function of the air temperature, the vapour concentration gradient from surface to air, and the aerodynamic resistance properties of the surface:

$$\lambda E = \frac{\Delta}{\Delta + \gamma} (R_n - G) + \frac{\rho\ C_p}{r_a} (D_z - D_0) \qquad (3:11)$$

where Δ = $\partial e_s / \partial T$ is the slope of the relationship between saturation vapour pressure (e_s) and temperature (T), and is calculated at the wet-bulb temperature by the method of Dilley (1968);

γ = $\Theta C_p / \varepsilon\lambda$, the psychrometric "constant" ($N\ m^{-2}\ K^{-1}$);

Θ = atmospheric pressure ($N\ m^{-2}$);

C_p = specific heat of dry air at constant pressure (assumed constant at 1005 $J\ kg^{-1}\ K^{-1}$);

ε = ratio of molecular weights of water vapour and air = 0.622;

ρ = density of air = $1.292 - (0.0047 T_d)$ ($kg\ m^{-3}$);

r_a = aerodynamic resistance to vapour transfer ($s\ m^{-1}$);

$D_{z,0} = (T_d - T_w)_{z,0}$, the wet-bulb depression at heights z and 0 (K);

$T_{d,w}$ = dry-bulb and wet-bulb temperatures (K).

Eq. 3:11 expresses what is known as the combination model of evaporation, first formulated by Penman (1948). For convenience it may also be written

$$\lambda E = \lambda E_{rad} + \lambda E_{pot}$$

where "rad" stands for "radiative", and "pot" stands for "potential" to suggest that the second term arises from the presence of a vapour concentration gradient. Research on the applicability of the combination model and its derivatives has shown that, despite difficulties in the evaluation of certain terms in the equation, the model is perhaps the most promising for eventual use on a routine basis for calculating evaporation (cf. Rouse and Stewart, 1972; Davies and Allen, 1973).

The meteorological data required for the model are wet and dry bulb temperatures at the surface and in the air near the surface, horizontal wind speed, soil heat flux (G), net radiation and (for calculation of γ) atmospheric pressure. Surface temperatures are very difficult to measure, and most variants of the combination model depend upon assumptions about how the surface

wet-bulb depression D_0 varies with soil moisture or atmospheric saturation. Of the remaining terms all save the soil heat flux are monitored routinely at many (but not most) weather stations. At Resolute, pressure, net radiation, wind speed and atmospheric wet-bulb depression are all measured hourly, and a rough estimate of G is possible from twice-daily measurements of soil temperature at 0.05, 0.10, 0.20, 0.50, 1.00, and 1.50 m below the surface.

3.3.2.1 Evaluation of Terms in the Combination Model

3.3.2.1.1 Soil Heat Flux and Surface Temperature The flux of heat into the ground (or out of it during winter) can be written as

$$G = k_g \frac{\partial T}{\partial z} , \qquad (3:12)$$

where T and z are temperature and depth and k_g is the thermal conductivity of the soil. There are no data available on the thermal properties of the soils of the Resolute area, but crude estimates for the Mecham basin as a whole are acceptable in view of the variability of its surface and the relatively small differences in conductivity between different kinds of soil. A value

of 1.5 $W\ m^{-1}\ K^{-1}$ is adopted for the soil cover of the Mecham basin; this is roughly the conductivity of a coarse sand or fine gravel of moderate porosity, with void spaces partly occupied by ice.

The energy represented by G is transferred, during most of the year, at the snow surface, while G is evaluated, with this method, in the ground beneath the snow. The problem can be corrected in part by supposing the conductive and other fluxes within the snowpack to be equal to the soil heat flux, or in other words by ignoring the details of the thermal regime of the snowpack and stipulating that in steady state conditions all the heat entering the pack leaves it again a "short" time afterwards. This is probably a rough approach to the true state of affairs, and can be accepted for estimates of G, but it gives unsatisfactory values for T_s, the surface temperature. Surface temperature, however, is only needed explicitly in the correction of r_a for non-neutral conditions, to be discussed later. G itself is given by the approximation

$$G = k_g \frac{T_0 - T_{0.05}}{0.05}, \qquad (3:13)$$

where the subscripts to T represent depth in the soil and T_0 is found by a least-squares fit of the form

$$T_z = a + b \exp(z) \qquad (3:14)$$

to the ground temperature data; the temperature at depth $z = 0.00$ is

$$T_0 = a + b. \tag{3:15}$$

Eq. 3:14 was chosen in preference to the more usual

$$T_z = T_0 \exp(bz) \tag{3:16}$$

because it fits better the temperature curve in the upper part of the profile. Since temperature varies not only with depth but also with time, eq. 3:16, although theoretically more sound, performs less well in practice.

The procedure adopted for estimating surface temperature T_s (as opposed to temperature at the ground surface T_0) was to assume heat flux equal in the ground and the snow, so that

$$\Delta T_s = G\frac{d_s}{k_s} \tag{3:17}$$

and

$$T_s = T_0 + \Delta T_s , \tag{3:18}$$

where d_s is depth of snow and k_s is the thermal conductivity of snow. k_s increases non-linearly with snow density, the values quoted by Kondrat'eva (1945) being about 1/30 to 1/2 of that found in soil.

In detailed meteorological studies it is customary to measure G directly with soil heat flux plates. These devices have shortcomings of their own, and in very detailed work it is necessary to measure or calculate all three components of the thermal conductivity,

$$k_g = \rho C \kappa , \qquad (3:19)$$

where ρ = density (kg m^{-3})

C = heat capacity (J kg^{-1} K^{-1})

κ = thermal diffusivity (m^2 s^{-1}) .

To determine heat capacity it is necessary to know the volume fraction of the soil occupied by its major components such as ice, water, organic matter and mineral constituents. Determination of thermal diffusivity requires analysis of soil temperature changes in two dimensions, time and space, and the assumption of a constant k_g masks the potentially important phase lags associated with the vertical propagation of heat in the ground. For example, the lowest depth used in the least-squares routine of eq. 3:14 is 1.5 m, at which temperature changes originating at the surface arrive much-damped with a lag of about 1/2 - 2 months; the three upper depths, on the other hand, are affected by changes with periods of less than one day. It is

not known what effect this has on heat flux determinations, or how important the effect is, but in a study at Norman Wells in the Mackenzie Valley, Beattie et al. (1973) found that fluxes determined by the gradient and flux methods did not agree. These authors, however, were not sure which of the two methods was performing the more accurately.

3.3.2.1.2 Resistance The resistance term for vapour transfer is variously interpreted as the time taken for a unit volume of air to exchange vapour with a unit area of surface, in which case it has dimensions $s\ m^2\ m^{-3}$, or by identifying it with the corresponding term for momentum transfer. In the latter case it is concentration of horizontal momentum

$$M = \rho u \tag{3:20}$$

divided by the shearing stress τ, which, assuming a wind speed profile in which wind speed increases with the logarithm of height above z_0 (Sellers, 1965), is given by

$$\tau = \rho\left[\frac{k\ u}{\ln(z/z_0)}\right]^2 \quad ; \tag{3:21}$$

dividing eq. 3:20 by eq. 3:21 yields

$$r_a = \frac{\left[\ln(z/z_0)\right]^2}{k^2 u}, \qquad (3:22)$$

where k = von Karman's constant = 0.41,

u = horizontal wind speed ($m\ s^{-1}$)

z = height of wind speed measurement (m),

z_0 = a roughness length characteristic of the surface (m).

A much-used rule of thumb for roughness length is that it is one tenth the height of the roughness elements (Priestly, 1959, p. 20; Eagleson, 1970, p. 230), and in this study the dominant roughness elements are taken to be grains of moderately old snow with diameters about 1 mm, giving z_0 = 0.1 mm. Beneath the snow, the roughest surfaces in the Mecham basin are on felsenmeer and talus slopes made of very coarse gravels, the smoothest on silty tills and estuarine sediments, so when snow depth is zero z_0 is set to 10 mm.

It is unfortunate that z_0 has to be guessed at, but the effect on r_a of changes in z_0 is small. Doubling or halving the roughness length changes the resistance by about 15-25 per cent in the roughness range 0.1-10 mm, while dividing or multiplying by ten produces changes of 40-75 per cent.

3.3.2.1.3 Wind Speed At Resolute wind speed is measured at a height of 12.2 m. Before any calculations are done with wind speed, it is adjusted to screen height (1.25 m) according to the relationship

$$u_{1.25} = u_{12.2} \frac{\ln(1.25/z_0)}{\ln(12.2/z_0)} \qquad (3:23)$$

giving

$$u_{1.25} = 0.68\ u_{12.2} \qquad (3:24)$$

when $z_0 = 0.0001$ m, and

$$u_{1.25} = 0.81\ u_{12.2} \qquad (3:25)$$

when $z_0 = 0.01$ m.

The logarithmic velocity profile is a very general feature of boundary layer mechanics, and the assumption of a logarithmic profile is the only assumption possible in the context of this study. When wind speed is measured at only one height some assumption about the profile is essential, for otherwise it would not be possible to derive and calculate the aerodynamic resistance. However, the logarithmic profile is a description of an ideal to which, because of topographical and other complications, reality only approximates roughly. Two studies which illustrate, at different scales, the

danger of assuming the logarithmic profile are Fraser (1959) and Langleben (1974).

3.3.2.1.4 Atmospheric Stability. The identification of resistance to vapour diffusion with resistance to momentum transfer rests on the assumption that the vapour diffusivity K_v is equal to the viscosity K_m. (In the equation

$$-\lambda E = -\rho\lambda K_v (\partial q/\partial z) \quad , \qquad (3:26)$$

where q is specific humidity, K_v is the analogue of K_m in eq. 3:21). It is now clear from studies such as that of Dyer and Hicks (1970) that this is true only when the atmosphere is in neutral equilibrium (i.e., when the temperature gradient is equal to the adiabatic lapse rate and turbulence is purely mechanical with no contribution from buoyancy); Webb's results (1970) suggest that K_v is also equal to K_m in stable conditions. In fact, a more critical assumption about the state of atmospheric turbulence, namely that neutral equilibrium prevails, has already been made for the purpose of adjusting the measured wind speed to screen height. Not only K_v but also K_m must be corrected in non-neutral conditions; the stability correction would appear in eq. 3:23 as

$$r_a = \frac{[\ln(z/z_0)]^2}{k^2 u} \phi_{v1} \qquad (3:27)$$

and an analogous term ϕ_m would appear in eq. 3:21, which is part of the derivation of both eqs. 3:23 and 3:24. The stability correction can be calculated empirically as a function of some other variable which measures stability, such as the Richardson number, and one acceptable form (see Dyer and Hicks 1970 and Businger, 1966) would be

$$\phi = 1/(1 - mRi)^n \qquad (3:28)$$

with m = 16, and m = 1/2 for vapour and 1/4 for momentum. The Richardson number is given by

$$Ri = \frac{g}{\theta} \frac{(\partial T/\partial z) + \Gamma}{(\partial u/\partial z)^2}, \qquad (3:29)$$

where g is acceleration due to gravity, θ is potential temperature in degrees Kelvin and Γ is the adiabetic lapse rate. The temperature gradient could be approximated by

$$\frac{\partial T}{\partial z} = \frac{T_d - T_s}{1.25}, \qquad (3:30)$$

where T_d is dry-bulb temperature, T_s is taken from eq. 3:18 and 1.25 m is screen height.

The attempt to correct for stability stalls, however, upon the need to know the wind speed gradient in order to calculate the correction: one purpose of the stability correction is to correct the wind speed gradient, and the procedure becomes circular. It is not possible, with the data now routinely available at Resolute, to correct the combination model for non-neutrality of the atmosphere. If the correction is to be applied, data on wind speed must be gathered at at least one new height, and an improvement in the estimation of surface temperature would also be desirable. Even with these additions to the available data it is not certain that the accuracy of the data would be enough to make the correction worthwhile.

There is, however, some evidence in the literature to support the view that neutrality may be assumed without serious consequences for the combination model. The stability correction is, after all, usually a small correction to a relatively conservative variable in a term which frequently vanishes from the model. This is a consolation in necessity, but work such as that of Bradley (1972) makes it seem not unreasonable.

3.3.2.1.5 Wet-bulb Depressions The surface wet-bulb depression is the last and probably the most important term in this adaptation of the combination model. Since it can not be measured, attempts have in the past been made to cope with the vapour concentration gradient in other ways, for example by making measurements in "special" situations. One such is where $D_0 = D_z = 0$: a saturated environment in which λE_{pot} vanishes from eq. 3:11. Another, of more interest here, is where $D_0 = 0 \neq D_z$, which leads to the "potential evaporation" that occurs from a surface at which moisture supply is not limiting. It is probably safe to assume that moisture supply is not limiting at a snow surface for, at least when it is at melting point, it behaves as a water surface, with most or all of the vapour transfer taking place between reserves of melt water and the atmosphere. In winter, when temperatures are very low, it is reasonable to suppose that vapour is transferred mainly between the outer atmosphere and the atmosphere of the pores in the snowpack, which is probably at or near saturation most of the time.

Although winter is much the longer of the two climatic seasons in the High Arctic, the vapour fluxes are likely to be much smaller than those in summer because both e_s and $\partial e_s/\partial T$ are very small at low temperatures. Summer evaporation, however, is more difficult to predict than winter evaporation, because among other things basin surfaces are partly snow-covered and partly snow-free. For a daily analysis this problem could be handled with a simple estimate of average fractional snow cover, but at the monthly scale of this analysis no great error is introduced by supposing the entire basin snow cover to disappear when the snow cover disappears at the weather station.

The question, what to do about D_0 during the snow-free period, is more difficult to answer and more critical in its influence on the results given by the model. Flood plain surfaces and slopes below semi-permanent snowpatches probably evaporate at the potential rate throughout the summer, and the entire basin is wetted by occasional rainfalls, which occur most heavily and most frequently in early August. The assumption that $D_0 = 0$ is therefore least accurate in mid-July, that is, after most of the snow cover has

gone but before moistening of the surface by rain. Mid-July is a period of moderate, though diminishing, supply of permafrost meltwater, but this supply is evidently insufficient to maintain evaporation, for the soil surface is sometimes observed to dry out at this period in the extensive well-drained parts of the basin.

Ideally, D_0 might be approximated as a function of measured soil moisture, but soil moisture is never measured routinely at weather stations and in any case a good average of soil moisture would be very expensive to obtain for an area close to 100 km^2. The device adopted here to account for soil moisture is a crude drying curve which works in the following way. After the disappearance of the snow cover, D_0 is set to zero for a period of I days; after I days have elapsed, D_0 is allowed to increase linearly for a period of J days until it is equal to D_z, so that on any day $D_0 = \frac{j}{J}D_z$, where j is the number of days from the start of drying; after the J days have elapsed, D_0 is set to D_z until more than 2.5 mm of precipitation falls on any one day or until the snow cover is renewed; if either of these things happens, D_0 is reset to zero and the accounting procedure begins again. The values chosen for I and J are 5

and 10 days respectively (actually 10 and 20 half-days since there are two calculations per day).

The form of the drying curve obtained in this way is intended to mimic observations which indicate that evaporation proceeds at or very close to the potential rate while soil moisture content is very close to soil moisture capacity, and that it drops away rapidly from the potential rate once the moisture content goes below some critical fraction of moisture capacity. Such observations have been made by Davies and Allen (1973), and by a number of workers including Van Bavel (1967) whose results are collated by Priestly and Taylor (1972). In extreme desiccation evaporation should drop to zero, but this situation is improbable in the Mecham basin, and λE_{rad} should always, if accurately evaluated, be a good estimator of the minimum evaporation from the basin.

The numbers chosen for I and J are arbitrary, representing only estimates of the average time that it would take for the basin average of D_0 to reach the measured D_z. Account has been taken in these estimates of the variability of average moisture supply in

different parts of the basin, and of the proportional distribution of well-drained and ill-drained areas, but no claim of physical "reality" can be made for I and J. The best that can be said of the surface wet-bulb depressions derived with their aid is that they should be reasonably close to the "true" basin averages.

The evaluation of D_0 in summer is thus the weakest point of the version of the combination model which is developed here. Two checks on the range within which actual evaporation varies are, however, available. Firstly, actual evaporation is unlikely to be much less than the value given by the first term of the combination model. Secondly, it is most unlikely that actual evaporation would exceed the published values of Class 'A' pan evaporation, which is measured at Resolute in July and August. Two values are published: firstly, the net water loss from the pan, which is greater than actual evaporation because evaporation pans are known to give results which are too high, due to their exposure above the ground surface and to conductive heat exchange through the pan walls; secondly, the "calculated lake evaporation",

which is obtained by empirical adjustment and which should be roughly equal to potential evaporation if the empirical adjustment is correct. The values obtained with the combination model are checked against these published values in a later section.

3.3.2.2 Advective Distortions Up to this section it has been assumed tacitly that horizontal imports and exports of heat are small enough that the energy balance and thus the evaporation can be calculated without accounting for them. If advection were important, however, it might affect the situation so greatly as to make the results worthless.

The characteristic or scale length associated with this study is about 100 km. Advective effects on scales ~1000 km are beyond its scope, but at smaller scales some thought should be given to them.

Firstly, local non-uniformities with scales ~1 km or less might be expected to influence the calculations, for this is roughly the separation between the wind mast, the radiometers and the other instruments at Resolute, and it is also roughly the distance from the instruments to the watershed of the Mecham basin. However, the terrain in the area is uniform at this

scale, and the only problem which might arise is the existence of the Resolute townsite and airstrip. The site is reasonably well exposed and the small amounts of heat and water vapour which it generates should normally diffuse rapidly.

At scales of about 100 km, the principal control on advection is the contrast between land and sea surfaces. There are long fetches over land all around the perimeter of the Mecham basin, the shortest land fetches decreasing to less than 5 km to the south and (in the southern part of the basin) to 5-10 km to the west. Offshore to the west there are sea fetches of several hundred km, and sea fetches of 50-100 km to the south-west and south. Surface contrasts should affect results largely through contrasts in moisture supply and albedo. Table 3:4 shows that there are three different surface regimes of importance, and that contrasts occur from July to September. The greatest contrasts are in late August and September, but the frequency of westerly and southwesterly winds at this season is only about 12 per cent (Meteorological Branch 1970). The Mecham basin can therefore be considered well sheltered from large advective fluxes of heat.

TABLE 3:4

Surface Contrasts at a Scale of 100 km

Season	Land*	Sea*
October - mid-June	Off-white to white, saturated	Off-white to white, saturated
mid-June - August	Dark grey, saturated to moderately dry	Off-white, saturated
August - September	Near-white to white, saturated	Black, saturated

*Albedo described on a colour scale from perfect black to perfect white

3.3.2.3 The Energy Balance at Resolute Evaporation, considered as a flux of water, is discussed in its proper place in the synthesis at the end of this chapter. Considered as an energy flux, however, it is also of some interest. Since net radiation is already measured, and the combination model demands an evaluation of the soil heat flux and results in an estimate of the latent heat flux, it is possible to calculate the convective heat flux as a residual and thus to produce a complete if imprecise statement of the energy balance. This is done here for the hydrological year 1969-1970, which is representative of

the years for which calculations have been made. Table 3:5 lists the four terms of the energy balance, together with auxiliary information, for each month and for the entire year; Figure 3:4 is a graph of the energy balance. Perhaps the most interesting feature of Table 3:5 is that the annual Bowen ratio ($\beta = H/\lambda E$) is 0.70; although monthly Bowen ratios are very high in winter, the corresponding fluxes are directed downward and are extremely small, so that the annual ratio is very close to the monthly ratio for July. Similarly, the annual mean for $\Delta/(\Delta + \gamma)$ suggests that about one-fifth of the available net radiation is used for evaporation, whereas a "true" annual value $\lambda E_{rad}/(R_n - G) = 0.48$ is close to the July mean.

