WATER BALANCE OF A SMALL HIGH ARCTIC BASIN

By

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A Thesis

Submitted to the Faculty of Graduate Studies

in Partial Fulfilment of the Requirements

for the Degree

Master of Science

McMaster University

January, 1978

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Frontispiece: Aerial view of the study area (outlined in black), 5 km north east of Resolute meteorological station. Photography was taken in early August 1976, with late lying snow occupying the valleys.

MASTER OF SCIENCE (1978)
(Geography)

MCMASTER UNIVERSITY
Hamilton, Ontario

TITLE: Water Balance of a Small High Arctic Basin

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NUMBER OF PAGES: viii, 108

ACKNOWLEDGEMENTS

I am greatly indebted to my supervisor, Dr. M.K. Woo for his friendship, encouragement and help during the field research and data analysis stages of this research project. Thanks to Dr. W. Rouse for his critical reading of various portions of this thesis and for other valuable discussions.

The field work was supported by a National Research Council of Canada grant and the Presidential Committee on Northern Studies of McMaster University. The generous logistical support of Polar Continental Shelf Project, Department of Energy, Mines and Resources is gratefully acknowledged.

Much thanks must go to my many friends in the Geography Department and at McKay Hall who made my stay in Hamilton so very enjoyable. Special thanks to Fred, Bill, Emile, George and Lief who made my stay at P.C.S.P. in Resolute so enjoyable. Thanks to Ed Zwick for his friendship in the field.

I would also like to thank Dr. S.B. McCann for initiating my field work in the Arctic.

TABLE OF CONTENTS

LIST OF FIGURES

LIST OF TABLES

ABSTRACT

A small drainage basin (area 33 km^2) near Resolute Bay, N.W.T., and three sub-basins with areas of 21, 10 and 0.65 km^2 were studied during the summer of 1976. All components of the water balance were measured or calculated, including basin snow storage, rainfall, streamflow and evaporation. A snow survey indicated that Atmospheric Environment Service data underestimated basin snow storage by 50% but weather station rainfall was representative of the study basin. Water balance studies showed that for the three larger basins, streamflow consistently accounted for 80% of the total incoming precipitation and evaporation 20%. The smallest basin was found to discharge only 67% of precipitation, leaving 33% available for evaporation. This difference was attributed to a larger percentage of wet areas in the small basin. The spatial variability of the water balance components was also demonstrated by calculating the values of all components for individual terrain units. It was found that the percentage of total precipitation which was discharged as streamflow varied from 53% for crests to 98% for gullies. Findings from this study agree with other studies in a similar environment. In general, streamflow as a component of the water balance is more important in the high Arctic than in the sub-Arctic regions.

CHAPTER 1

INTRODUCTION

1.1 Introduction

Previous hydrologic studies in the Canadian high Arctic have concentrated on streamflow (Church 1971, Cogley 1971, Cogley 1975, Ballantyne 1975, McCann, Howarth and Cogley 1972), mainly in terms of the streamflow regime and the volume of run-off. Few studies emphasized the other components of the water balance such as evaporation or precipitation. For example, it is well known that precipitation in the Arctic is underestimated and that fairly large regional variations exist (Cogley 1975, Hare and Hay 1974, Walker and Lake 1973), but there has been no attempt to determine the exact amount of this underestimation. Another unknown component of the northern hydrologic system is evaporation. Detailed work has been done in the subarctic (Rouse, Mills and Stewart 1978), but little is known about the importance of evaporation in the high Arctic (Hare and Hay 1971).

Another neglected aspect of Arctic hydrology is the nature of the hydrologic system which controls the transfer between the input and the output components of the system. This includes such problems as the effect of channel opening processes during the melt period.

Future development in the high Arctic will require better knowledge of the availability of water in the northern hydrologic system. It is thus necessary to survey our existing body of information, then to advance our knowledge in Arctic hydrology by furthering research efforts in the areas where our understanding is deficient.