On an annual basis, the radiative flux is small in comparison with low-latitude radiation, but positive nevertheless. The evaporative flux is larger than the convective flux, but whether or not it is "negligible" depends entirely upon the context of discussion; in a global or continental context its importance depends upon one's assessment of it in comparison with fluxes found at lower latitudes; in the Arctic context its significance is clearly not to be belied. The annual soil heat flux is negative, as is the case in each of the years examined. Although it would be unwise to infer much from this,

TABLE 3:5

The Energy Balance, 1969-1970

Year	Month	R_n	H	G	λE	λE_{rad}	λE_{pot}	$\frac{\Delta}{\Delta+\gamma}$	D_z-D_0	Bowen ratio
69	09	10	-13	14	8.9	-0.7	9.6	0.36	0.13	-
	10	-28	-20	-4.5	-4.3	-5.3	1.0	0.26	0.05	4.6
	11	-26	-13	-12	-1.1	-1.1	0.1	0.09	0.01	12
	12	-24	- 6.7	-17	-0.4	-0.4	0.0	0.10	0.00	18
70	01	-28	- 5.8	-21	-0.2	-0.2	0.0	0.07	0.00	29
	02	-38	-15	-22	-0.6	-1.0	0.4	0.07	0.01	25
	03	-22	0.4	-24	1.0	0.3	0.8	0.08	0.08	0.40
	04	1.1	15	-16	2.1	1.9	0.2	0.10	0.05	7.1
	05	38	35	-13	16	14	2.0	0.25	0.27	2.2
	06	83	36	14	33	30	3.8	0.41	0.34	1.1
	07	157	58	23	76	64	13	0.46	0.44	0.76
	08	103	46	18	39	38	1.1	0.44	0.03	1.2
Whole year		19	10	- 4.6	14	12	2.7	0.22	0.12	0.70

Fluxes are in W m^{-2}, (D_z-D_0) in deg C; symbols as in the text

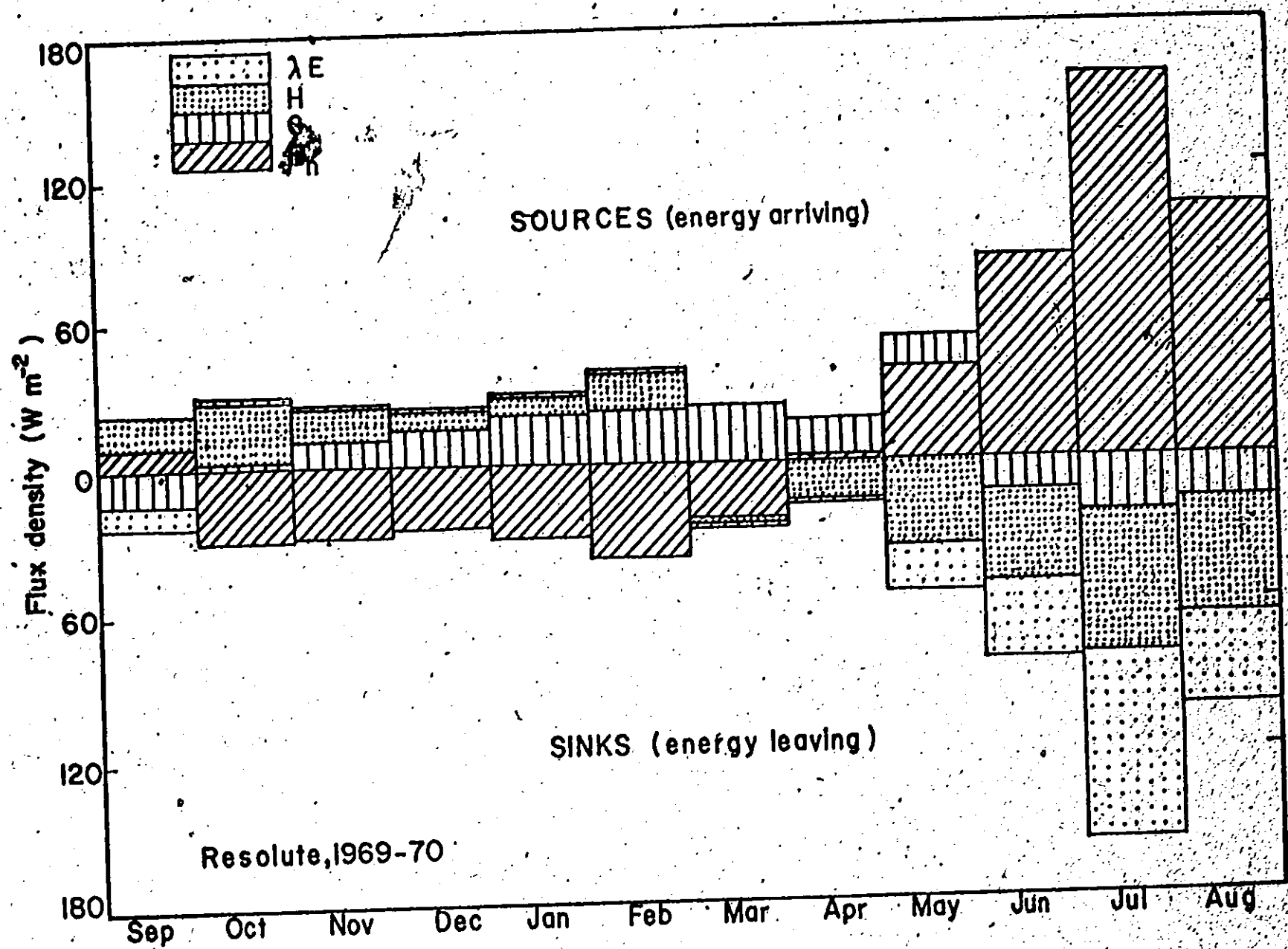


Figure 3:4. The energy balance, 1969-70.

the sign of the soil heat flux is at least consistent with the slight cooling of the Northern Hemisphere in recent decades, which has been inferred from other kinds of evidence.

On a monthly scale, inspection of the columns of Table 3:5 shows that, from the onset of winter until about December the terrestrial radiation of the polar night is fuelled mainly by heat withdrawn from the atmosphere. From December onward the major source of heat is the ground, although the atmosphere continues to give up heat until March, which is the end of the polar night. Cooling of the ground continues into May, by which time the evaporative flux is a substantial fraction of total heat transfer to the atmosphere. By July-August the evaporative flux is larger than the convective flux, for a different energy regime prevails over snow-free surfaces. All three heat fluxes reach annual maxima in July. However, the conductive thermal diffusivity of the soil is much smaller than the turbulent thermal diffusivity of the atmosphere, and this means that the atmosphere cools much more rapidly than the ground at "nightfall" in September-October, just as the ground is slower to warm up in April and May.

3.3.2.4 A Monthly Version of the Combination Model

Monthly evaporation as presented below is a sum of values obtained twice daily. These diurnal values are probably of questionable worth for interpreting diurnal or even day-to-day evaporation changes, but they are the best available sources for the monthly figures. They are, however, costly to use.

A practical problem arose during calculation of the water balance: the necessary data were not available on magnetic tape beyond the end of calendar 1972. Since values were wanted for the hydrological year 1972, the possibility was investigated of using published monthly means as inputs for the combination model. In the "Monthly Record" and "Monthly Radiation Summary" (Atmospheric Environment Service, 1972 ff.) all of the required information can be found, except that wet-bulb temperature must be reconstructed tediously from dry-bulb temperature and vapour pressure. Twice-daily means are printed for net radiation, ground temperatures, air temperatures and pressure, the last two being reported for synoptic hours which are two hours away from the hours of ground temperature measurement. A single daily mean snow depth and a single monthly mean wind speed are published also.

These data were used to calculate monthly evaporation for January 1972 to August 1973, giving eleven months with twice-daily calculations as controls (data were missing for July 1972). In the monthly version of the model, snow-covered and snow-free parts of each month were treated separately. For snow-free periods the rainfall records were searched and each day was given a weight of the same kind as that explained in sec. 3.3.2.1.5. The surface wet-bulb depression for the month was taken as the monthly weight times the monthly mean of the depression at screen height.

The performance of the monthly version was measured against that of the twice-daily version. The closeness of the fit is demonstrated in Figure 3:5 and in the regression statistics

$$E_d = a + b\,E_m, \qquad (3:31)$$

where $a = 1.63 \pm 3.23$,

$b = 0.998 \pm 0.062$,

$S_e(\bar{E}_d) = 3.23$,

$S_e(b) = 0.062$,

$r = 0.983$, $n = 11$.

Here S_e is standard error, r is correlation coefficient, n is sample size and E_d, E_m are evaporation from the

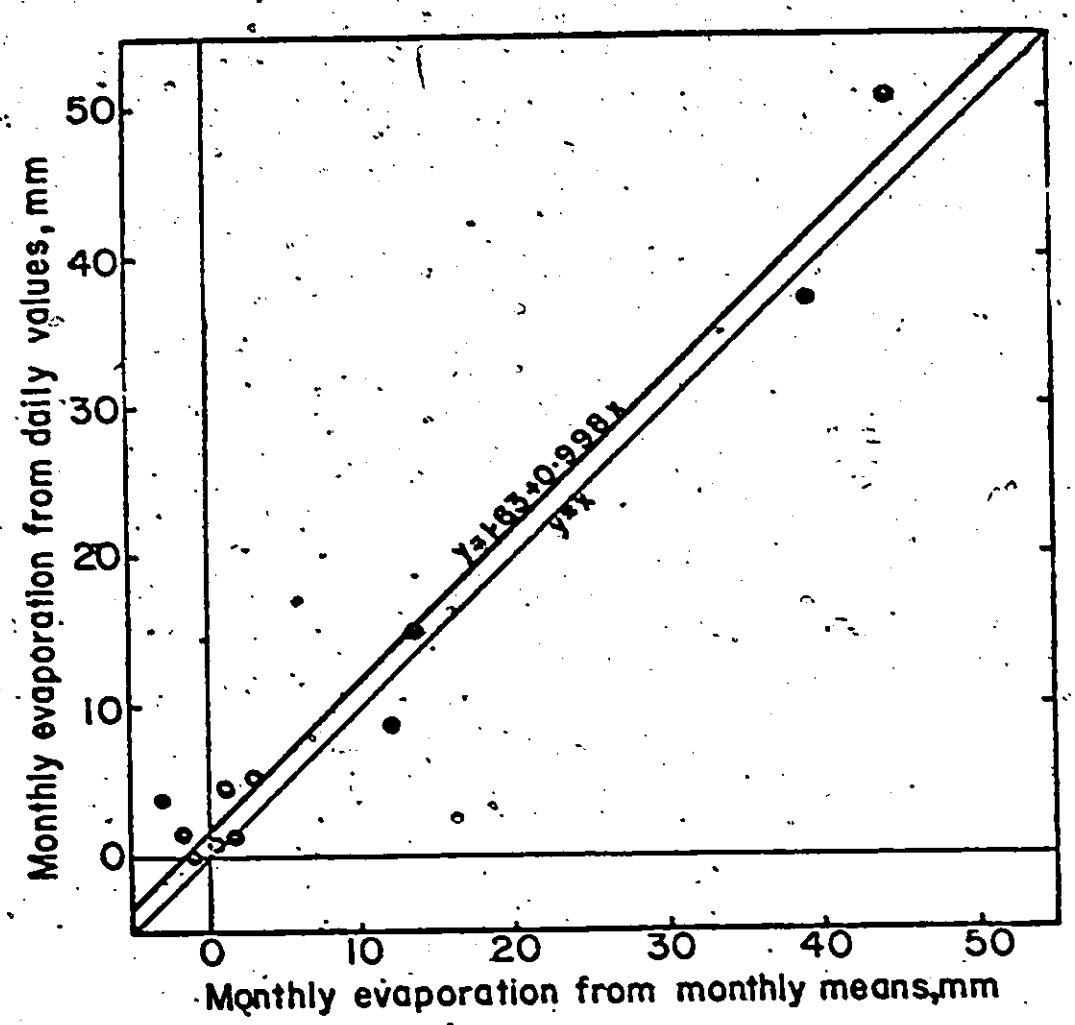


Figure 3:5. Evaporation from twice-daily and monthly versions of the combination model.

twice-daily and monthly versions respectively in millimetres. The standard errors suggest that, with a very high probability, eq. 3:31 is equivalent to

$$E_d = E_m \qquad (3:31a)$$

The greatest difference between the two versions for 1972 is only 6.6 mm.

This result has two implications. First, the water balance for 1972-73 can safely be considered comparable with those for earlier years. Second, the authorities charged with measuring climate could, with some re-arrangement and at trivial computational expense, publish the raw data and results for monthly and annual evaporation at Resolute in less than a page of print. Few stations in Canada have the range of routinely collected data which is available at Resolute, but the procedure developed here should be widely applicable, certainly in sparsely vegetated areas, and with some confidence in many regions above the treeline. Below the treeline complications appear such as the variation of z_0 with windspeed for plant surfaces which yield to the wind, and some of the assumptions permissible at Resolute would be less appropriate. A number of

technical improvements which are desirable include the elimination of guess work in the estimation of z_0, k_g and other factors, the addition of wind speed measurements at screen height and temperature measurements as close to the surface (of the snow or the ground) as possible, and the development of a routine for measuring soil moisture content in summer.

3.3.3 Snowfall and Rainfall For the water balance calculations, the six-hourly totals of snowfall and rainfall recorded at Resolute have been summed to give monthly totals. Problems encountered with the measurement of precipitation were discussed in Section 1. One of these problems, which is peculiar to areas such as the High Arctic, is the underestimation which results from too frequent emptying of precipitation gauges and consequent over-reporting of "trace" precipitation (Jackson, 1960); the problem is aggravated in the High Arctic by the very high frequency and long duration of very light precipitation. A reasonably effective way of allowing for this problem is to count the number of traces reported and to regard each of them as a quantity of precipitation. In this analysis, then, each six-hourly trace recorded is assumed to represent 0.064 mm (0.0025 inches) of precipitation, and a new term

$$T_r = \delta N_t \qquad (3:32)$$

where N_t is the number of traces and $\delta = 0.064$ mm, is added to eq. 3:1:

$$S + R + T_r - E - Q = 0 \qquad (3:1c)$$

3.3.4 Discharge - The hourly discharges measured at the mouth of the Mecham River were added to give monthly totals, and because of the close attention devoted to it the discharge term is considered to be the most accurately known of the eight in eq. 3:1. The values used to give the discharge totals are regarded as accurate to ± 25 per cent at worst and in most cases to ± 10 per cent. The discharge for 1969-70, which was an exploratory year, is probably less accurate than the discharges quoted for the following years.

3.4 Results of Water Balance Calculations

Each term of eq. 3:1c was computed for each month of the hydrological years (beginning on 1 September) 1969-70 to 1972-73. For 1972-73 the calculations for E were performed on monthly data published in printed form.

The results of the water balance calculations are displayed in Tables 3:6-9, and Figure 3:6 represents the water balance for 1970-71. The principle feature illustrated is the gross disparity between inputs and outputs. For the years with evaporation data, there is no case in which measured inputs exceed 40 per cent of outputs, and even when allowance is made for trace precipitation the gross disparity remains. Clearly, the estimation procedures for one or more of the terms in the water balance are grossly wrong. The trace term T_r does not bear on the essential problem, but in any case the problem gets worse if it is discarded; changing the value of δ has the same effect if δ is reduced and if, improbably, δ is twice the value chosen (i.e., if δ has its maximum possible value) the resulting improvement is small. Much faith has already been placed in the data used to estimate discharge, and the attention devoted to the combination model should indicate that some confidence is about to be placed in the evaporation figures, at least to the extent of claiming that (say) $E \approx Q \approx P$. Precipitation, then, seems to be the culprit for the failure of eq. 3:1.

TABLE 3:6

The Water Balance, 1969-70

Year	Month	S	R	T_r	E	Q	Balance
		(millimetres)					
69	09	27.2	5.6	3.0	9.2	5.6	21.0
	10	24.1	0.0	2.6	-4.0	0.0	30.7
	11	3.8	0.0	2.2	-1.0	0.0	7.0
	12	4.6	0.0	3.4	-0.4	0.0	8.4
70	01	2.8	0.0	4.4	-0.2	0.0	7.4
	02	2.0	0.0	2.9	-0.5	0.0	5.4
	03	2.5	0.0	2.4	1.0	0.0	3.9
	04	4.6	0.0	2.8	2.0	0.0	5.4
	05	0.8	0.0	1.3	15.2	0.0	-13.1
	06	12.7	4.3	3.5	31.4	3.5	-14.4
	07	1.3	13.7	1.7	82.0	132.3	-197.6
	08	7.2	24.4	2.2	42.2	39.3	-47.7
Total		94	48	32	177	181	-184

TABLE 3:7

The Water Balance, 1970-71

Year	Month	S	R	T_r	E	Q	Balance
		(millimetres)					
70	09	13.7	0.0	3.6	14.5	0.0	2.8
	10	4.4	0.0	4.7	-2.1	0.0	11.2
	11	2.5	0.0	3.4	0.8	0.0	5.1
	12	1.8	0.0	4.0	0.4	0.0	5.4
71	01	1.8	0.0	1.5	1.2	0.0	5.1
	02	4.6	0.0	4.5	1.0	0.0	8.1
	03	0.2	0.0	2.8	2.2	0.0	0.8
	04	11.4	0.0	2.9	7.3	0.0	7.0
	05	15.5	0.0	4.0	16.2	0.0	3.3
	06	4.6	0.5	1.5	80.7	134.5	-205.6
	07	1.8	12.4	1.5	119.3	123.4	-227.0
	08	16.5	40.9	3.0	62.9	38.3	-40.8
Total		79	54	37	304	297	-431

TABLE 3.8

The Water Balance, 1971-72

Year	Month	S	R	T_r	E	Q	Balance
		(millimetres)					
71	09	14.9	0.0	3.3	5.5	0.0	12.7
	10	9.6	0.0	3.6	1.7	0.0	11.5
	11	1.8	0.0	3.9	2.5	0.0	3.2
	12	3.6	0.0	5.3	1.0	0.0	7.9
72	01	3.8	0.0	4.4	1.0	0.0	7.2
	02	3.3	0.0	4.4	1.1	0.0	6.6
	03	3.3	0.0	4.4	4.6	0.0	2.9
	04	0.2	0.0	4.2	5.3	0.0	-0.9
	05	6.4	0.0	4.7	15.0	0.0	-3.9
	06	9.4	0.0	4.1	37.1	0.0	-23.6
	07	0.0	6.1	2.0	89.1	131.0	-212.0
	08	9.1	18.8	2.6	50.9	7.4	-27.8
Total		62	25	47	215	138	-216

TABLE 3:9

The Water Balance, 1972-73

Year	Month	S	R	T_r	E	Q	Balance
			(millimetres				
72	09	7.9	0.0	3.8	12.1	0.0	-0.4
	10	3.0	0.0	5.7	-2.6	0.0	6.1
	11	3.3	0.0	5.9	-0.9	0.0	8.3
	12	1.5	0.0	7.2	0.0	0.0	8.7
73	01	4.6	0.0	6.5	0.1	0.0	11.0
	02	1.0	0.0	5.0	0.2	0.0	5.8
	03	0.8	0.0	6.6	0.6	0.0	6.8
	04	1.0	0.0	3.5	4.4	0.0	0.1
	05	2.0	0.0	3.4	22.5	0.0	-17.1
	06	0.0	5.8	2.0	96.3	64.7	-88.5
	07	5.6	15.5	1.6	91.0	18.7	-68.3
	08	7.9	17.0	2.4	43.9	21.8	-38.4
Total		39	38	54	268	105	-164

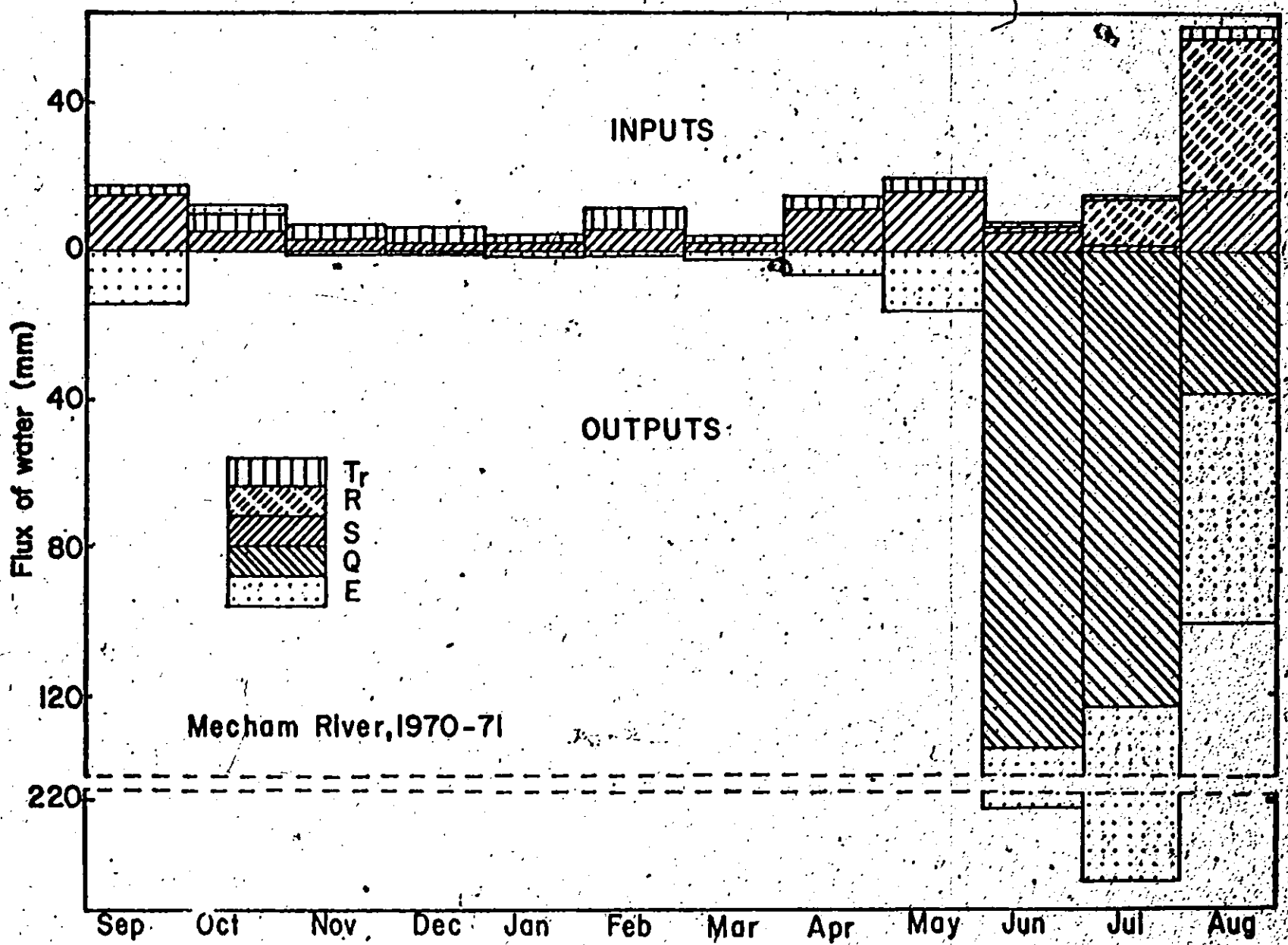


Figure 3:6. The water balance, 1970-71.

Before this conclusion is pursued, however, a careful re-examination of the evidence in favour of the discharge and evaporation figures is called for.