1.2 Literature Review

The water balance of northern river basins is composed of the following major components: snowfall, rainfall, streamflow and evaporation. Of these components, streamflow is the most studied and the best understood (Ambler 1974, Carlson 1974, Cogley 1975, Church 1974, Cook 1967, Dingman 1973, Kane and Carlson 1973, Mackay and Loken 1974, Pissart 1967). In general, rivers in the high Arctic follow a basic annual cycle. During early June snowmelt begins, but most of the snowmelt is accomplished within a one week period in mid to late June. Much of this melt water does not leave the basin until the snow-choked channels are flushed by a catastrophic flood during which 90% (Cook 1967) of the mean annual flow leaves the basin. After the flood, flow declines, approaching zero by early to mid-September. There is no flow until the next June. In detail, the nature of the channel opening processes is qualitatively described by Pissart (1967). Streamflow during the melt period is characterized by diurnal variations depending on the daily variations in radiation and thermal conditions (McCann and Cogley 1971).

The major characteristics of streams flowing in permafrost areas are summarized by Church (1974). He notes that the dominant role of permafrost is to keep the water near the surface. As a result, surface runoff is usually relatively rapid but recession flows are highly variable. In rolling, non-vegetated terrain recession flows are normally short, while in basins dominated by tundra, heath or muskeg, they may be very long. Newbury (1974) also demonstrated that in permafrost areas, there is an increase in the proportion of precipitation available to runoff.

In the Canadian high Arctic, the amount of streamflow data are sparse and there are no long term data to allow proper statistical analysis. Although data collection programs are being expanded by the Water Survey of Canada, resource development poses an urgent need for hydrologic information. Therefore better understanding of the northern hydrologic system is required to enable streamflow modelling and hence a prediction of long term changes in the hydrologic system.

In terms of precipitation, many studies have noted a gross underestimation by official weather stations (Cogley 1975, Church 1974, Cook 1960, Dingman 1973, Findlay 1966, Hare and Hay 1971, Wedel 1977). Findlay (1966) found that 37% of the actual snowfall was not caught, while the respective figure for rainfall was 16%. Cook (1960) considered that trace rainfall events at Resolute can add substantially to the measured total. While Dingman (1973) noted that condensation and snowfall were major problems, Hare and Hay (1971) believed that the underestimation of snowfall was the most serious problem. Despite the general belief that precipitation in the far north is severely underestimated (Church 1974), very few studies have attempted to show the actual value of the underestimation.

Using both lysimeters and microclimatic methods, the importance of evaporation in the subarctic regions has been established (Church 1974, Dingman 1973, Ferguson et al 1970, Findlay 1966, Findlay 1969, Nebiker 1957, Nebiker and Orvig 1958, Watts et al 1960). The few evaporation studies carried out in the high Arctic are site specific and do not provide basin wide data to allow comparison with other components of the water balance (Addison 1972, Addison 1975, Smith 1976, Weller and Holmgren 1974). Addison's studies on Devon and King Christian Islands showed that evaporation rates were highly variable over short distances and that evaporation was extremely small over certain dry surfaces. But no data on the proportion of the basin which was covered by each surface type were given and therefore basin evaporation cannot be estimated.

This review shows that most of the previous studies have concentrated on individual components of the water balance rather than considering them as parts of a northern hydrologic system. In recent years, there are a number of studies on entire drainage basins (such as Cogley 1975, Findlay 1966, Holocek and Vosahlo 1975, Wedel 1977).

In view of the vastness of the Canadian high Arctic and our lack of information on these areas, water balance of Arctic basins will play a role in contributing to our knowledge of permafrost hydrology.

1.3 Objectives

To improve our understanding of the hydrologic system of high Arctic regions, the research was carried out to study the water balance of a small basin in the high Arctic. More specifically the objectives are:

(1) to determine the total basin snow storage at the end of winter and determine the representativeness of the Resolute weather station snowfall data.

(2) to determine the total amount of summer precipitation and stream discharge.

(3) to obtain an accurate estimate of evaporation as the residual of the water balance equation.

(4) to assess the accuracy of calculating seasonal evaporation as a function of equilibrium evaporation, over the dominant surface types found in the research basin.

(5) to assess the spatial variation of the water balance components outlined above, both at the basin and for individual terrain units.

CHAPTER 2

STUDY AREA AND METHOD

2.1 Location

Field work was carried out in a basin (74° 43'N, 94° 59'W; area 33 km^2) approximately 5 km north of Resolute Airport, Northwest Territories (figs. 2.1 and 2.2). To study the spatial variability of the water balance components, this basin was further divided into three subbasins with areas of 21, 10 and 0.65 km^2. This study area was chosen because

(1) it was close to the weather station at Resolute, thus enabling a comparison of field data with standard weather station data,

(2) most parts of the basin were within easy reach from Resolute, thus enabling the collection of a large amount of data with limited manpower.