3.4.1 Re-examination of Discharge Figures A close inspection of the estimation procedure for discharge brings no radical new flaws to light, beyond those mentioned in section 1 as being common to almost all measurements of discharge and water level. Indeed, the amount of time spent in 1971 on calibrating the stage-discharge rating curve (Fig. 1:16) was probably longer than average, and it is hard to picture a more stable channel cross-section that the one chosen in that year and illustrated in Fig. 1:15. One cause for doubt which has not yet been mentioned is that occasional interpolations have been made silently in some of the August and September rows, to allow for unmeasured runoff from rainstorms arriving after the removal of the stage recorder at the end of the field season. All of the measured rainfall was routed directly to the same month's discharge column. The amounts of water involved in this armchair control of the water balance were 11.4 mm in August 1971,

24.4 mm in August 1970 and 5.6 mm in September 1969. The procedure is the most rational way of accounting for the destination of the rainfall, but again the larger problem remains if the solution to the lesser problem is discarded. For the water year 1969-70, reducing discharge by 30 mm (= 5.6 mm + 24.4 mm) and ignoring trace precipitation makes it possible to balance the water budget if it can be supposed that there was net condensation of 8 mm.

3.4.2 Re-examination of Evaporation Figures The possibility remains that it is unjustified to claim that E ≈ P in the High Arctic; that negative values should actually be envisaged for E. There is plenty of scope for scepticism about the evaporation figures produced above, for they result from stretching a sensitive model to and perhaps beyond its limit: as far as is known, the combination model in this form has not been applied before to routinely-gathered information from a weather station.

It is particularly unfortunate that there are not yet any high quality control data against which to compare these results, but checks on the magnitude of the calculated vapour fluxes can be made with reference to the results produced by other techniques of calculation.

3.4.2.1 Comparison with Pan Evaporation The Atmospheric Environment Service attempts to measure evaporation with Class 'A' pans in July and August at Resolute, and its attempts are successful about half of the time. Even in summer Resolute has an extreme environment, in which the pan water is apt to freeze. Table 3:10 shows the results achieved with an evaporation pan in the period of this study. For comparison with the published values E_{rad} and E are also tabulated, and a separate computer job was run to generate the evaporation which would have occurred had the surface been saturated throughout each month: to do this, the "drying curve" algorithm was overridden with a command to set D_0 equal to zero.

The comparison is only meaningful for two of the five months for which evaporation figures were published. The published "estimates" do not fit the pattern which appears in July 1970 and July 1971, and the most conservative explanation is that the estimates are merely summations of incomplete data. The relationship $E_{rad} < E < (E_{rad} + E_{sat}) < E_{Lake} < E_{pan}$ is to be expected from theory and from the arguments advanced earlier in this chapter, and this is the pattern which appears in July 1970. In July 1971 the pattern is

TABLE 3:10

Model Evaporation and Pan Evaporation, 1969-1973

Year	Month	E_{rad}	E	$(E_{rad}+E_{sat})$	E_{lake}	E_{pan}
70	07	68	82	112	119	130
70	08	41	42	78	-	66*
71	07	76	119	175	152	176
71	08	30	63	70	62*	68*
73	07	56	91	101	86*	89*

Data in mm; E_{lake} is calculated lake evaporation, E_{pan} is net water loss from pan; E_{rad} is the vapour flux corresponding to λE_{rad}; E_{sat} is the vapour flux corresponding to λE_{sat} (= λE_{pot} with $D_0 = 0$).

similar except that $(E_{rad} + E_{sat}) \approx E_{pan}$, suggesting that E_{sat} is too large. The complicated empirical correction used to calculate E_{lake} may give results slightly too high for the lake evaporation which it is intended to reproduce: in July 1970

$$E_{lake} / E_{pan} = 0.92 \qquad (3:33a)$$

$$(E_{rad} + E_{sat}) / E_{pan} = 0.86 \qquad (3:33b)$$

In any case, actual evaporation E from a partly dry land surface should be substantially less than both E_{lake} and E_{pan}, and the combination model gives, for July 1970,

$$E / E_{pan} = 0.63 \qquad (3:34a)$$

and for July 1971

$$E / E_{pan} = 0.68 \qquad (3:34b)$$

As far as model performance is concerned, the least optimistic conclusion which can be drawn from eq. 3:34 is that the model's results are not underestimates of actual evaporation.

It is much harder to fix a lower limit for the accuracy of the model. With some reservations, however, E_{rad} is taken to be that lower limit for present purposes. Much more detailed field work in the future

might establish this point. Strict depiction of reality would demand that the drying out of the soil proceed until $E = 0$, not until $E = E_{rad}$. Only the justification of intuition can be advanced in defence of the latter position, which when restated is that the basin average of active layer moistness never decreases to a state in which the available energy is unable to withdraw water from the soil.

Comparison of model evaporation with pan evaporation, then, suggests that E represents a most probable value of actual evaporation, and that the range $E_{rad} < E < (E_{rad} + E_{sat})$ gives upper and lower limits.

3.4.2.2 Comparison with a Large Scale Approach

Priestley and Taylor (1972) have collated the best data presently available in the literature, and have argued that "for apportioning the available energy over substantial saturated land areas" the following relation holds:

$$\lambda E = \Lambda \Psi (R_n - G) \ , \qquad (3:35)$$

where $\Psi = \Delta/(\Delta + \gamma)$,

$\Lambda = 1.26$ (Priestley and Taylor's α).

Eq. 3:35 states that evaporation is proportional to (but not equal to) the available energy times a function

of temperature. Λ was determined empirically to be equal to 1.26, as an average of pooled averages of all suitable measurements from nine studies with the common property that they were done very carefully. However, the individual observations seem to range from $\Lambda = 1.00$ to $\Lambda = 1.50$: unpooled averages and standard deviations range from 1.08 ± 0.01 to 1.34 ± 0.05 and 1.33 ± 0.21.

The "Priestley-Taylor equation", eq. 3:35, is a very promising instrument for separating the available energy over large areas, but its physical basis can not yet be considered established. It may mean no more than that, to a good approximation, $\lambda E_{sat} = \lambda E_{rad} / 4$. Studies are needed to provide a physical interpretation of Λ if there is one. For example, Davies and Allen (1973) succeeded in showing that, through its dependence on the Bowen ratio, Λ is a function of temperature, but that it should be virtually constant between 15°C and 30°C; they also showed that Λ is strongly dependent on surface wetness.

Priestley and Taylor define Λ as

$$\Lambda = \frac{\lambda E}{\Psi(\lambda E + H)} , \qquad (3:36)$$

where $(\lambda E + H) = (R_n - G)$. The results from Resolute were inspected for occasions when Λ could be estimated by

$$\Lambda = \lambda E / \lambda E_{rad} , \qquad (3:37)$$

which is equivalent to eq. 3:36. Monthly values were rejected if they did not satisfy the criteria that

i) they should come from the light season April - September, when evaporative and convective fluxes are both reasonably large;

ii) they should have an evaporative flux directed upward (no condensation);

iii) they should have $\Lambda < \psi^{-1}$, implying no advection of warm air: i.e., the convective flux should be directed upward.

21 months were found to satisfy each of these criteria, and on average for these months

$$\Lambda = 1.47 \pm 0.37 \qquad (3:38)$$

Strictly, eq. 3:35 is equivalent not to eq. 3:36 but to

$$\Lambda = (\lambda E_{rad} + \lambda E_{sat}) / \lambda E_{rad} , \qquad (3:39)$$

where $\lambda E_{sat} = \rho C_p (D_z / r_a)$.

λE_{sat} is a hypothetical quantity, but it has been calculated for purposes related to Table 3:10 for a total

of six months, two of which fail to satisfy criterion iii. For these six months Λ averages 1.46 from eq. 3:37 and 1.96 from eq. 3:39. The overall average of eq. 3:37 is well above Priestley and Taylor's determination, but when the sums are done "properly" (eq. 3:39) the Resolute value is even further from 1.26.

For the four annual figures obtained at Resolute, Λ (from eq. 3:37) averages 1.57 and ranges from 1.23 to 1.73.

A number of different interpretations of the high value of Λ at Resolute are possible. The Priestley and Taylor data may be wrong; the Resolute data may be wrong; or there may be a plausible framwork into which both values of Λ can be fitted. One shrinks from the first interpretation, and an inclination to the second is, perhaps, natural. On the other hand, although Priestley and Taylor used excellent raw material, the raw material at Resolute is average to very good for the context in which eq. 3:35 is intended to apply. The third interpretation, then, deserves some attention.

Λ has already been shown to vary with temperature and soil moisture. It is logical to expect that it should also be a function of surface resistance to vapour transfer, for in a sense the coefficient $(\Lambda - 1)$

takes the place of λE_{pot} in eq. 3:11, and λE_{pot} is a function of the three variables just mentioned. Of the nine averages used by Priestley and Taylor, some were obtained over water, some over crops such as alfalfa and snap beans, and one over recently ploughed bare soil. The dependence of Λ on surface resistance is obscure on the evidence now at hand; curiously, Λ was equal to 1.08 over the bare soil, which most closely resembles the surface of the Mecham basin. However, the nature if not the magnitude of the dependence can be stated: Λ should be inversely proportional to r_a and therefore directly proportional to windspeed and surface roughness (as measured by z_0). The Mecham basin and the Resolute area are relatively windy and relatively smooth, but since $\Lambda \propto [1/\ln(1/z_0)]^2$ and $\Lambda \propto u$ (i.e., Λ changes slowly with roughness) a high rather than a low Λ is implied.

Another possibly important control on Λ may be the physiological resistance offered by plants to the transpiration of water. There is no significant resistance of this sort at the surface of the Mecham basin, but over vegetated surfaces (such, incidentally, as occur in other parts of the High Arctic) stomatal

and other physiological resistances may become very large. This contrast, again, implies that Λ should be relatively high for the Mecham basin.

As far as Davies and Allen (1973) were able to show, Λ varies very slowly with temperature, but it is still not known whether it increases at temperatures below about 10 - 15°C, where there are few measurements. This question will not be decided until good measurements are obtained relating the Bowen ratio to temperature in cool conditions with D_0 equal to zero. However, a plausible if inconclusive case can be made for the approximate correctness of combination model estimates of λE_{pot}, based on the observation that the average monthly value of Λ at Resolute is within the range of values whose mean, determined by Priestley and Taylor, is 1.26; the case is supported by the theoretically argued dependence of Λ on surface resistance, and further support may be drawn from a possible variation with temperature which remains uncertain.

<u>3.4.3 Examination of Precipitation Figures</u> Whatever attitude is taken towards the adequacy of the evaporation estimate, the conclusion is inescapable that not all is

well with the published measurements of precipitation. If any positive value is chosen for E, the discrepancy in the water balance increases from a point at which it is already, in most years, substantial. Q is always greater than (S + R) and is greater than (S + R + T_r) in 3 years out of 4.

If it is accepted that actual evaporation lies between E_{rad} and (E_{rad} + E_{sat}), with E as its most probable value, the measure of the precipitation problem can be drawn by evaluating P as a residual of the water balance. For convenience the annual water balances are summarized in Table 3:11. The following

TABLE 3:11

The Annual Water Balance, 1969-1972

Year	S	R	P	T_r	E	Q	Σ_{in}	Σ_{out}	Balance
1969-70	94	48	142	32	177	181	174	358	-184
1970-71	79	54	133	37	304	297	170	601	-431
1971-72	65	25	90	47	215	138	137	353	-216
1972-73	39	38	77	54	268	105	131	373	-242

Data in mm; Σ_{in}, Σ_{out} = total inputs, outputs

relationships between terms, stated very roughly, are of interest:

$$T_r \approx R; \quad (3:40)$$

$$P + T_r \approx Q, \text{ excepting 1970-71}; \quad (3:41)$$

$$E \geqslant Q > P \text{ always}; \quad (3:42)$$

$$T_r / (E + Q) = 0.11 \text{ on average}; \quad (3:43)$$

$$(E + Q) / P = 3.95 \text{ on average}. \quad (3:44)$$

In words, about as much trace precipitation goes completely unrecorded as goes into the record as rainfall. With an allowance for the unrecorded precipitation the water balance will balance with reasonable accuracy, but only if evaporation is ignored. However, discharge always exceeds measured precipitation and, moreover, evaporation usually exceeds discharge.

Evaluating P as a residual of the water balance, it turns out that about 10 per cent of actual precipitation falls at rates less than 0.125 mm (0.005") in six hours, and so would never be measured by present practice. More strikingly, present measurement practices have a gauge catch deficiency of about 75 per cent. Most work on the performance of precipitation gauges deals in terms of gauge catch deficiency, but the problem is more simply and more graphically stated in

the following way: mean annual precipitation at Resolute is not 5.36" but ~21". This is about the same as the measured precipitation at Winnipeg, San Francisco or London, England.

This bald statement can be qualified in a number of ways, but the most expedient is to calculate the range within which the true value of actual precipitation should lie. The error in discharge is about 10 per cent, and therefore the upper limit of the range is $(E_{rad} + E_{sat}) + 1.10\ Q$; on average for the years studied,

$$(E_{rad} + E_{sat} + 1.10\ Q)\ /\ P = 5.03 \qquad (3:45a)$$

The lower limit of the range is $E_{rad} + 0.90\ Q$, and on average

$$(E_{rad} + 0.90\ Q)\ /\ P = 2.94 \qquad (3:45b)$$

Generalizing eqs. 3:44 and 3:45, actual precipitation over the Mecham basin is at least three times greater than the precipitation measured at Resolute, most probably about four times greater, and possibly greater still.

<u>3.4.4 Measurement of Precipitation</u> The article by Black (1954), referred to above (sec. 3.3.1), is entitled "Precipitation at Barrow, Alaska, greater than recorded". Its main finding was that actual precipitation is "at least two

to four times" the recorded precipitation at Barrow. The problem of measuring precipitation with accuracy is by no means new, and Black's article is merely one snowflake in a blizzard of reports which either, like this one, define the problem, or describe attempts to solve it. One of the most recent projects concerned with improvements in the technology of precipitation measurement is reported by Larson and Peck (1974) and Larson (1971 a,b, 1972 a,b). In summary the problem is that any precipitation gauge which projects above a surface generates instability in the windfield; air is forced upward and outward by the gauge, and the vertical wind must exceed the settling velocity of some fraction of the precipitating particles of water; the faster the wind the fewer the particles which enter the orifice of the gauge, and the greater the gauge catch deficiency. Because snowflakes are lighter and sometimes larger than raindrops, it is much more difficult to measure snowfall than rainfall, and the difficulty increases with wind speed.

Precipitation measurements, then, are least accurate where wind speeds are high and snowfall is a

large portion of total precipitation. In most middle and low latitude localities the error of precipitation measurement is less serious than it is in the High Arctic for these reasons. In many hydrological studies the greatest uncertainty is in the estimation of evaporation or of quantities related to storage, such as interflow and baseflow, although Crawford and Linsley (1966), for example, have pointed out that the most critical factor in watershed simulation is the input from rain gauge networks. The High Arctic is hydrologically anomalous in that outputs from the systems of water circulation exceed the measured inputs to them: evaporation estimated from a model can be considered more accurate than precipitation measured with gauges.

The conventional approach to correcting gauge catch deficiencies, when they are not ignored completely, is to place shields around the gauge orifice with the design of deflecting updraughts of air. It is universally appreciated that these shields improve accuracy only slightly, but they are the least costly way of approaching the problem. It is equally widely acknowledged that the most accurate way to measure precipitation is to place a gauge in the centre of a well-sheltered open

area, such as a small clearing in a forest. With properly chosen geometrical configurations (World Meteorological Organization, 1969), it is possible in places like forest clearings to measure precipitation essentially without error due to atmospheric turbulence. In exposed areas the forest clearing can only be simulated at some expense with materials such as snow fencing. However, artifical clearings or other fencing arrangements such as wind baffles are the only solution to the problem of direct measurement of precipitation in places like Resolute.

This does not mean that there are no alternatives in the estimation of precipitation. Hamon (1971), for instance, has described a "dual gauge approach" which gives good estimates of actual precipitation through a regression on the catches of one shielded and one unshielded gauge. The evaluation of precipitation as a residual of eq. 3:1 has been shown above to give better results than conventional gauge measurements, and this approach may be useful where diagnosis rather than prediction is the aim. Finally, a very simple way to measure P for the annual (or winter) water balance is to conduct an extensive survey of snowpack water equivalent on a judiciously chosen date at the end of

winter - early June in the latitude of the Mecham River. If no large snowfalls follow the date of the survey a very good estimate of annual precipitation should result, since theory demands that errors in rainfall measurement during summer should be much less than errors in snowfall measurement. Daily snow depth measurements are already taken routinely at Resolute, but the crucial measurement of snow density is missing at present. Snow course measurements which lead to estimates of water equivalent are made twice monthly, but the course consists of only ten stakes; the observations are sometimes omitted, and are not always made between the last large snowfall and the disappearance of the snow.

In June 1972, however, a 20-stake snow course was surveyed within the Mecham basin and gave the results which were reported in sec. 2.2. To recapitulate, the estimate of the snowpatk water equivalent from the snow course was four times the estimate from snowfall measurements during the winter, providing an interesting if circumstantial indication of the merits of the approach advocated above.

3.5 Summary

An annual water balance equation derived specially for unglacierized catchments in the High Arctic includes several terms in addition to the more familiar elements of the classical water balance equation. Of these additional terms, however, some can be dismissed as negligible at least in the Mecham basin.

Changes in the amount of water stored in snowbanks, for example, must be very small, since there are few semi-permanent snowbanks and these few seem to be relatively stable. Changes in lake storage can also be discounted because there are very few lakes in the basin. Changes in active layer storage - the year-to-year differences in the moisture content of the layer which thaws in summer - are of possibly greater importance. Although no data on soil moisture can be brought to bear on the problem, reasoning from what is known about the available water at the time of freeze-back each fall suggests that ordinarily permafrost meltwater can be neglected in the annual water balance. None of the years of study seem to have been exceptions to this norm.

The flux of blowing snow across the watershed of the basin requires more careful attention in the light of hints in the literature about its significance. Arguing from the topographic disposition of traps for blowing snow, however, and exploiting the findings of experimental research in Siberia, a strong case can be made for neglecting the flux of windborne snow. Only in basins much smaller than that of the Mecham River should the flux be a potentially large component of the water balance.

For the Mecham basin, then, if not necessarily for other High Arctic catchments, it should be possible to solve the water balance in terms of its conventional elements, precipitation, evaporation and discharge. Precipitation is measured at Resolute, and discharge was measured for the purposes of this study, but evaporation must be calculated indirectly.

The method chosen for these calculations was to apply the combination model to meteorological data on net radiation, dry-bulb, wet-bulb and soil temperatures, and windspeed. Approximations and assumptions are needed at several places in the model, but these are considered acceptable given the coarse scale at

which the model is applied. A point of possible practical value is that monthly mean data give results which are not statistically distinguishable from the estimates using twice-daily data which are the basis of most further calculations. Improvements in the model would include the elimination of guesswork in the estimation of some parameters and the addition of wind speed measurements at screen height and surface or near-surface temperature measurements; the most desirable improvement, however, would be to get reliable estimates of surface wetness so that the vapour concentration gradient between air and surface might be reliably known.

If the data now available are less than ideal for successful exploitation of combination-model theory, they are still unusually comprehensive at Resolute. Comparison of the results with pan evaporation measurements implies that the model estimates are close to the actual evaporation occurring from the basin surface. Further encouragement comes from a comparison of the results with the large-scale approach of Priestley and Taylor (1972), which has excited much recent interest as a simplification of the rather

demanding combination model. The estimates of this study do not agree with the estimates of the Priestley-Taylor approach, but plausible reasons can be given for the disparity. If the reasons are accepted, the performance of the model developed here can be regarded with more confidence.

The model results suggest more evaporation than might be expected considering the low temperatures which prevail in the study area. The main conclusion of the water balance analysis, in fact, is that basin outputs of water - by evaporation and discharge - greatly exceed the measured inputs - by precipitation. A numerical allowance for unrecorded trace precipitation does not help to solve this problem, and it can only be assumed that present techniques of precipitation measurements give amounts which are much too low: on average, actual precipitation is 400 per cent greater than measured precipitation. There are ways of improving precipitation measurements, but for water balance purposes the easiest way is to conduct an extensive survey of snowpack water equivalent as close to the end of winter as possible.

If, as the work of this section suggests, there is considerably more water circulating through the

Mecham River basin than would appear from published climatic normals, it is natural for a geomorphologist to suspect that these extra fluxes of water may imply larger fluxes of sediment. In an environment which is classified climatologically as a desert, the realization that in fact there are appreciable quantities of water flowing over land surfaces prompts enquiry into the geomorphic role of this running water. In reality the geomorphic enquiry prompted suspicions about the water balance, and not vice versa. But the geomorphic enquiry is better seen in the context of the water balance, and therefore follows it in logical sequence in the next section.

SECTION 4

FLUVIAL TRANSPORTATION OF SEDIMENT

4.1 Introduction

Classically, geomorphology has been concerned with the estimation of rates of landscape denudation. This concern can be traced long ago in the history of the discipline (e.g., Guettard, 1768-83, and Targioni-Tozzetti, 1752, quoted in Chorley, Dunn and Beckinsale, 1964), but only in the recent past have attempts to estimate denudation been characterized by reasoning from measurement and accumulation of data. In this section the results of programmes of measurement of sediment concentrations in the Mecham River and two streams in Ellesmere Island are presented. The programmes were mounted in pursuit of evidence to vindicate the impression conveyed by pictures such as Figure 1:1 - that, while there are distinctively periglacial processes and landforms, periglacial *landscapes* are moulded as much by fluvial as by cryogenic working of the sediments of which they are made. The evidence can be condensed into a few numbers which state the rates of denudation observed during the period of study, but many of the details of denudation are interesting

in themselves, and these details are discussed in their turn. The observed rates have added significance when findings from Section 3 about the water balance are borne in mind: in general, measurements of sediment transport give qualified support to the argument that the High Arctic is different, not only hydrologically but also geomorphologically, from the picture of it which is conventionally presented. But the qualifications themselves are illuminating.

Studies of the sediment flux in streams draining this periglacial environment have an immediate application, particularly in areas such as the Canadian High Arctic which are entering on phases of rapid economic expansion. Engineering construction can have dramatic effects on streams, and the converse is true: witness the destruction of parts of the Mackenzie Highway between Artic Red River and Inuvik by the spring flood of 1972. The sets of data on which this section relies were not collected with a view to solving a problem of this sort, but rather with a view to defining reliably the sediment load carried by single Arctic streams over short periods of time. Such data may prove useful in the future, and for the present they permit a rough evaluation of the geomorphic role of running water in at least a part of the High Arctic.

Considerable interest attaches to the role of running water in dissolving the limestone and dolomite bedrock of the Mecham basin. Thermal climatic control of carbonate rock solution has been a subject of controversy for some time, and a later section of this chapter helps to shed light on the controversy.

4.2 Stream Load

A common way of discussing the load carried by streams is to express the total load as a sum of three fractions: bed load, suspended load and dissolved load. Of the three fractions carried by the Mecham River, data are available for only two, suspended load and dissolved load. It is almost impossible to measure bed load accurately even under ideal circumstances, and such circumstances did not obtain on the Mecham River. Crude "guess estimates" of bed load have been made, but for suspended and dissolved load successful programmes of measurement were conducted. Sampling was first undertaken in summer 1970, but the major part of the programme was carried out in 1971, when over a 60-day period 100 samples of suspended sediment and about 175 samples for solute analysis were taken. For the analysis of solutes, all samples were analyzed for Ca^{++}, Mg^{++},

HCO_3^-, pH and, in theory but not in practice, specific conductivity (the conductivity bridge being out of order for a part of the season); 60 samples, one per day, were analyzed for Na^+, and 9 for SO_4^{--} and Cl^-. A few samples were measured for Si, which proved to be present in very small quantities, and on the basis of work done previously on the waters of "Jason's Creek" (Cogley, 1972) a decision was made to ignore K^+, Al and Fe.

4.2.1 Bed Load The bed load carried by the Mecham River is carried almost entirely at times of high discharge. In 1971 a simple record was kept of periods and channel widths over which bed material was in motion through the channel cross-section opposite the base camp, conditions being noted each time the river was waded (on an average, several times each day); when wading was impossible because of high stream velocities, it was assumed that bed material was in motion. The record indicated that even during the snow-melt flood bed load was carried for not more than a few hours each day, and that only rarely did bed material move at all points on the cross-section. After the end of the flood, bed material was only entrained by the stream for a few hours during the rainstorm of 4 August.

From the hydraulic theory of bed material movement developed by Einstein (1950) and others, bed load transport may be expected to be proportional to a characteristic grain diameter of the bed material, to roughness properties of the channel and to the energy slope of the sediment-bearing water flow. Accurate solutions of the bed-load equations derived by Einstein are extremely difficult to obtain, but an earlier, semi-empirical generalization derived by Meyer-Peter, Favre and Einstein (1934) and containing no terms for roughness is more amenable to treatment. However, no particular merit other than simplicity is claimed for it in the present context. Bed load transport is defined as

$$q_s = \left[aD\left(\frac{q^{2/3}S_0}{D} - b\right)\right]^{3/2} , \qquad (4:1)$$

where q_s = bed material discharge (kg s^{-1} per metre of cross-section),

q = water discharge (kg s^{-1} per metre of cross-section),

D = a representative grain diameter (m),

S_0 = energy slope,

a,b = coefficients with values of 2.5 and 17 respectively.

For the ~~Mecham~~ River values of D = 0.02 m and S_0 = 0.0089 (obtained from maps of the basin) are appropriate.

Calculations using these values and measured values of q were made for the appropriate periods (a total of 200 hours) and channel widths (an average of 3 m) in 1971. The result suggests that a total of 1.06 t km^{-2} was discharged as bed load during that year, but this figure must of course be treated with caution. It is no more than a rough indication of actual bed material transport.

It can not be said whether a figure of 1.06 t km^{-2} y^{-1} reflects accurately the importance of bed load relative to suspended and dissolved load, more accurate values for which are discussed below. It is reasonable, however, to conjecture that bed material is comparatively unimportant in the total load of the Mecham River, for the channel of the river has a low energy slope. The bed material, also, consists largely of coarse material derived from raised beach deposits and mechanically weathered fragments of limestone and dolomite.

Another bed load equation which can be used as a check on eq. 4:1 is that of Kalinske (1947). Explanations of the theoretical derivation of this formula will

be found in Henderson (1966) and Carson (1971). The derivation is based on the assumption that the velocity of particles in traction is proportional to the shear velocity

$$u_* = (g\ R\ S_0)^{\frac{1}{2}} \quad , \qquad (4:2)$$

where g = acceleration due to gravity = 9.81 m s^{-2},

R = hydraulic radius = A/P,

A = cross-sectional area of flow (m^2),

P = length of wetted perimeter (m)

It is assumed also that bed material transport is governed by the ratio of inertial forces, exerting drag on grains in the bed, and gravitational forces which resist drag. A dimensionless number, similar to the Froude number, which expresses this ratio for grains is the Shields number

$$Sh = u_*^2/gD[(\rho_s - \rho_f)/\rho_f]$$

$$= RS_0/1.6D \qquad (4:3)$$

where ρ_s = density of grains = 2600 kg m^{-3} ,

ρ_f = density of water in which grains are immersed = 1000 kg m^{-3}.

The Kalinske bed load equation itself can be written

$$Q_b = 10W_b\rho_s u_* DSh^2 \quad , \qquad (4:4)$$

where Q_b = bed material transport (kg s^{-1}),

W_b = width over which bed material is in motion (m),

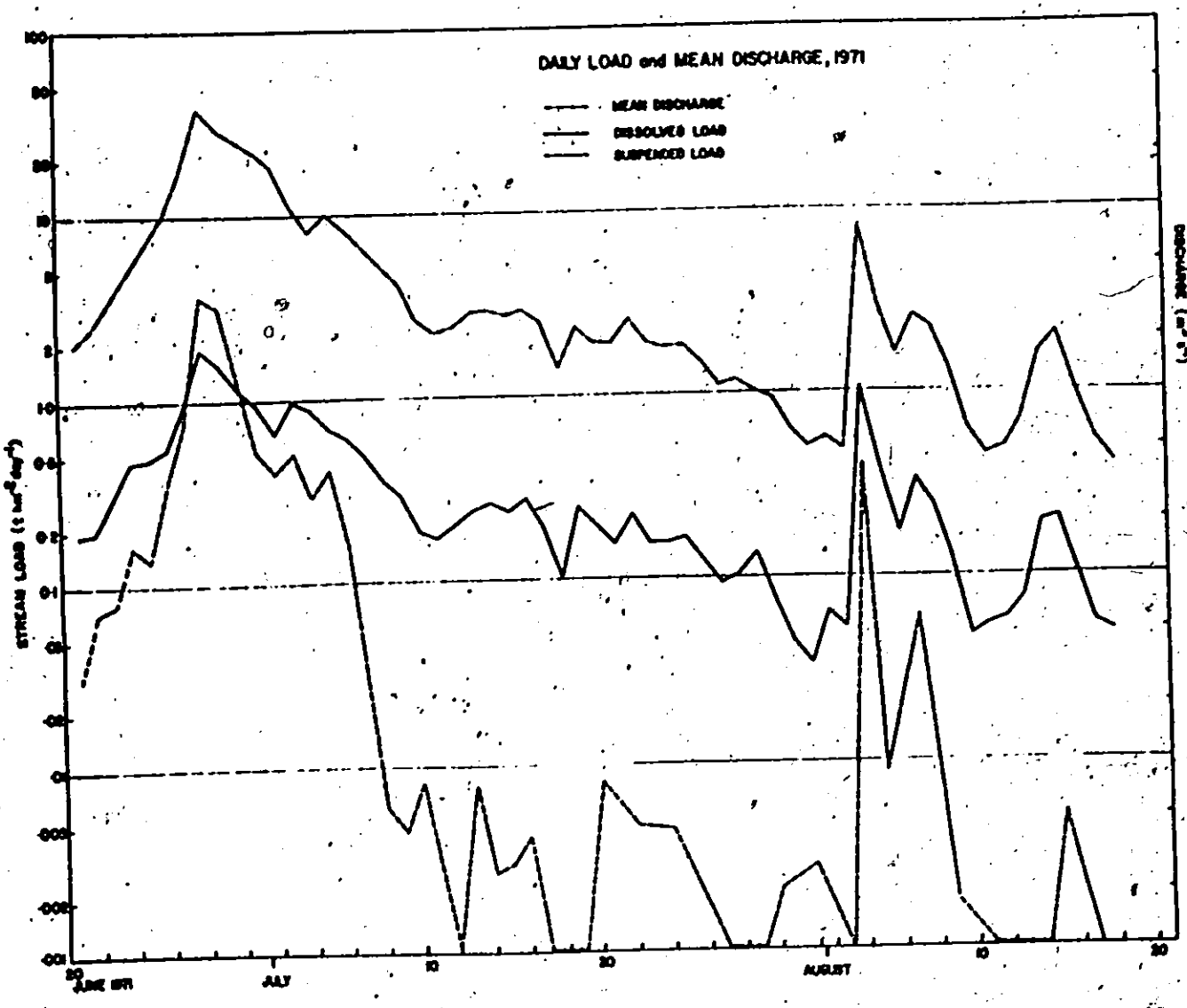
and the number 10 is dimensionless and represents the intercept of a fitted curve. Rewriting eq. 4:4 in terms of basic quantities,

$$Q_b = 31789 W_b R^{2.5} S_0^{2.5}/D, \qquad (4:5)$$

if; as for eq. 4:1, W_b = 3 m, S_0 = 0.0089 and D = 0.02 m, and if an average for R is taken from discharge measurements as 0.2 m, eq. 4:5 gives Q_b = 0.64 kg s^{-1}. Over a period of 200 hours this works out to a total seasonal transport of 459 t, or 4.70 t km^{-2}. Considering the extent of the assumptions made, this value is in good agreement with that from eq. 4:1, and the implication is reinforced that bed load is relatively small in the Mecham River. It remains, however, no more than an implication.

4.2.2 Suspended Load

4.2.2.1 Mecham River The suspended and dissolved load of the Mecham River for each day of observation in 1971 are graphed on a logarithmic scale in Figure 4:1, together with the logarithm of daily mean discharge. Values on a daily basis were calculated by extrapolating



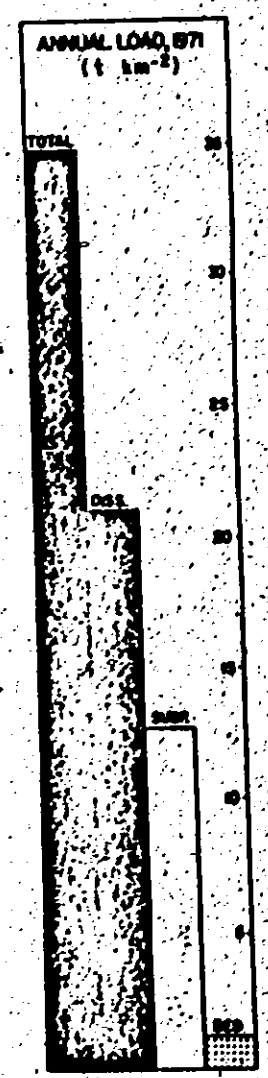


Figure 4:1. Discharge and load, Mecham River, 1971.

the available instantaneous values halfway to each of their neighbours in time, multiplying by an appropriate factor and summing or averaging the results for each day. Suspended sediment concentrations were measured on a weight per weight basis, and dissolved load was calculated as discharge times the sum of species concentrations which were measured individually in mol m^{-3} (1 mol m^{-3} = 1 mol l^{-1}).

Sampling frequency was varied so as to be greater at higher discharges. For some lesser dissolved species, concentrations had to be extrapolated over periods as long as one week, and the lower concentrations of suspended sediment were extrapolated over two days. These errors of extrapolation should not be excessively large.

Østrem, Ziegler and Ekman (1970) and Ziegler (ed., 1972) draw attention to some important problems in the extrapolation of instantaneous sediment measurements for estimation of total stream load. Their detailed records for glacial streams in Norway show occasional bursts of anomalously turbid water, which would be missed by a daily sampling programme. Bursts of this sort occur in the Mecham River during the

spring flood, when snow dams across the channel are forced aside by the accumulating water: one such happened on the afternoon of 22 June 1971, when suspended sediment concentration increased from 22 p.p.m. at 1525 h to 1300 p.p.m. at 1540 h, then decreased to 12 p.p.m. at 1555 h. Since these bursts are infrequent and short on the Mecham River, they do not contribute much to its total annual load.

The sums of the daily values of load are depicted in the bar graph on the right of Figure 4:1, together with the estimate of bed load described above. The inaccuracy in these annual sums due to measurement and extrapolation error is increased by an error which arises from upward adjustment by extrapolation to allow for load carried in the ~10 days of flow before measurements started and the 10-20 days of flow after measurement ended. These adjustments are, however, relatively small (very small in the case of the end-of-season adjustment). The measured total load for the year was 34.04 t km^{-2}, the adjusted total load 34.89 t km^{-2}. A rounded figure of 35 t km^{-2} y^{-1} is accepted as being a truer reflection of the accuracy of the measurement. The major result shown in the bar

graph of Fig. 4:1 is that, by a substantial margin, solution is the main mechanism for the removal of the products of erosion from the Mecham River basin. Only at the highest observed discharges does suspended load exceed dissolved load, and for most of the runoff season suspended load is practically nil. (Suspended sediment concentration varied from nil to 215 p.p.m., excluding the very short period mentioned above, when it reached 1300 p.p.m.; dissolved solids concentrations, on the other hand, ranged between 62 and 134 p.p.m.)

Suspended load was measured as 12.7 km^{-2} y^{-1}, and this is consistent with the observation that the Mecham River has low energy slope and flows through a terrain dominated by coarse surface sediments. One question which arises about this figure concerns its representativeness of other parts of the High Arctic. A number of facts can be adduced to support an assumption that the Mecham River is representative of a more extensive area than its own drainage basin. The Allen Bay and Read Bay Formations cover a considerable part of the Arctic archipelago, and strata which are lithologically similar are also widely distributed. Throughout the area underlain by these calcareous

sediments precipitation is low and vegetation cover is sparse, being restricted to a small number of low-growing plant species. There is also an approximate coincidence of limestone and/or dolomite bedrock with a regional physiography which features well-preserved plateau surfaces, slightly dissected and fringed by lowlands of limited extent. In the Arctic regions which resemble southeast Cornwallis Island there is scope for considerable variation in the factors which control the suspended load of streams. Stream gradient and the availability of different sizes of sediment are two of these factors. Although there are no data to support an assertion, it seems reasonable at least to suggest that the Mecham River may be representative of conditions in these areas.

4.2.2.2 "Sverdrup" and "Schei" Rivers It is interesting to enquire into the variation of sediment yield over still more extensive areas, for geomorphology has long been concerned with the rates at which different landscapes are eroded. For example, much higher suspended loads would be expected of streams draining terrain in which most of the unconsolidated surface sediment was fine-grained. This is so in south central Ellesmere Island, where streams have concentrations of suspended

sediment between 100 and 1000 p.p.m. even at moderate flows, and which reach several thousand p.p.m. at high flows. These streams drain an area which is mantled by much recent glacial and fluvioglacial detritus, and although the same strata of the Allen Bay and Read Bay Formations outcrop in their drainage basins as are found in that of the Mecham River, the most extensive beds are of weak Devonian and Tertiary sediments. These rocks weather readily to a regolith which contains very little gravel and only small amounts of sand.

Discharge and sediment data on two Ellesmere Island streams, the "Sverdrup River" and Schei River", were obtained in 1973 (McCann, Cogley, Woo and Blachut, 1974). Figure 4:2 shows the location of these two streams, Figures 4:3a and 4:3b the pertinent data collected for them. The "Sverdrup" basin (above the measurement section) is about 1630 km^2 in area, of which about 1250 km^2 is occupied by a section of the Ellesmere Ice Cap. The "Schei" is tributary to the "Sverdrup" and drains 91 km^2, 32 km^2 of which is ice-covered. (The confluence of "Schei" and "Sverdrup" is downstream of the measurement section on the "Sverdrup".)

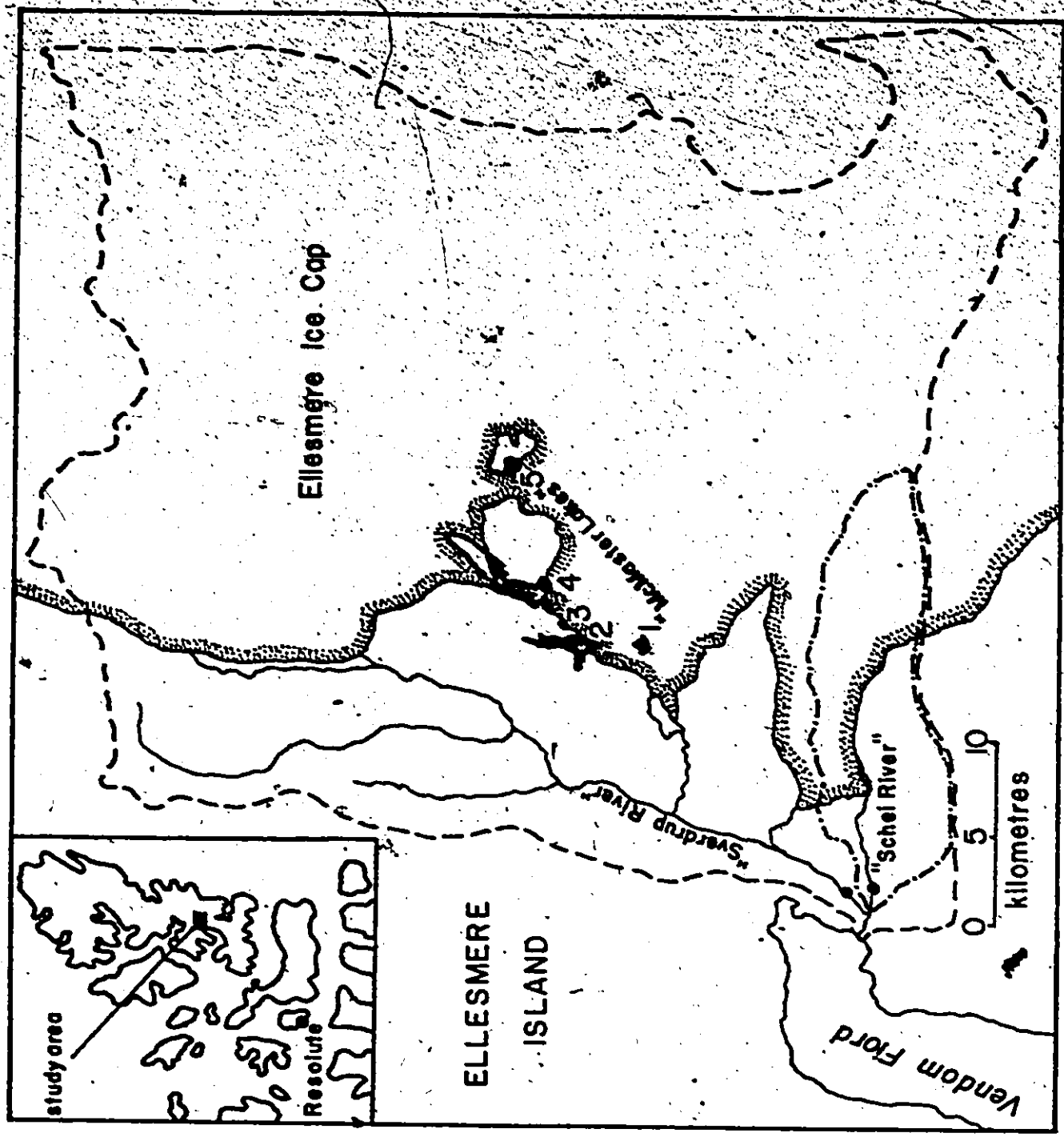


Figure 4:2. Location of "Sverdrup River" and "Schei River", Ellesmere Island.

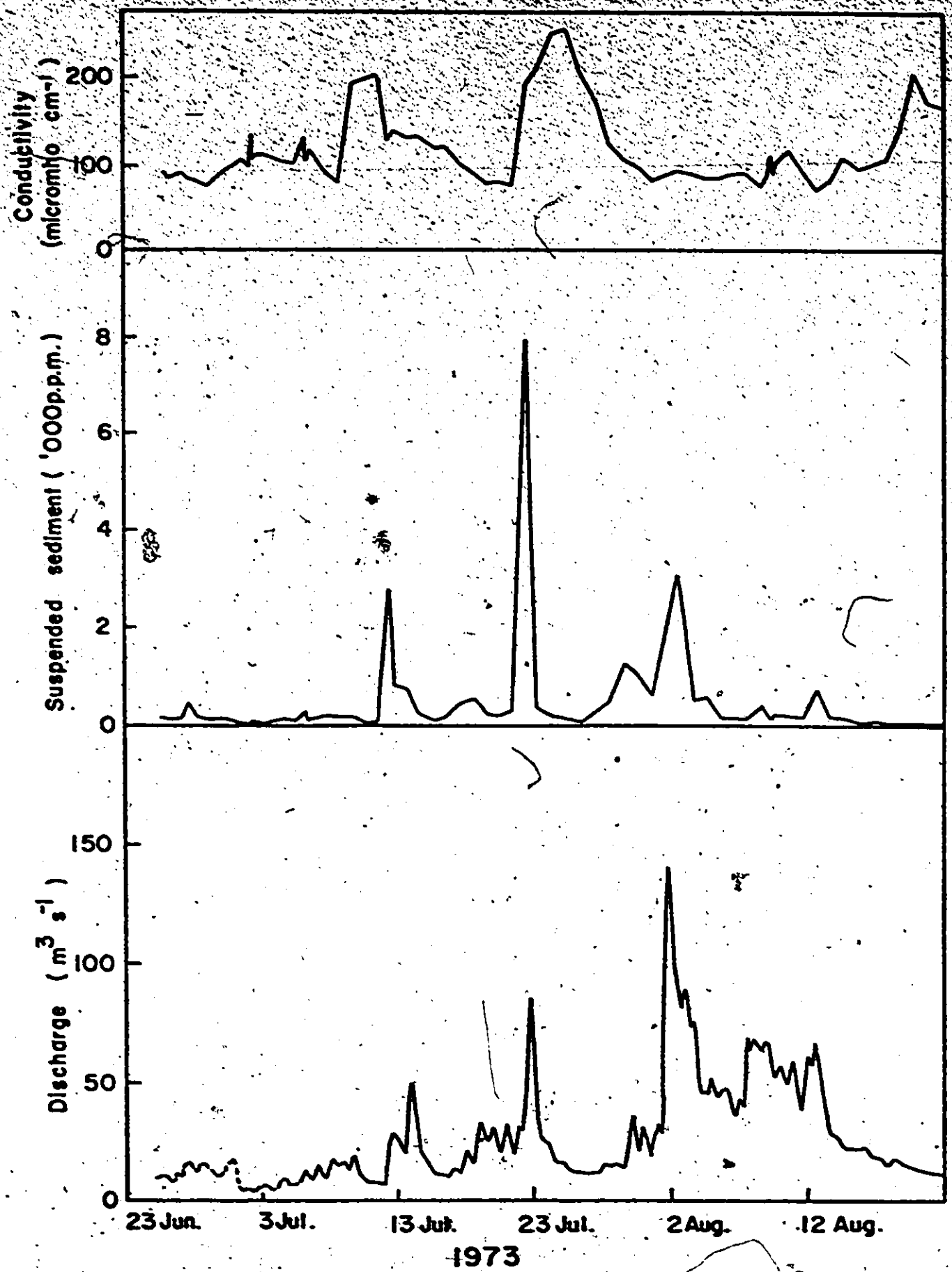


Figure 4:3a. Discharge and load, "Sverdrup River", 1973.

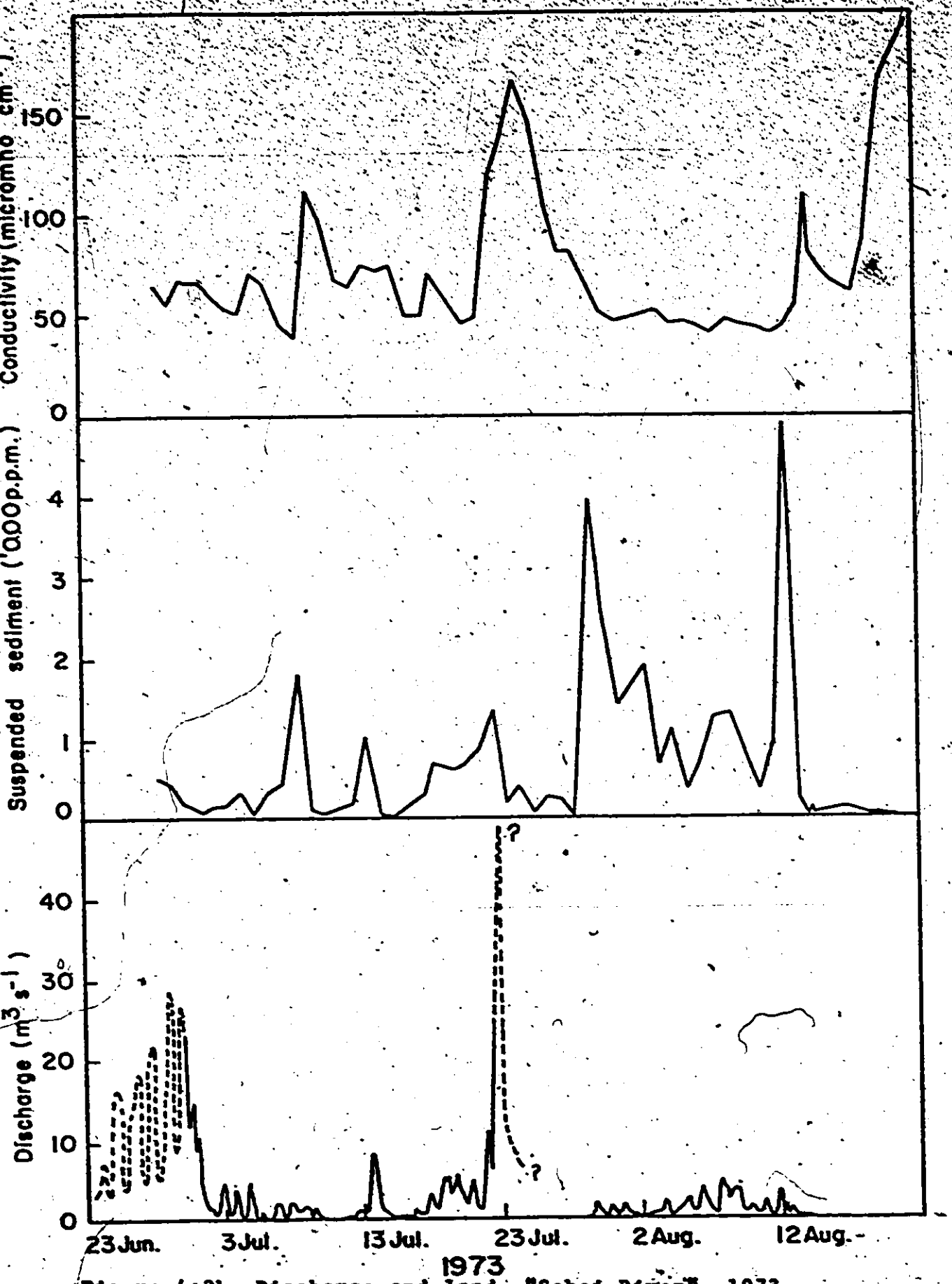


Figure 4:3b. Discharge and load, "Schei River", 1973.

The difference in water regime between these streams and the Mecham River arises because the two storage terms of eq. 3:1 - storage in the frozen state and storage in lake basins - are much more important on Ellesmere Island than on Cornwallis Island. In particular, the "Sverdrup River" hydrograph is affected by jökulhlaups: catastrophic emptyings or partial emptyings of ice-dammed lakes in the interior of the basin. A jökulhlaup on 1 August 1973 produced the largest discharge of the year, considerably greater than the response to a rainfall of 54.6 mm on 21-23 July. The 24-hour rainfall total for 22 July (49.4 mm) exceeds the maximum 24-hour total (42.4 mm) recorded in 26 years at the nearest weather station, Eureka, and the return period for the rainstorm as a whole may be over 50 years; yet there are indications that jökulhlaups similar to that of 1 August recur annually or almost annually.

The suspended sediment concentrations in both the "Sverdrup" and the "Schei" were very high during much of the 1973 runoff season. Most of the runoff water leaving the ice cap surface was free of sediment, and none was observed issuing from the base of the ice.

However, each stream leaving the ice flows over glacial detritus, generally at a high gradient, and quickly acquires a load of sediment from that source.

Concentrations of suspended sediment in the "Sverdrup" ranged from 36 to 8100 p.p.m. with a mean of 444 p.p.m. (N = 70), the higher concentrations being associated with the higher discharges. However, the highest discharge of the season - 146 $m^3 s^{-1}$ during the jökulhlaup - produced a concentration of only 3200 p.p.m., in marked contrast to the concentration recorded at the peak of the rainstorm flood response on 22 July - 8100 p.p.m. at 67.5 $m^3 s^{-1}$. Most of the stream load on 1 August must have been derived from the channel bed and banks, mainly in the steepest portion of the channel, in the gorge immediately downstream from the lake outlet. The high concentrations during rainstorm floods came not only from bed and bank sediment along a short length of channel but also from land surfaces and channel walls throughout the entire basin.

On average the "Schei" carried more suspended sediment per unit of volume than the "Sverdrup", mean concentration being 585 p.p.m. (N = 60) with extremes

of 16 and 4900 p.p.m. It is possible that the "Schei" was more turbid even than the "Sverdrup" after the rainfall of 22 July, but the sample for that day was taken from the "Schei" several hours after the flood peak and yielded only 1350 p.p.m. A higher average concentration in the "Schei" than in the "Sverdrup" would be expected because of its steeper bed slope and its consequently higher velocities of flow.

The suspended load carried by the "Sverdrup" between 25 June and 20 August 1973 amounted to 80.6 tonnes per square kilometre, a figure which should be rounded to 80 t km^{-2} considering the inaccuracies of sampling at the rate of about one sample per day. The figure is certainly an underestimate of the total load, for a substantial load must have been carried in the period before 25 June, and a lesser amount was probably removed after 20 August, Note that bed-load is not included in this estimate, and that dissolved load also has not been calculated, although readings of conductivity (Fig. 4:3) suggest solute concentrations somewhat lower than those found in the Mecham River.

It is known that the peripheral ice of the Ellesmere Ice Cap is frozen to its base, and also that

it is mostly very clean ice, suggesting that the proportion of the stream sediment load originating directly from the glacier ice is very small. (Whatever the direct glacial contribution, about 65 per cent of the total annual load of the "Sverdrup" was carried by rainstorm and jökulhlaup runoff on 22 July and 1 August.) If this is so, the ice-free area rather than total basin area should be used to calculate load per unit area, which yields a seasonal suspended load of 342 t km^{-2}. This would still be an underestimate of total seasonal denudation, but it would represent intense fluvial activity by any standards (v. infra): it corresponds, in a misleading but much-used notation, to a denudation rate or basin surface lowering of 132 mm per thousand years at a sediment density of 2.6 t m^{-3}. Of course it is known that surface lowering is not uniform and specifically, for instance, that the ~37000 tonnes carried past the measurement section on 1 August were the residue of net erosion in the 2-4 km long gorge downstream of "McMaster Lake 2" and probable net deposition in the 12-15 km of channel between the gorge and the sampling station.

The seasonal suspended load of the "Schei River" (from 28 June to 20 August) was calculated as 150 t km^{-2}. This figure, however, is even more uncertain than that for the "Sverdrup" because it includes a rough estimate (10 t km^{-2}) for the period 22-28 July, when no discharge records were gathered. On the assumption that direct glacial contributions to stream load are negligible, the seasonal suspended load increases to 232 t km^{-2}. This is lower than the comparable figure for the "Sverdrup" (342 t km^{-2}), but there are no permanent ice-dammed lakes, and therefore no jökulhlaups, in the basin of the "Schei": when the load carried by jökulhlaup waters is discounted, the figures for load per unit of ice-free area are 232 t km^{-2} for the "Schei" and 237 t km^{-2} for the "Sverdrup". These figures are very close to each other, as one would expect for two similar and neighbouring streams. For its size, the more steeply-sloping "Schei" may have done slightly more geomorphic work than the "Sverdrup", but the geomorphic significance of the jökulhlaup should be obvious.

4.2.2.3 Comparisons On the basis, then, of single measurement years (not the same), a very great contrast emerges between the "Sverdrup" and "Schei" on the one hand and the Mecham on the other. Even when the rare rainstorm of 22 July 1973 is allowed for, south central Ellesmere Island must be admitted to have a higher-energy geomorphic regime than Cornwallis Island; certainly its surface material is more readily erodible. South central Ellesmere Island in fact resembles central Baffin Island in both of these respects. Church (1972) studied two streams in central Baffin Island, the Lewis River in 1963-65 and the Ekalugad River in 1967-8. Suspended load was variable in the Ekalugad system, ranging in different sections of it from 0.8 to 49.1 $t \cdot km^{-2} \cdot y^{-1}$; the average annual suspended load of the Lewis River was 143 $t \cdot km^{-2}$. Table 4.1 contains information abstracted from Church's Table 21. Its most striking feature is the proportional distribution of load between the three fractions: bed load usually accounts for more than 80 per cent of total load, dissolved load for less than 10 per cent. The possibility that this proportional distribution in favour of bed load might also be found in south central Ellesmere Island

TABLE 4:1

Denudation in Central Baffin Island

Stream	Bed Load*	Suspended Load	Dissolved Load†	Total Load
Lewis R, 1963	1049	233	5.1	1286
Lewis R, 1964	236	68	2.5	307
Lewis R, 1965	483	127	3.1	613
Lewis R, mean	589	143	3.6	735
Ekalugad R.:				
Middle, 1967	570	36	2.4	609
Upr. South, 1967	1114	49	4.7	1168
Upr. South, 1968	29	0.8	2.1	32

All figures are t km^{-2} yr^{-1}

*Calculated from the equation of Meyer-Peter and Müller (1948)

†From laboratory measurements of individual species.

is now being investigated. Church's measurements led him to conlude that the relatively small contribution of solutes to total load was "the characteristic by which arctic fluvial activity is distinguished from that of all other regions". This conclusion, eminently reasonable on the granites and gneisses of Baffin Island and possibly also in south central Ellesmere Island, where some of the bedrock is calcareous, is evidently not apt for the Mecham River, as will be seen later.

To place the limited results from the High Arctic in a wider context, it is necessary to consult the records of officially-maintained programmes of suspended sediment sampling. In Canada there is a network of sediment sampling stations for which records are published by the Sediment Survey of Canada. Most of these stations are in the south of the country, and there are none in the North West Territories with "good" records. When stations whose records are officially described as "poor" are eliminated, there remain twenty four stations with "fair" or "good" records and, of these, four are greatly influenced by artificial modifications.

To explore the relationship between stream gradient and sediment load, the records of the remaining

twenty stations were fitted to a least squares model with total basin relief as the independent variable (Table 4:2). Suspended sediment load for the model was taken as the mean of annual loads in the period 1965-1967, although not all of the stations were in operation throughout the period. The corresponding basin relief is the difference in altitude between the highest point in the basin and the sediment energy gradient which partially determines the load of sediment carried. More sophisticated measures of the gradient are conceivable, but they would be time-consuming to obtain and more elaborate than is justifiable for this simple model. The rationale for the use of basin relief is that, other things being equal, the solid sediment supplied to a stream comes either from the bed of its channel or the surface of its drainage basin. The amount arriving from these two sources depends at least partly upon their slopes, and although the contributions from the two are hard to separate and may vary in relative importance, total basin relief should be a rough measure of the two in combination. The model, based on the twenty official data points, gives annual suspended load

TABLE 4:2

Suspended Sediment Data for Canadian Streams

Stream	Area (km^2)	Relief (m)	Load ($t\ km^{-2}\ y^{-1}$)	Period of Measurement
Columbia at Revelstoke, B.C.	26936	3312	115	1967
Fraser at Hope, B.C.	202796	3911	114	1965-7
Kootenai at Porthill, B.C.	35483	3088	70.7	1967
Streater Main Stem at Nanton, Alta.	5.96	381	12.7	1967
Oldman at Brocket, Alta.	4403	2093	205	1967
Red Deer at Bindloss, Alta.	43512	2720	55.7	1967
Willow at Claresholm, Alta.	1155	1423	120	1965-7
South Saskatchewan at Hwy. 41, Alta.	66304	2686	113	1967
South Saskatchewan at Hemsford, Sask.	116550	2742	98.0	1965-7
Pembina at Windygates, Man.	7822	379	11.0	1965-7
Red at Ste. Agathe, Man.	116550	524	86.2	1965-7
South Tobacco at Miami, Man.	94.5	200	121	1965-7
Turtle at Laurier, Man.	1083	433	31.2	1966-7
Peace at Peace R., Alta.	186479	3566	168	1967
Big Otter at Vienna, Ont.	697	152	124	1967
Thames at Ingersoll, Ont.	518	119	18.5	1965-7
Kennebacsis at Apohaqui, N.B.	1101	405	17.8	1967
North Nashwaaksis at Royal Rd., N.B.	26.9	241	6.42	1966-7
Fraser at Archibald, N.S.	9.12	94	1.45	1967
Brudenell at Brudenell, P.E.I.	35.5	73	13.3	1967

$$L_s = 1.6R^{0.5} \qquad (4:6)$$

with $r = 0.658$, $100r^2 = 43$ per cent. The predicted value of L_s for the Mecham River, with a relief $R = 194$ m, is 22.3 t $km^{-2}\ y^{-1}$, to be compared with a measurement of 12.7 t $km^{-2}\ y^{-1}$. The crudity of the model is underscored by the result of an attempt to relate load to the quantity

$$R/(\sqrt{2A}) ,$$

where A is basin area in m^2. This quantity should be a better measure of basin slope than R alone, yet it is not significantly correlated with suspended load.

In general, the Mecham carried a load similar to those of streams in the Maritime Provinces but substantially smaller than those of streams in the Prairie and Cordilleran regions, where energy gradients are high and/or regoliths are thick. The loads of the "Sverdrup" and "Schei" were greater than any of those in Table 4:2 and much greater than those predicted by eq. 4:6 (their respective basins have reliefs of ~1500 and ~750 m, implying loads of 40-200 t $km^{-2}\ y^{-1}$).

Langbein and Schumm (1958) investigated the relationship between the suspended load of streams and basin precipitation. They took as raw material information collected by the U.S. Geological Survey on suspended sediment concentration and discharge for streams in the United States. Recognizing that the erosive effectiveness of a given amount of precipitation would not be the same everywhere, they used a variable which they called "effective precipitation": the amount required over any basin to produce the measured amount of runoff, allowing for evaporative losses which are regarded as a function of temperature. These losses were accounted for with a relationship graphed by Langbein (1949) for a mean annual temperature of 50°F (10C), which is well fitted by the cubic

$$P_{eff} = 225.8 + 4.076Q - 6.14 \times 10^{-3}Q^2 + 3.48 \times 10^{-6}Q^3 \quad , \qquad (4:7)$$

where P_{eff} is effective precipitation and Q measured runoff, both in mm.

There is great scatter in the data sets used by Langbein and Schumm, attributed by them to topographic and geologic factors. However, by grouping the data they argued that the relationship between effective

precipitation and sediment yield could be seen to be twofold: the more precipitation, the more energy available and therefore the more erosion; but on the other hand, the more precipitation the more biomass to deflect the energy and therefore the less erosion. Langbein and Schumm argue that there should be zero load at zero precipitation, that peak yield occurs at ~250-300 mm effective precipitation (which in effect means in the semi-arid southwestern U.S.A.), and that yield decreases up to ~1000 mm and remains nearly constant as precipitation increases above 1000 mm., The ungrouped runoff data on which this argument is based are graphed in Figure 4:4, and the curve drawn on the figure approximates and paraphrases what has been elevated in Fairbridge (1968; cf. article by Ritter, p. 173) to the status of the "Langbein-Schumm Rule".

The "rule" is intuitively reasonable, implying that mechanical erosion by water is greatest in arid and semi-arid shrublands and grasslands, least in the very dry deserts and moderate in the forested regions of the world. It must be re-examined, however, in the light of the data presented here and that gathered by Church. Following eq. 4:7, P_{eff} for the Mecham in 1971 was 968 mm, a figure which is too large;

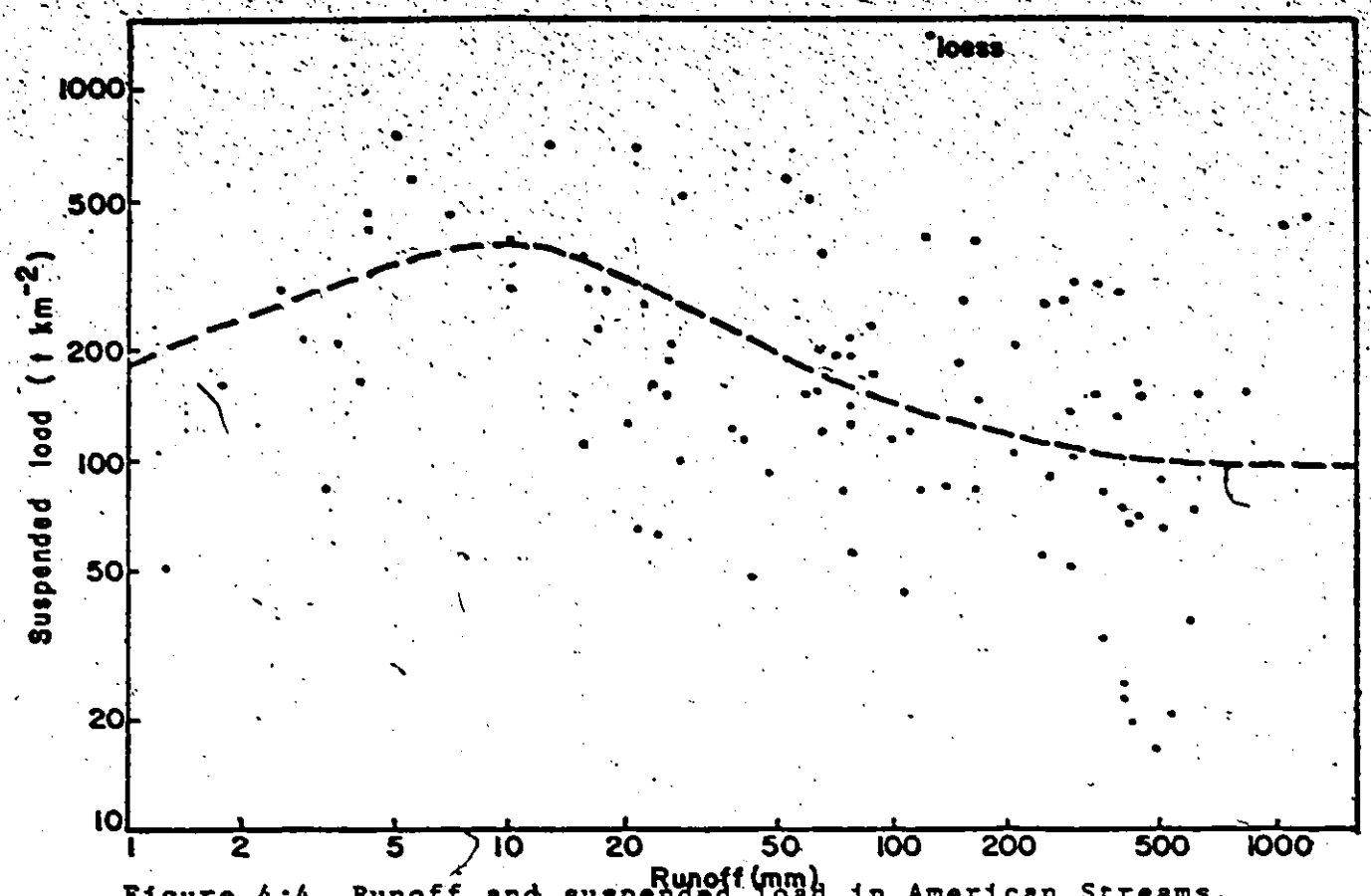


Figure 4:4. Runoff and suspended load in American Streams.

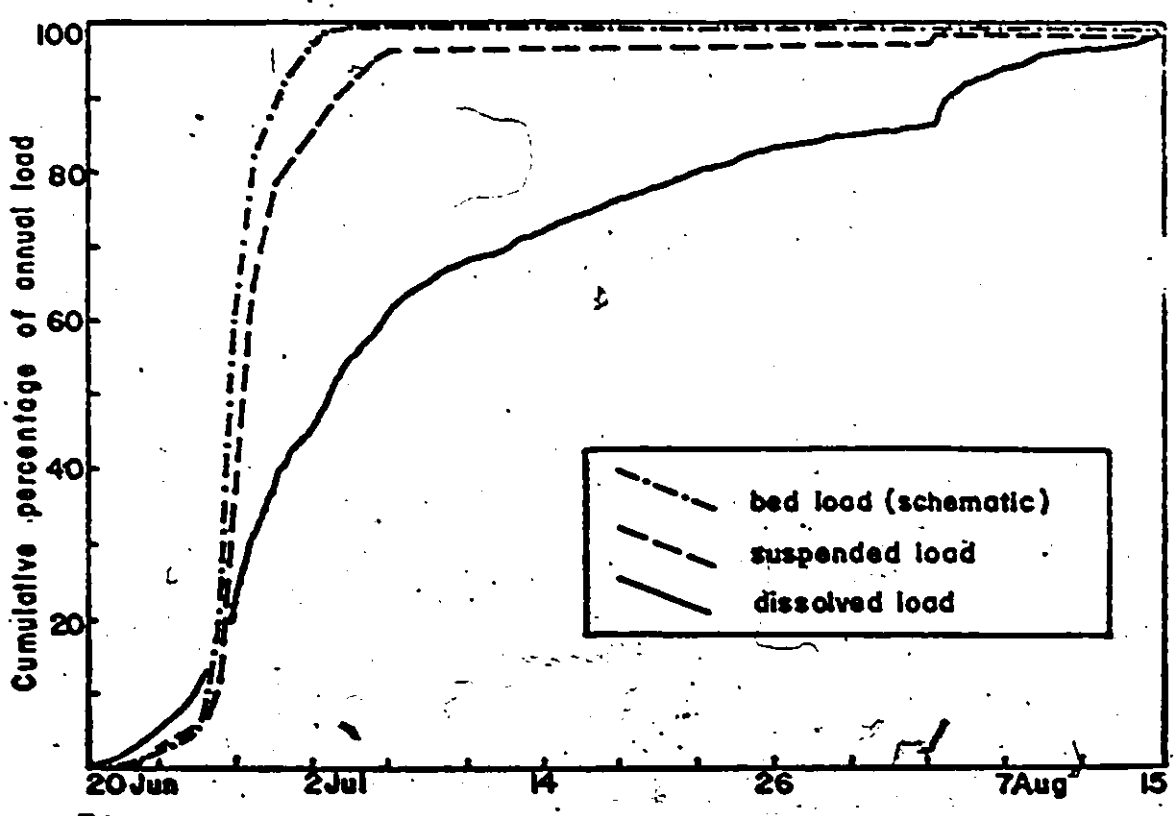


Figure 4:5. Cumulative load, Mecham River, .971.

following eq. 3:1 we get an effective precipitation of 601 mm, at which the rule requires high sediment yield; even following the spirit rather than the letter of the rule, we should expect a very high yield from the barren surface of a basin in a semi-arid climate. Yet the Mecham River in 1971 carried a load which was very low judged by any of these criteria. The "Sverdrup" and the "Schei" in 1973, and the Lewis in 1963-5, carried loads in keeping with the spirit of the Langbein-Schumm Rule but not with its letter, for the "effective" precipitation in these cases was concentrated onto relatively small areas of terrain from larger areas covered by ice.

Two factors seem to be important for an explanation of how the Langbein-Schumm Rule applies in Arctic terrains. The first is geological and topographic: the gentle slopes and coarse regolith of southeast Cornwallis Island make it a region which is resistant to mechanical erosion; for opposite reasons, and also because glacial meltwater is an important component of its geomorphic regime, south central Ellesmere Island is a region which is susceptible to erosion. Roughly speaking, sediment yield in High Arctic drainage basins should be higher in areas in the east

which have greater relief than does southeast Cornwallis Island, and higher generally in areas thickly mantled by drift or floored by relatively incoherent clastic sediments of Mesozoic and Cainozoic ages. Such sediments are widespread in the western High Arctic and in parts of Ellesmere and Axel Heiberg Islands. Geological and topographic controls of this kind were recognized by Langbein and Schumm but deliberately circumvented by their grouping procedure.

The second factor concerns the precipitation-yield relationship in general. The geomorphic analogy between high-latitude "deserts" and the true deserts and adjacent semi-arid regions is appealing; superficially, the sparseness or absence of vegetation, and the availability of only moderate amounts of precipitation, suggest that the same quantity of incoming precipitation should do the same amount of work in whichever of the two kinds of regime it falls. The situation in detail suggests otherwise.

All of the precipitation in low-latitude semi-arid regimes falls as rain in storms of greater or lesser intensity; the classical magnitude-frequency relationship expressed in the "Wolman diagram" (cf. Leopold, Wolman and Miller, 1964, p.80) should hold here, and most of the geomorphic work will be done

during events of moderate magnitude and moderate frequency. More importantly, however, a large contribution to the total erosion will be made by the impact of the rain on the surface. Both raindrop mass (Laws and Parsons, 1943) and raindrop velocity increase with rainfall intensity, and intensity increases with annual amount of rainfall, so the energy applied to the basin surface should increase with at least the square of rainfall amount because of the velocity term. (Interestingly, Langbein and Schumm fitted to their data a curve which increased with the power 2·3 of effective precitipation.)

In high-latitude semi-arid regimes, less than half of the precipitation normally falls as rain. At once this implies that a different kind of relationship will hold between the magnitude and frequency of geomorphic events. Many insignificant winter events are summed into a single large event which occurs in spring once a year. Usually this annual snowmelt flood is the largest event of the year in unglacierized basins and, insofar as this is the case, the magnitude-frequency problem degenerates to the problem of assessing long-term climatic variability: most of the geomorphic work is done during events of a known high frequency and an unknown moderate magnitude.

The concentration of winter precipitation into the spring flood tends to increase the amount of work done, but against this tendency must be set the fact that the impact of the precipitation on the surface is negligible. Wilkinson (1972) has shown convincingly that snow meltwater can and does move substantial quantities of sediment on Arctic slopes, but it is well known that rainwater runoff is also effective in moving sediment on slopes. In the High Arctic the kinetic energy of raindrop impact is worth considering only on two or three occasions each year, and this is probably the most significant difference between high- and low-latitude semi-arid geomorphic regions.

To restate this contrast, the same quantity of incoming precipitation should do less work in a region such as the Queen Elizabeth Islands than it would do in a region such as the American South West. South central Ellesmere Island is partly a special case because it is glacierized, but the evidence from the "Sverdrup River" and "Schei River" suggests that the work done by running water may be greater per unit of incoming precipitation in many high-latitude drainage basins than in grass-covered and forested mid-latitude

catchments such as are found in southern Canada (Table 4:2) and the northern U.S.A.

The problem of assessing the amount of fluvial activity in the High Arctic, and comparing the region with others in different regimes, is, however, less important than the most basic point which has been established by this and by some earlier investigations: that water is the major agent of landscape lowering at high latitudes and that, moreover, it is very probably the major agent of geomorphic changes in general.

4.2.3 Dissolved Load

4.2.3.1 Controls Most of the sediment removed by the Mecham River from its basin in 1971 was carried in solution. Generally, one should expect dissolved load to predominate over solid load in calcareous terrain where relief is low. The controls upon transportation of material in solution are, however, quite different from those on transportation of suspended material. Solution, like most chemical processes, proceeds at a rate which is a function of temperature, and temperature should therefore be a major and direct control on removal of dissolved material from the landscape. More important than the rate of solution, however, is the solubility of most salts in water: this also is a func-

tion of temperature. Another control on solution is the availability of water; in all but extremely dry geomorphic regimes the water supply is not a limiting factor, yet there is one important effect related to water supply which bears on the magnitude-frequency distribution of solute transport. As discharge in a stream channel increases, the rate at which a unit volume of water is able to interact with a unit area of channel wall decreases, and opportunities for solution also decrease. The concentration of solutes, although not the load, therefore decreases with discharge, and a higher proportion of the geomorphic work is done during low magnitude-high frequency events. In the Mecham River, for instance, dissolved load is the only significant component of total load during most of the flow season (Table 4:3, Figure 4:5).

While availability of water is not a limiting factor, the availability of soluble solids often is limiting. It is probable that most stream waters are unsaturated with respect to those mineral species which are both common in rocks and soluble in water, and , for this reason if for no other, dissolved load should be comparatively small in igneous, metamorphic and clastic sedimentary terrains. In limestone and dolomite terrains, however, the major soluble species are abundantly available to runoff waters, and the ratio

TABLE 4:3

Magnitude and Frequency of Water and Sediment Flows, Mecham River, 1971 Measurement Season

Daily mean discharge ($m^3 s^{-1}$)	Time	Total discharge	Percentage of Total dissolved load	Percentage of Total suspended load
0-2	48	11	16	1
2-5	25	14	18	2
5-10	12	15	20	9
10-20	8	22	19	19
20-40	7	38	27	69

Time = 60 days

Mean discharge = 3.5 $m^3 s^{-1}$

*Total dissolved load = 20.26 t km^{-2}

*Total suspended load = 12.72 t km^{-2}

*These figures apply to the measurement season only

of dissolved to suspended load should increase with water temperature and decrease with surface gradient. The dissolved load itself, as the product of discharge and concentration, should increase with effective precipitation (approaching an upper limit) and water temperature (which could be approximated by air temperature or even latitude in a coarse analysis).

There is, however, one more important control on solution, which applies in the most common situation where the solute source is limestone or dolomite. This final control is the availability of CO_2, which increases very greatly the solubility of carbonate minerals in water. The two major sources of CO_2 are the atmosphere and the vegetation cover of a given terrain. Again, in a coarse analysis, vegetation cover or biomass could be approximated by air temperature, or alternatively by latitude corrected for elevation. However, even in detail significant control of limestone solution by vegetation cover may occur.

4.2.3.2 Comparisons The dissolved load measurements made on the Mecham River should be seen in two contexts: they should be compared, firstly, with values for other streams irrespective of drainage basin lithology and, secondly, with values for other limestone and dolomite

regions. Table 4.4 is a list of data on dissolved load for streams in the U.S.A., taken from Leopold, Wolman and Miller (1964, p. 76), with Mecham River and Alberta information included for comparison.

The streams in Alberta were studied by Drake (1974), and the loads are calculated from regressions of monthly sample results on discharge. The two figures for each stream are estimates based on two assumptions: that the entire basin contributes to the load, and that contributions come only from carbonate outcrops (which cover 31.4 per cent of the Athabasca basin and 33.4 per cent of the North Saskatchewan basin). Leopold, Wolman and Miller refer to unpublished work by Langbein and Dawdy on a data set similar to that of Langbein and Schumm (1958); the results of this research suggest that dissolved load increases with runoff to a maximum of 60-80 t km^{-2} y^{-1} in the runoff range above 250 mm, which corresponds in the United States to an "effective precipitation" of roughly 600-1000 mm. In the Canadian Arctic 250 mm of runoff is approximately equal to the same amount of effective precipitation. The Mecham River transports more dissolved material than some of its semi-arid counterparts in the American South West, yet it seems clear

TABLE 4:4

Dissolved Load in American Streams

Stream	Runoff (mm)	Dissolved Load ($t\ km^{-2}\ y^{-1}$)	No. of Years of Record	Dissolved Load / Suspended Load
Little Colorado at Woodruff, Ariz.	2.69	1.01	6	0.012
Canadian at Amarillo, Tex.	11.0	2.71	1	0.019
Colorado at San Saba, Tex.	16.2	2.85	5	0.069
Bighorn at Kane, Wyo.	51.7	5.80	1	0.136
Green at Green River, Ut.	57.2	27.0	20	0.132
Colorado at Cisco, Ut.	121	78.8	20	0.293
Iowa at Iowa City, Io.	160	62.7	3	0.410
Mississippi at Red River Landing, La.	171	37.1	3	0.358
Sacramento at Sacramento, Cal.	319	35.4	3	0.804
Flint at Montezuma Ga.	421	19.4	1	0.330
Juniata at Newport, Pa.	445	71.9	7	1.758
Delaware at Trenton, N.J.	596	51.5	4	0.828
Mecham at Resolute, N.W.T.	284	20.3	1	1.593
Athabasca at Hinton, Alta.	566	68.2-137	2	1-2?
North Saskatchewan at Rocky Mtn. Ho., Alta.	404	57.9-113	2	1-2?

that thermal effects are at work reducing its dissolved load relative to that carried by mid-latitude streams with similar amounts of actual runoff. The streams in Table 4:4 all drain large basins (from 7500 to 3×10^6 km^2 in area) and so are not strictly comparable with the Mecham River; more important than disparity in size, however, is variety of lithology. More striking relationships emerge when the Mecham River is compared with other limestone and dolomite streams: a clue to the form of these relationships is given by the Juniata River, which drains much of the Appalachian karst of central Pennsylvania and which removes over three times as much dissolved material as the Mecham River with less than twice the runoff. The number of good published figures for dissolved load in carbonate terrain is very small, so discussion must be based on solute concentrations rather than loads.

Climatic controls on limestone solution have been the subject of controversy every since Corbel (1959) argued that rate of removal should decrease with temperature because the solubility of CO_2 also decreases with temperature. Corbel's notion has been effectively refuted by the theoretical reasoning of Bogli (1960) and by an accumulation of empirical evidence to the

contrary (e.g. Ford, 1971, Cogley, 1972). It is now known that the dependence of water solvent capacity on temperature is far more important than the inverse dependence of CO_2 solubility on temperature, and moreover that the enrichment of water with biogenic CO_2 is of great importance in itself. The extent of this enrichment is a function of the thoroughness with which the water is able to mix with organic matter before and during the solution process; this also is grossly dependent upon temperature through the dependence of biomass on temperature.

The gross relationship between solute concentration and temperature is well illustrated in recent work by Harmon, White, Drake and Hess (1975). These workers argue, following Drake and Harmon (1973), that waters draining calcareous terrain should be segregated into types based on position within local hydrogeologic environments: surface runoffs constitute one type, but they choose to work exclusively with waters from conduit flow and diffuse flow springs. This selectivity, and the procedure of grouping many analyses to produce mean values for large geographical regions, enables them to remove much of the noise which obscures the wider relationships in which they are interested. They find

good linear relationships between water temperature and each of the dependent variables Ca^{++}, HCO_3^- and log P_{CO_2} (where P_{CO_2} is the equilibrium partial pressure of CO_2 calculated from pH measurements); illustrative data on some of these relationships are given in Table 4:5 below, to which are added comparable data for analyses of Mecham River waters.

TABLE 4:5

Chemistry of North American Spring Waters

Region	Ca^{++} (mol m^{-3})	HCO_3^-	(deg C)	(N m^{-2})
Alberta	0.52	1.22	3.3	32.8
Virginia	0.80	1.84	9.5	343
Pennsylvania	1.13	2.91	9.7	532
Kentucky	1.18	2.63	11.7	376
Missouri	1.78	4.15	14.8	1060
Texas	2.14	5.07	22.2	2660
Mexico	3.75	4.72	23.5	3060
Surface runoff, Mecham basin*	0.51	1.37	3.0	45.0

*Only samples with ion balance error < 5% used; mean error = 2.90%

The mean solute concentrations in the Mecham River are close to the means for springs in the alpine karst of the Alberta Rockies. In the High Arctic there is no such thing as a spring with which to compare Alberta waters. However, since most of the Alberta springs are in alpine karst, and are fed by rapid conduit flow of water having little contact with biogenic CO_2, similarity between the two types in the two regions is to be expected.

Considering the results of work which has been done in the past decade, some of which are mentioned above, the conclusion that solute concentration (and therefore solute load) increases with temperature must now be regarded as quite unequivocal. There is, however, good evidence that the relationship is linear, and on physical grounds one should expect a finite amount of solute removal even at the freezing point. Because of the inverse relationship between solute concentration and discharge, dissolved load should always be a less variable quantity than suspended or bed load.

Other things being equal, then, the amount of chemical erosion in the High Arctic should be less,

though not by much, than at lower latitudes. Depending upon other properties of the arctic terrain, chemical erosion may vary greatly in relative importance. In non-carbonate basins it will be moderate or insignificant except where the relief is extremely low or the surface is otherwise very resistant to mechanical erosion. In carbonate basins in the High Arctic, chemical erosion will be moderate to overwhelming in importance.

4.2.3.3 Details of the Solution Process The course of solution during the 1971 runoff season is shown in detail in Figure 4:6, on which are plotted discharge, Ca^{++} concentration, saturation with respect to calcite, equilibrium partial pressure of CO_2 and pH. The changes in Ca^{++} concentration are representative of changes in concentration of the other major species, Mg^{++} and HCO_3^-, and the observation that there are about 175 measurements of these three species (with 60 measurements of Na^+, 9 of $SO_4^=$ and Cl^- and 3 of Si) should inspire some confidence in the composite figure for dissolved load discussed in the preceding section. Too much confidence would be misplaced, but at least the sampling density was adequate.

Some features of the solution process which are obscured by a general treatment are brought to light

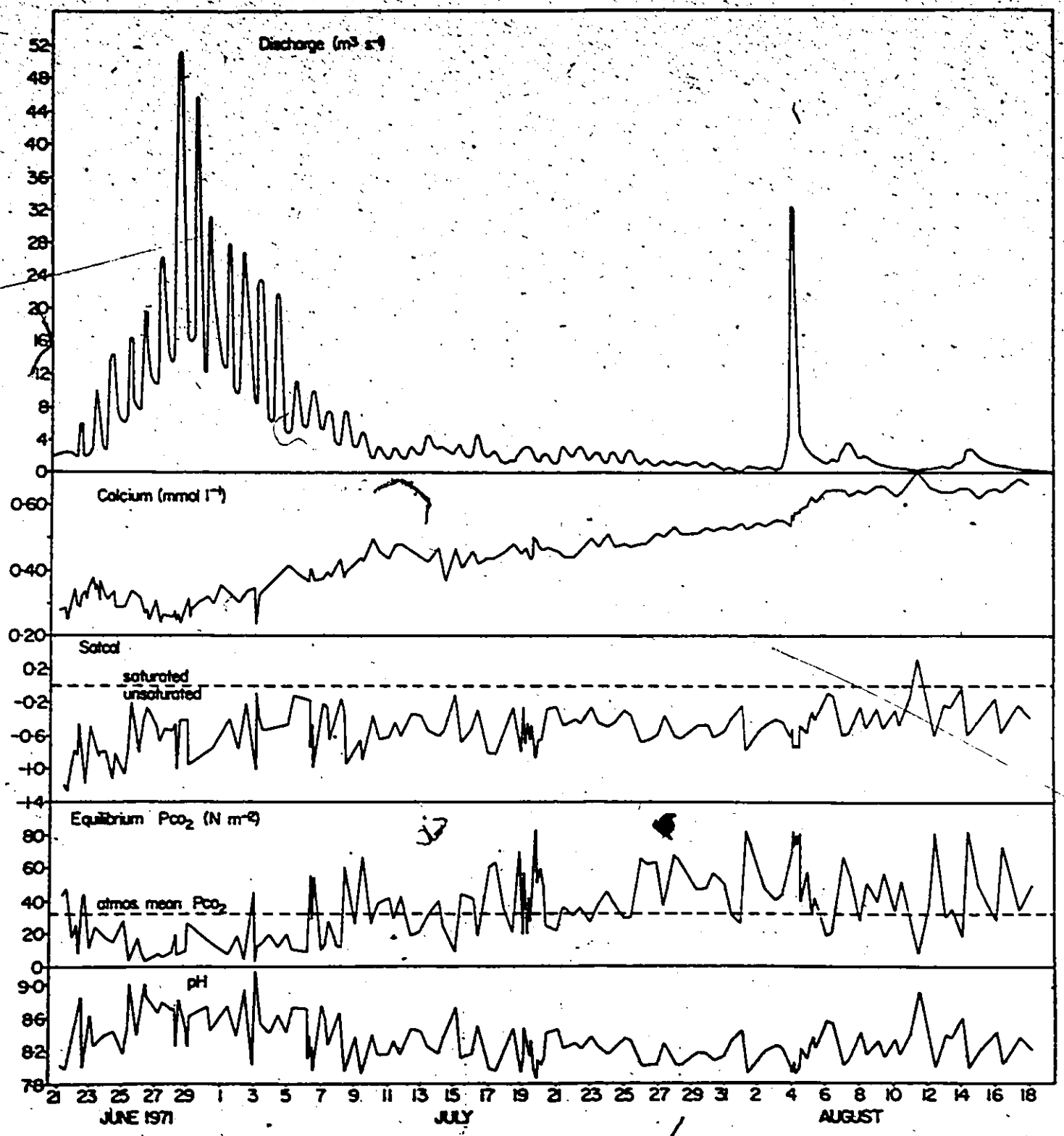


Figure 4:6. Chemical properties of runoff water, Mecham River, 1971.

or hinted at in the several graphs of Fig. 4:6. The composition of the dissolved load is one of these features. For all practical purposes the Mecham basin waters can be considered to be of the calcium-bicarbonate type, as is shown in Figure 4:7 (slightly modified for illustrative purposes).

Table 4:6 shows analyses of a small number of samples from snow and rainwater. The analyses suggest that most of the Na^+, $SO_4^=$ and Cl^- in solution in the Mecham River has an immediate origin in the atmosphere, and that very little solution of minerals other than calcite and dolomite takes place within the basin. Cogley (1972) reported that there is very little K^+, Al or Fe in the waters of "Jason's Creek", which drains over the same beds as the Mecham River, and in 1971 three tests of Mecham River water showed Si concentrations of 0.004, 0.004 and 0.007 mol m^{-3}.

The molar ratio of Ca^{++} to Mg^{++} is illustrated in Figure 4:8, which has points for all samples taken in southeast Cornwallis Island in 1970-72. The mean ratio is 3.56, similar to the ratios measured in southwest Devon Island by Cogley (1971, Fig. 8:4) and higher than ratios commonly found in lower latitudes. The high Ca: Mg ratio is partly a lithological effect and

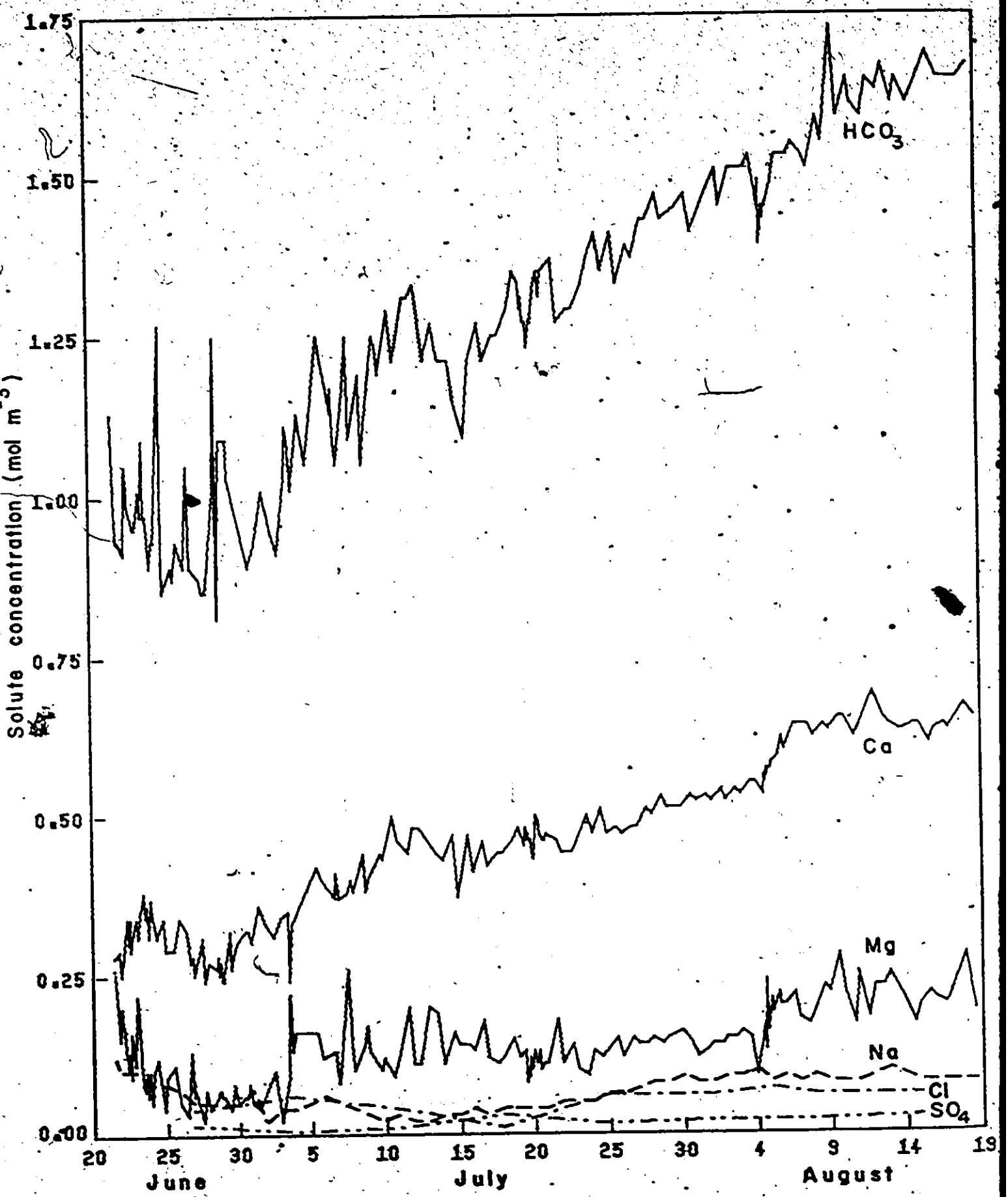


Figure 4:7. Solute concentrations, Mecham River, 1971.

TABLE 4:6

Chemistry of Rainwater and Snow

Sample location and date	pH	Ca	Mg	Na	K	HCO_3	SO_4	Cl
				(mol m^{-3})				
"Jason's Creek", 280770	6.00	0.06	0.06	0.02	0.01			
Mecham R., 010871*	6.44	0.05	0.02			0.04	0.003	0.10
Mecham R., 040871	6.42	0.06	0.01			0.01	0.002	0.01
Mecham R., 140871	6.64						0.003	0.01

*Snow

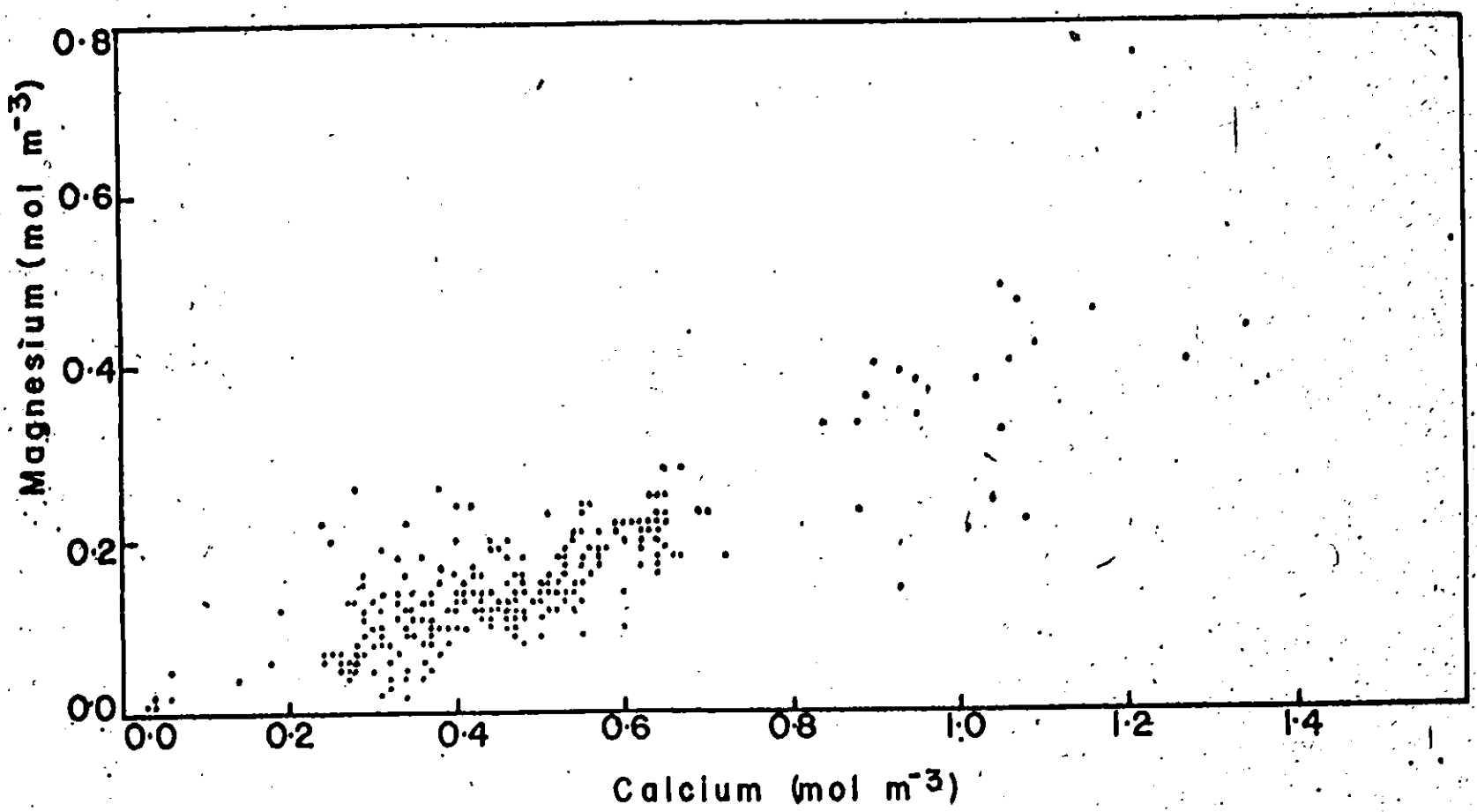


Figure 4:8. Calcium/magnesium ratios, south-east Cornwallis Island, 1970-72.

partly due to differences in the solubility of calcite and dolomite: dolomite is less soluble than calcite, but the dependence on temperature is more marked in the case of dolomite, such that proportionally less of it is dissolved from the same carbonate rock as temperature decreases.

The dilution which results as discharge increases is illustrated for Ca^{++} in Figure 4:9. This aspect of limestone solution in the High Arctic was documented by Cogley (1971, 1972); suffice it to say here that the relation

$$Ca^{++} = 0.49Q^{-0.16}, \quad r = 0.760, \quad N = 119 \qquad (4:8)$$

noted for the Mecham River in 1971, is very similar to that noted in the earlier work just mentioned ($Ca = 0.42Q^{-0.12}$). The difference in the two types of sample plotted on Fig. 4:9 will be referred to in a later paragraph.

The graph of P_{CO_2} in Fig. 4:6 offers interesting insights into the process of solution itself. The first notable feature is that mean P_{CO_2} for all samples on the graph is 31.6 N m^{-2} (45.1 N m^{-2} for samples with $I_e < 5$ per cent); away from sources of pollution, global atmospheric mean P_{CO_2} is at present about 31-32 N m^{-2}.

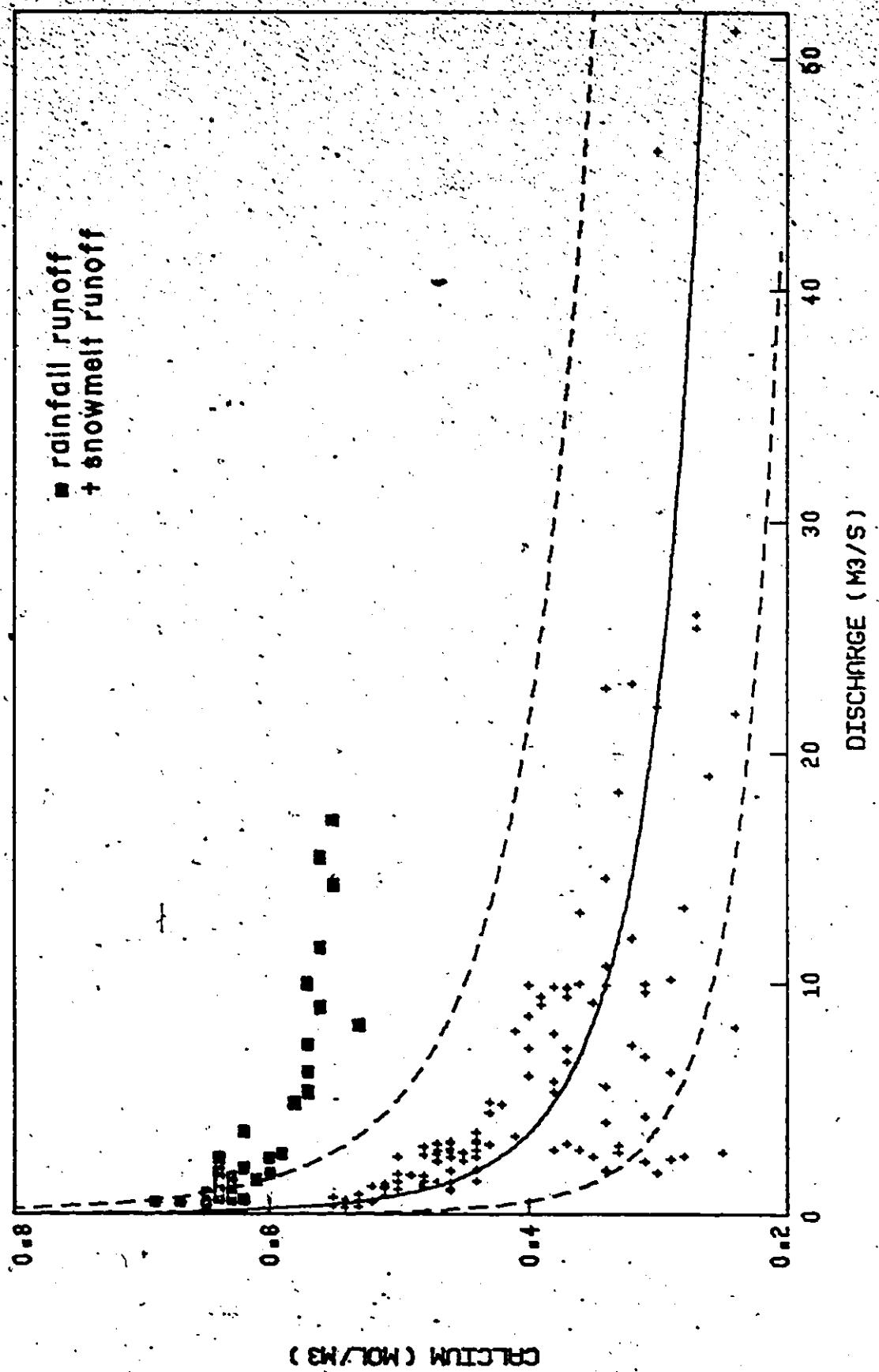


Figure 4:9. Discharge and dissolved calcium, Mecham River, 1971.

Clearly there is very little enrichment of the Mecham waters by organic CO_2.

The expression "P_{CO_2}" denotes (equilibrium) partial pressure of CO_2; for a water, if Henry's Law applies, this is the pressure which would be exerted by the gaseous CO_2 in a hypothetical atmosphere with which the water would be in equilibrium. In most natural waters, equilibrium CO_2 pressures much greater than 31-32 $N\ m^{-2}$ are compelling evidence for admixture of biogenic CO_2. Soil percolation waters in grassed and forested mid-latitude areas often have CO_2 pressures 100 or more times the global atmospheric mean.

The difference between the all-sample mean P_{CO_2} given above and the mean for samples with low ion balance error is partly explained in Figure 4:10a, and partly in the seasonal trend of Fig. 4:6. P_{CO_2} is low early in the flow season, declining to as low as one-tenth of the global mean at the height of the snowmelt flood; during the remainder of the summer P_{CO_2} increases erratically, and in August most values are between 1 and 2.5 times the global mean. This moderate seasonal increase is best explained as an effect of the very modest increase in photosynthetic

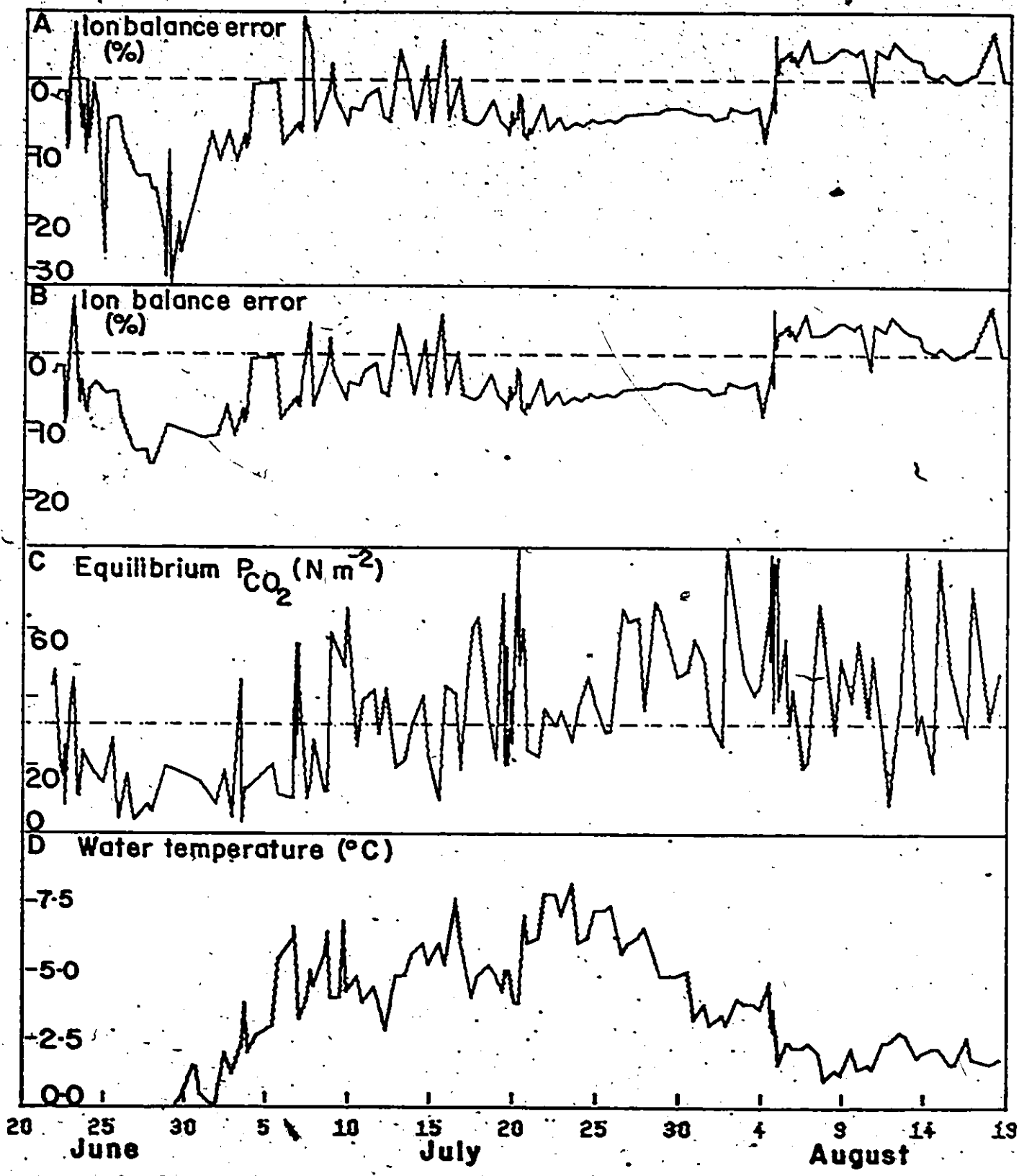


Figure 4:10. Ion balance error and related properties, Mecham River, 1971.

activity which occurs through the summer, but the relative depletion of CO_2 in snowmelt flood waters requires special attention.

Unfortunately, as Fig. 4:10 shows, many of the samples taken during the flood had high negative ion balance errors, implying either underestimation of Ca^{++} and Mg^{++} or overestimation of HCO_3^-. Accurate measurements of HCO_3^- (and pH) are essential for accurate calculations of P_{CO_2}, the latter being a function of temperature and of hydrogen and bicarbonate ion activities. An extended discussion of the difficulties of pH and alkalinity measurement was given in Section 1, where it was concluded that the pH measurements used here have accuracies of +0.04-0.02 pH units, and that most measurements of HCO_3^- were accurate to $\pm$ 0.02 mol m^{-3}. The seasonal fluctuation of pH shows an erratic decrease through the summer, but it was generally high throughout the flood, rising above pH 9.00 - a value rarely exceeded in unpolluted natural water - on two occasions. This is one reason for low P_{CO_2}.

The question of the accuracy of alkalinity measurements is difficult to answer, for the reasons

discussed in Section 1. If analyses with ion balance errors greater than 5 per cent are rejected, very few analyses from the flood period survive. Without doubt, some of these analyses have produced bad data, but it is possible to save many of them by adopting a more lenient criterion for rejection, on the ground that measurement error itself is large as a percentage when concentration is low:

$$|I_e| > 5 + 100 \frac{0.20}{C_+ - C_-} , \qquad (4:9)$$

where C_+ and C_- are total cation and anion charges respectively. Of about 175 samples through the season, 153 pass this test, and their P_{CO_2} values are graphed in Fig. 4:10b. They still illustrate CO_2 depletion during the snowmelt flood.

The explanation for this phenomenon is probably related to that offered for similar observations by Ek (1964, 1966) and supported by the measurements of Clement and Vaudour (1968) and Ford (1971). This is that CO_2 is expelled during the compaction of snow and its transformation to ice, a hypothesis which is also supported indirectly by the work of Kelley Weaver and Smith (1968) and Coyne and Kelley (n.d.; 1972) at Barrow, Alaska. These workers measured gaseous CO_2

within and at the bottom of snowpacks, and in the ambient air, with infrared gas analyzers. They found that there was always a CO_2 pressure gradient from the tundra surface through the snow to the ambient air, and concluded that the soil was the source of the gas. However, it seems equally if not more probable that the source was in fact the snow covering the soil.

Whatever the mechanism of CO_2 depletion, it can now be considered well established that snow and ice meltwater is on the average less aggressive w.r.t. calcite than rainwater. A disparity between the two water types was noted at "Jason's Creek" in 1970 and was discussed by Cogley (1972, Fig. 4); the same disparity was noted for the Mecham River in 1971, but in considerably more detail (cf., in Fig. 4:6, the solutional events accompanying the rainstorm of 4 August, and in particular see Fig. 4:9). The effect of the disparity on total dissolved load, as considered in previous sections, is to reduce the load as the ratio of snowmelt to rainfall runoff increases, but the amount of the reduction is very small. Effects may, however, be noticeable in karst landform genesis.

4.3 Summary

Observations made principally during the summer of 1971 showed that the Mecham River carried bed load for only a small part of its flow season: for a few hours each day during the snowmelt flood, and at high stage during rainstorm floods. A crude application of two different bed load equations implies a low total for seasonal bed material transport of a few tonnes per square kilometre. This implication is in keeping with the coarseness of the stream's bed material and the low energy slope of its channel.

The suspended load carried during 1971 was heavier, amounting to 12.7 t km^{-2}. While this single number is little to go on, there is reason to believe that the Mecham River is geomorphically representative of other parts of the High Arctic, and (cf. sec. 2.6) that 1971 was, at least climatically and hydrologically, an average year for the region.

Measurements made in 1973 on the "Sverdrup" and "Schei" Rivers in south central Ellesmere Island provide information from an environment contrasting with that of the Mecham basin in several respects. Both of the streams have glacierized basins, but the glacier is a source of very little sediment; glacial meltwater, however, runs over a proglacial

terrain in which there is much unconsolidated fine sediment, and suspended loads are therefore much higher than in the Mecham River.

Comparisons of the Mecham River with streams farther afield show that it differs from Baffin Island streams in that its dissolved load is absolutely and proportionally much larger; the Baffin Island streams flow over preCambrian igneous rocks which are poor in solutes. A regression of stream load against basin relief for 20 streams in other parts of Canada gives moderately good results, and shows that the load of the Mecham River is similar to those, for example, of streams in the Maritime Provinces, but much smaller than those of most of the western Canadian streams. The Langbein-Schumm Rule, that basin sediment yield is greatest in semi-arid regions, fails for the Mecham River, which carried much less sediment in 1971 than the rule would dictate. It seems that the failure of the rule can be attributed to the coarser-than-average sediment supply of the Mecham basin, and also to the relative unimportance of raindrop impacts in dislodging material from barren, semi-arid surfaces in the Arctic. While freeze-thaw activity should compensate for this effect to some extent, it is probable that the difference in rainfall intensities is a significant

distinction between high- and low-latitude geomorphic regimes.

Most of the sediment removed by the Mecham River from its basin in 1971 was carried in solution: 21.1 t km^{-2} of a total of 35 t km^{-2}, most of the 21.1 t km^{-2} occurring as the species Ca^{++}, Mg^{++} and HCO_3^-. For much of the flow season suspended sediment concentration was small or nil, but dissolved solids concentrations increased as discharge decreased. 88 per cent of total suspended load was carried in 15 per cent of the time at discharges greater than 10 m^3 s^{-1}, while only 46 per cent of the dissolved load was carried in the same time.

A comparison of Mecham River solute data with data from studies of temperate and tropical karst shows that solute concentration increases with temperature. Mecham River waters contain less dissolved carbonate material than warmer waters from lower latitudes, thus helping to dispel the notion of Corbel (1959) that the converse relationship holds.

The factor which confuses the reasoning of Corbel is his neglect of organic CO_2. Waters enriched in organic CO_2 dissolve much more calcite than meteoric water with equilibrium P_{CO_2} close to the

global atmospheric mean. Almost all samples from southeast Cornwallis Island have P_{CO_2} close to or below the global mean. It seems probable that the low-P_{CO_2} waters derive from a snowpack from which CO_2 has been lost during compaction; those with average P_{CO_2} are average waters for the terrain, which, since it is barren, provides almost no biogenic CO_2; waters with above-average P_{CO_2} (and more dissolved Ca^{++}, allowing for the inverse relationship with discharge) are found later in the season as photosynthetic activity increases modestly, or in rills flowing through small patches of vegetation.

The inverse dependence of limestone solution on temperature is thus more accurately described as a dependence on production of CO_2, and more generally as a dependence on biomass. Even in the cold climate of the High Arctic, high solute concentrations may be expected in vegetated areas such as those found on calcareous till or alluvium. There are indications (Woo, private communication) that this is indeed so, and there are indications also that solute concentrations are "anomalously" high in the cool boreal forest waters of the Nahanni region of the Mackenzie Mountains (Brooke, Cowell and Ford, 1975). It has

still to be demonstrated that the solute-temperature relationship fails also in hot karst regions which are biological deserts.

SECTION 5

SUMMARY

"Why, why, why! Weh, O weh! I'se so silly
to be flowing but I no canm stay!"

James Joyce, Finnegan's Wake

5.1 Results

It will be well to begin this ending by stating again the main findings of the study. In the three preceding sections the study was subdivided into discussion of the Mecham River's runoff regime, its water balance and its role as an agent of denudation. This subdivision is maintained for purposes of summary.

5.1.1 The Arctic Nival Regime The Mecham River flows only during summer, its flow season lasting probably for 60-90 days. Flow begins in June or early July. The observed or estimated dates of first flow range from about 10 June to 4 July. Annual maximum discharge follows 13 or 14 days after the start of flow, on dates ranging from 24 June to 17 July during the period of study. Annual maxima have ranged from 30 to 60 $m^3 s^{-1}$,

or in round figures 0.3-0.6 $m^3 s^{-1} km^{-2}$ (27-55 $ft^3 s^{-1} mi^{-2}$). The period of high snowmelt discharges is short; it lasts for 10 to 15 days and the recession from the snowmelt peak rapdily leads to low discharges which prevail through the rest of summer except after rainstorms.

Stream responses to rainfall are rapid and shortlived, but antecedent moisture conditions are important in determining the magnitudes of responses to given amounts of rain. A flood in August 1973 began with a 20-fold increase in discharge (from 1.4 to 28.7 $m^3 s^{-1}$) in 4 hours; this flood had a recession constant of 14.1 hours, which is similar to recession constants calculated for other Mecham River floods and is short in comparison with those found in more vegetated catchments further south. Both smaller responses to larger rainfalls and (proportionally) larger responses to smaller rainfalls have been observed, but the pattern of rapid rise and rapid recession has recurred in each flood. The example quoted illustrates also the possibility of a rainfall response larger than the spring snowmelt maximum. 1973 had the driest winter on record at Resolute; the dry winter led to a subdued spring flood, which was followed in the summer by a pronounced response to a moderately heavy rainstorm.

The lag time of the basin response to this rainstorm was about 9 hours.

Similar lags were observed for the daily response of the basin to snowmelt generated by radiative fluxes of energy entering the snowpack. The average lag for 21 days in 1971, for example, was 8.0 hours, although individual time differences between radiation and discharge maxima ranged from 5 to 11 hours. Diurnal fluctuations tend to be less pronounced late in the season, when little snow is left to be melted and rainstorm flood recessions drown the small contribution of meltwater. While they occur, however, these diurnal fluctuations often involve a doubling of discharge in a few hours: on the day of highest flow in 1974, discharge increased almost 7 times in 8 hours.

Simulating and predicting the spring snowmelt flood are tasks which have still to be tackled successfully in the High Arctic. Although the energy transfers which determine when snow begins to melt are well understood in theory, it is more difficult to work them out in practice, and a full energy balance solution is not possible for snowmelt in the basin of the Mecham River. At the minimum, it is necessary to know snow temperature, but snow temperature data are not collected routinely at Resolute.

With the available data, all that has been achieved is an approximately true rule-of-thumb for dating the start of flow: this usually happens 4-6 days after the first day with mean air temperature above freezing. If the date on which flow begins is taken as known, more success can be claimed in modelling the subsequent course of the snowmelt flood, although the results are still only moderately good.

The best results are obtained with a regression model which simulates the energy balance by using its several components as independent variables. The poorest aspect of the model is its underestimation of the highest peak discharges, and improvements might be made by disaggregating the basin into spatial sub-units, which would contribute water with successively longer lags to the basin outlet.

In the 5-year discharge record for the Mecham River the most striking difference between annual hydrographs is in their timing. Although peak magnitudes and other properties of the hydrograph also vary from year to year, an examination of the longer weather record shows that the five runoff years included two years with average precipitation, the two driest years on record and the second wettest. No trends or

persistence can be detected in the weather record, and the variance of annual precipitation about the mean is moderate.

The extremeness of single events is also moderate. No daily rainfall exceeding one inch (25.4 mm) has yet been recorded, and the 500-year daily rainfall at Resolute is only 45.2 mm, although heavier rains than this have fallen at other High Arctic stations.

5.1.2 The Annual Water Balance The representativeness of the study period is encouraging when the results of water balance analyses are considered. In the High Arctic, unglacierized basins may have water balances which are much affected by terms other than those of the classical continuity equation. There are no changes in deep groundwater storage, but active layer storage or permafrost watermelt may possibly be substantial in certain circumstances. Over a number of years, however, changes in the moisture content of the active layer must sum to zero. Other forms of storage, in lakes and in semi-permanent snowpatches, may be neglected in the Mecham River basin.

Windblown snow has sometimes been proposed as a substantial element of the water balance, but it too

may be neglected in the Mecham River basin, and probably in other basins of its size. A snow transfer model of the terrain in and around the basin shows, (if it is true that snow is not blown over more than a few kilometres each winter), that the amount of snow blowing across the drainage divide must be extremely small in comparison with the amount falling on the basin.

Of the remaining terms in the water balance, snowfall, rainfall and discharge are all measured for the Mecham River, while it is necessary to calculate evaporation. The combination model can be applied successfully to the routinely available data for calculating the vapour flux, although, as with the problem of modelling the snowmelt foood and events leading to it, improvements would be desirable. Such improvements are, again, provision for measuring snow temperature (and in particular snow surface temperature), and also provision for measuring or estimating surface wetness. Not knowing how moist the surface is (when it is free of snow) entails not knowing with enough confidence the vapour concentration gradient from surface to atmosphere, and this is probably the major weakness in the version of the combination model developed here. Lesser weaknesses are in knowledge of the wind speed gradient, and of per cent basin snow cover.

In spite of its weaknesses, however, the performance of the model seems reasonable. On the two occasions when a check is possible, calculated monthly evaporation is, as it ought to be, less than measured pan evaporation. Moreover, encouragement can be taken from a calculation of the empirical coefficient of Priestley and Taylor (1972). These workers suggest that in a wide variety of situations this coefficient is equal to 1.26 on average, but at Resolute for monthly evaporation during the light season it is equal to 1.47 ± 0.37.

This high value, it can be argued, occurs because of the windiness of Resolute. Priestley and Taylor's coefficient has already been shown to vary with temperature and surface wetness, and it is reasonable to expect that it should also increase with wind speed. The second term in the combination model is a function of these three variables, and Priestley and Taylor's coefficient replaces this second term.

Although the results must be considered approximate, it is possible to solve the energy balance at Resolute using measured net radiation, calculated latent and soil heat fluxes, and sensible heat flux calculated as a residual. In a sample year the annual

Bowen ratio was 0.70, which was very close to the monthly ratio for July, the month when fluxes are largest. In the same year, about 20 per cent of the energy available for turbulent transfer (i.e., of latent and sensible heat) came from the ground, and indeed the annual soil heat flux was directed upwards in each of the years studied.

In the last of the study years it was necessary to use published monthly means instead of twice-daily data in the evaporation calculations. A check on both methods of calculation indicates that very little extra uncertainty is introduced with the published means, and the model may therefore be a relatively cheap tool for calculation of the vapour flux.

The fluxes calculated for water balance purposes in this study proved to be unexpectedly large. In the four water years evaporation exceeded discharge in three, and in the fourth was about equal to discharge.

The major result of the water balance analysis can be expressed in the statement that measured and calculated outputs of water from the Mecham River basin are much greater than measured inputs; there is good reason to believe that the discrepancy is due to underestimation of the inputs. If the sum of evaporation and discharge is assumed to represent true precipitation,

the underestimate due to rain and snow gauge deficiencies is on average 400 per cent. That is to say, average true precipitation at Resolute is approximately 530 mm (21"), while average measured precipitation is only 136 mm (5.36").

Checks on the accuracy of this statement are hard to find, but if discharge is assumed accurate to ± 10 per cent and the minimum evaporation is assumed to be given by the first term of the combination model, actual still exceeds measured precipitation by 300 per cent. If maximum evaporation is assumed to be given by setting the surface wet-bulb depression to zero for all calculations, the underestimate increases to 500 per cent.

Unrecorded light or "trace" precipitation can be discounted as an important source of the missing input water. A small numerical allowance for each 6-hourly report of trace precipitation adds up to an annual total which, while it is typically of the same order as the annual rainfall, is still only about 10 per cent of the amount of water leaving the basin.

For hydrologic purposes, the simplest way around the problem of precipitation underestimation is probably to collect extensive information on snowpack

water equivalent at the end of winter. Indeed, circumstantial support for the major finding of the water balance study comes from a survey of this sort made in June 1972. The survey (on a 20-stake snow course) showed a snowpack water equivalent four times that of the snowfall measured at the Resolute weather station in the preceding winter.

5.1.3 The Geomorphic Role of Running Water Bed load is always difficult to measure, and especially so in coarse-bed streams such as the Mecham River. However, it is known that during the year of closest observation (1971) the bed material of the Mecham River was in motion for only a small part of the total flow season: at the discharge maxima of the spring flood, and for a few hours during a response to a rainstorm. A crude application of two empirical bed-load formulae suggests that the total bed load in 1971 was proportionally much less than the suspended and dissolved loads. The two formulae gave figures of 1 and 4 tonnes per square kilometre for the seasonal bed load.

Suspended load in the same year was 12.7 t km^{-2}. Most of this was carried during the spring flood, when suspended sediment concentrations of several hundred

p.p.m. (and, very briefly, of over 1300 p.p.m.) were measured. Concentrations were very small or nil through most of the remainder of the season when discharge was low.

The Mecham River flows in a gently-sloping channel, through terrain of moderate relief in which the supply of fine sediment is not abundant. Its suspended load is comparable with that of streams answering to similar descriptions in other parts of Canada. A regression of suspended load against basin relief shows that the Mecham River carries much less sediment than streams in western Canada which have high relief. Streams in the Maritimes with similar relief have similar suspended loads.

The suspended loads of two streams in south-central Ellesmere Island were measured in 1973, and in comparison with the data from the Mecham River this additional information illustrates the significance of sediment supply in determining stream load. The two streams both have glacierized basins, and although the glacier ice contributes very little sediment they both flow through proglacial terrain in which there is much fine sediment. Their basins also have greater relief than that of the Mecham River, but their

higher suspended loads - 340 and 230 t km^{-2} - may be interpreted as being due mainly to the presence of glacier ice as a source of runoff and to the abundant supply of fine sediment.

A wider comparison for the Mecham River data can be found in the Langbein-Schumm Rule (Langbein and Schumm, 1958), which says that suspended load is greatest in semi-arid regions, decreasing to moderate amounts in wetter (sc. forested) areas and to zero in the drier, arid deserts. Given the findings of sec. 5.1.2 it is hard to say whether the Mecham basin is arid or semi-arid, but in any case its yield of suspended sediment is much lower than predicted by the rule. It is possible that this is because the rule was deduced from data on streams in warm semi-arid regions. In such regions the importance for sediment yield of raindrop impact on exposed loose sediment is undoubtedly much greater than in the High Arctic. As shown in sec. 5.1.1, extreme rainfalls in the High Arctic are light by temperate-latitude standards, and the number of these extreme events is also small.

Most of the Mecham River's sediment load in 1971 was made up of dissolved solids. The solute load was 21.1 t km^{-2}, which leads to an estimate for

the total load of 35 t km^{-2} (or 13 millimetres per thousand years). This result is consistent with the knowledge that the Mecham basin is in limestone and dolomite bedrock, and as might be expected the great bulk of the dissolved load occurred as the species Ca, Mg and HCO_3.

Solute concentrations, unlike suspended sediment concentration, varied inversely with discharge, and the removal of solutes was spread more evenly through the flow season. The mean ratio of Ca to Mg, 3·56, was relatively high, which is to be attributed partly to the lesser solubility of dolomite than of calcite at low temperatures.

The thermal control exerted on solution is such as to reduce solution at lower temperatures. This is illustrated by a comparison of the Mecham River's load with that of streams in the U.S.A. The variability of dissolved loads is less than that of suspended loads, but still the Mecham River carries less material in solution for the same amount of runoff than do other limestone streams in areas such as Pennsylvania.

The reduced solutional work done by colder waters is, however, better explored through data on

solute concentrations. Waters in and around the Mecham basin have concentrations lower than in most limestone regions in warmer climates, the range in the Mecham River itself, for example, being 0.24-0.69 mol m^{-3}. The average Ca concentration of Mecham River samples is very close to that calculated for spring waters above the treeline in the Rocky Mountains of Alberta. The same average plots very close to a regression line of Ca against water temperature recently presented by Harmon et al. (1975), for which the raw material was information from limestone regions in Mexico, the U.S.A. and Alberta.

The dependence of solution on temperature is an indirect one. There is more solution in lower-latitude regions because their waters have a higher partial pressure of CO_2 and are hence more aggressive with respect to calcite. The higher P_{CO_2} values occur because of biogenic enrichment as meteoric water passes through vegetation canopies and soils. In the Mecham River basin, and other cold limestone regions, there is almost no organic material from which runoff water is able to obtain CO_2, and average P_{CO_2} is almost the same as the global atmospheric mean. Saturation with respect to calcite is observed to occur at concentrations

very close to those predicted theoretically for a system with the global mean P_{CO_2}.

The proportionally small number of samples from the Mecham basin which do have higher P_{CO_2} also have higher Ca concentrations. This shows that at the regional scale the Ca-temperature relationship is meaningful mainly in so far as it reflects a coincidence of relationships between Ca concentration and biomass and temperature and biomass. Data from cold regions with high biomass appear to depart from the Ca-temperature relationship, and it is possible that a departure in the opposite direction may occur in hot regions with low biomass.

One group of Mecham River samples shows not just an absence of biogenic CO_2 added to the atmospheric norm, but a considerable depletion below that norm, some samples having a P_{CO_2} only one tenth of global mean P_{CO_2}. These are waters measured during the snow-melt flood, and it is believed that their unusual chemistry can be ascribed to their origin in the winter snowpack. Expulsion of CO_2 may occur during compaction of the snow, which when it melts may produce meltwaters with reduced P_{CO_2}.

5.2 Reflections

The picture drawn in these pages is of a High Arctic environment in which water has a more prominent place than in the conventional view. Climatically arid by most standards of classification, the Mecham River basin is by no means the dry and frosty place which it might be thought without first-hand acquaintance. Anyone who has walked its slopes in summer could testify as much on the evidence of muddy boots, but the achievement of this study has been to document numerically such circumstantial impressions, to offer reasons why this desert should be so suspiciously muddy and to give some geomorphic measure, figuratively speaking, of its muddiness.

The Mecham basin and its locality are not arid because in the first place much more precipitation falls there than is measured. In the second place evaporation, although substantial, is proportionally much less important than in the low-latitude deserts. Water, therefore, is always found in considerable quantities, whether frozen in winter or liquid in summer, and if the Mecham basin is indeed a physiological desert this is partly because it is cold, mainly because it is calcareous, and not at all because it is arid.

Given this abundance of water in the High Arctic environment, some social implications can readily be foreseen. High Arctic society has coped up to the present without much management of its water resources, but these resources will require careful attention in the near future. The peculiarity of the arctic nival regime is that most of the year's runoff occurs in a few weeks, and the problem of predicting socially hazardous floods becomes largely a problem in predicting exceptionally snowy winters. Damage to pipelines and other structures is more likely to result from high snowmelt discharges than from high rainfall response. If really large settlements are ever established in the High Arctic they will require special provisions for their water supply. But the results of this study suggest a larger water supply than do official measurements of precipitation, and impounding and storing enough water for a large settlement may then be a correspondingly less difficult problem.

A wider and perhaps more worrying social implication depends for its significance on whether precipitation is underestimated as much at other stations as at Resolute. Some writers have been concerned that the Arctic is a sort of global dump for the atmospheric contaminants of world society, and

particularly of the mid-latitude societies of the Northern Hemisphere. Pollutants put into the atmosphere in these middle northern latitudes are thought to be carried by westerly winds into higher latitudes, where they fall out and accumulate on and near the surface. If this hazard is a serious one (which is not yet known), how much more serious would it become if it were learned that 400-per cent underestimates of precipitation were widely typical of meteorological efforts in the Arctic?

The wide view of the Arctic as a geomorphic environment seems also to suffer from underestimates of water, although in this case the underestimates of amounts are compounded by underestimates of their effectiveness. The main geomorphic argument of this thesis was stated quite simply in the introduction to be that running water must do geomorphic work whatever the circumstances which surround its running. If water runs off High Arctic terrain, it must carry some sediment away with it. The data collected to reinforce the argument show, not so much that the Mecham basin is the scene of vigorous and hitherto unsuspected fluvial activity, but that it is geomorphically and in particular fluvially a rather quiet place with a low-energy

regime. There are details to set this fluvial regime apart from others, such as the regular annual recurrence of the snowmelt flood to modify the more familiar magnitude-frequency relationship for geomorphic events.

Some of these details tend indeed to show that the Mecham basin suffers a reduction in the intensity of geomorphic processes because of its cold climate. Although limestone solution is the most important process of denudation in the basin, it proceeds less rapidly than in lower-latitude limestone regions because the Mecham and its locality are cold. Here is a genuine way in which the periglacial environment sees less fluvial activity than temperate-latitude and other environments.

But the point which has been demonstrated is precisely this: there are good reasons why the Mecham basin is fluvially a low-energy environment, and these reasons have little to do with the coincidence in the basin of fluvial and cryogenic events. Cryogenic processes are interesting and challenging peculiarities, but the geomorphic laws which apply elsewhere apply here also in the periglacial environment. The extrapolation of a law saying that there is less solution

at lower temperature can be matched with the extrapolation of a law saying that more suspended sediment is transported in steep channels with fine-grained beds. Both extrapolations have equal validity in the High Arctic, and whatever the significance of periglacial processes in the region its streams continue to flow and to do geomorphic work as do all streams everywhere.

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