2.2 Topography and Terrain Units

Topographically the basin can be divided into four regions (fig. 2.3):

(1) Plateau: an area with general elevation of 120-190 m surrounding the eastern and southern portions of the basin (fig. 2.4).

(2) Rolling Terrain: a low-lying area (elevation of 75 to 120 m) with a relative relief of less than 10 m (fig. 2.4).

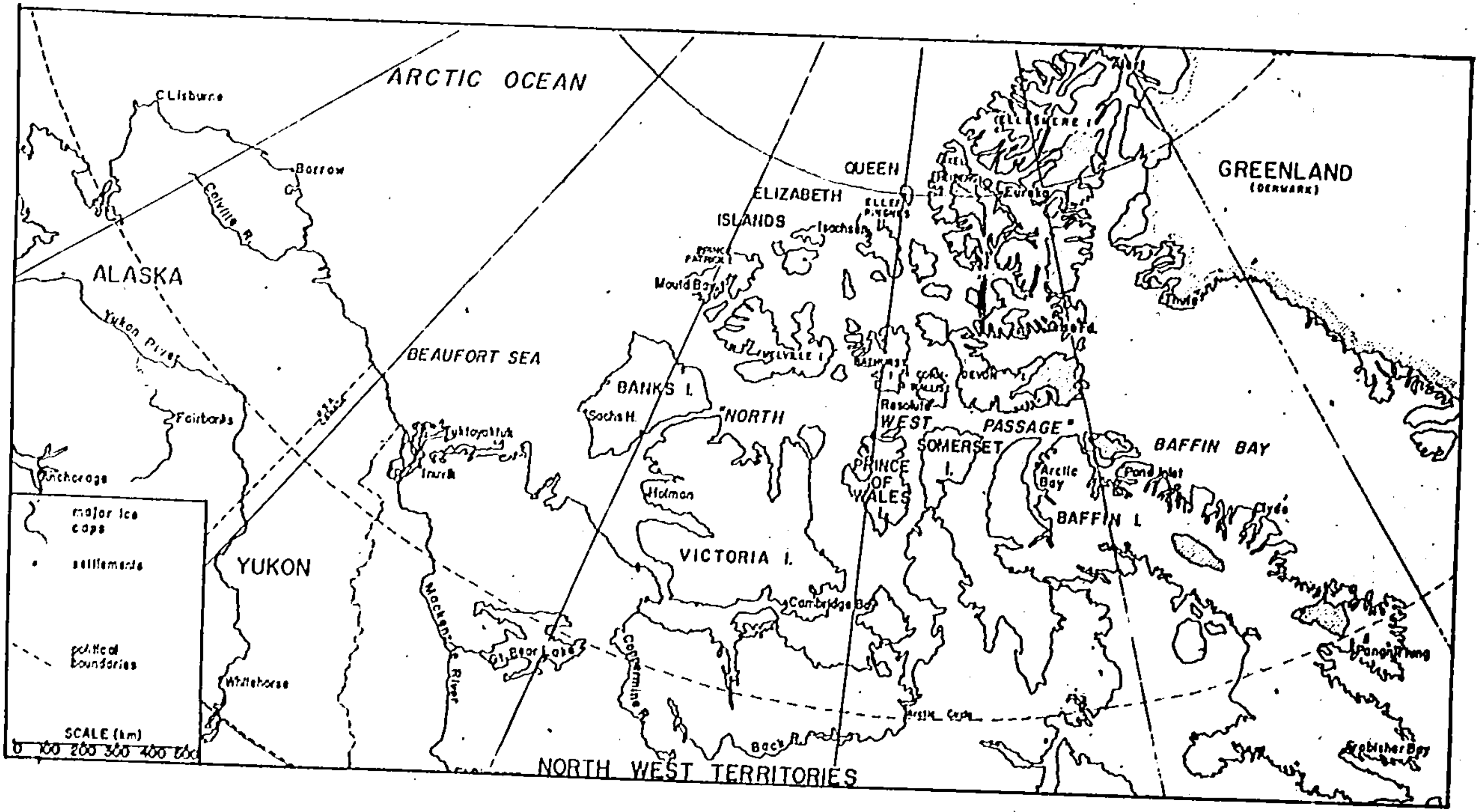


Fig. 2.1 The Canadian Arctic

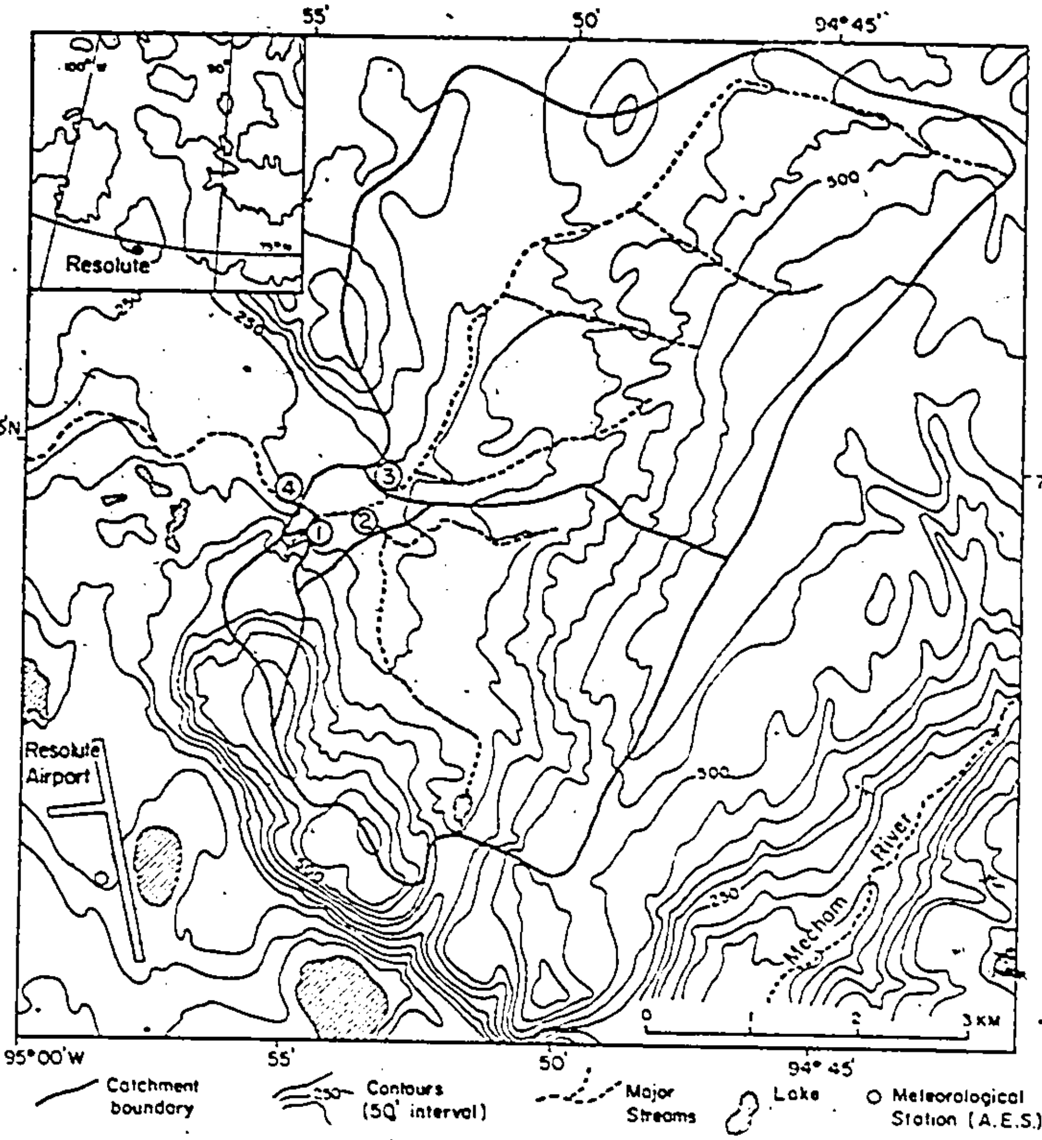


Fig. 2.2 Topography of the study area near Resolute, Cornwallis Island, N.W.T. The basin areas are: (1) 0.5 km^2, (2) 10 km^2, (3) 21 km^2, (4) 33 km^2.

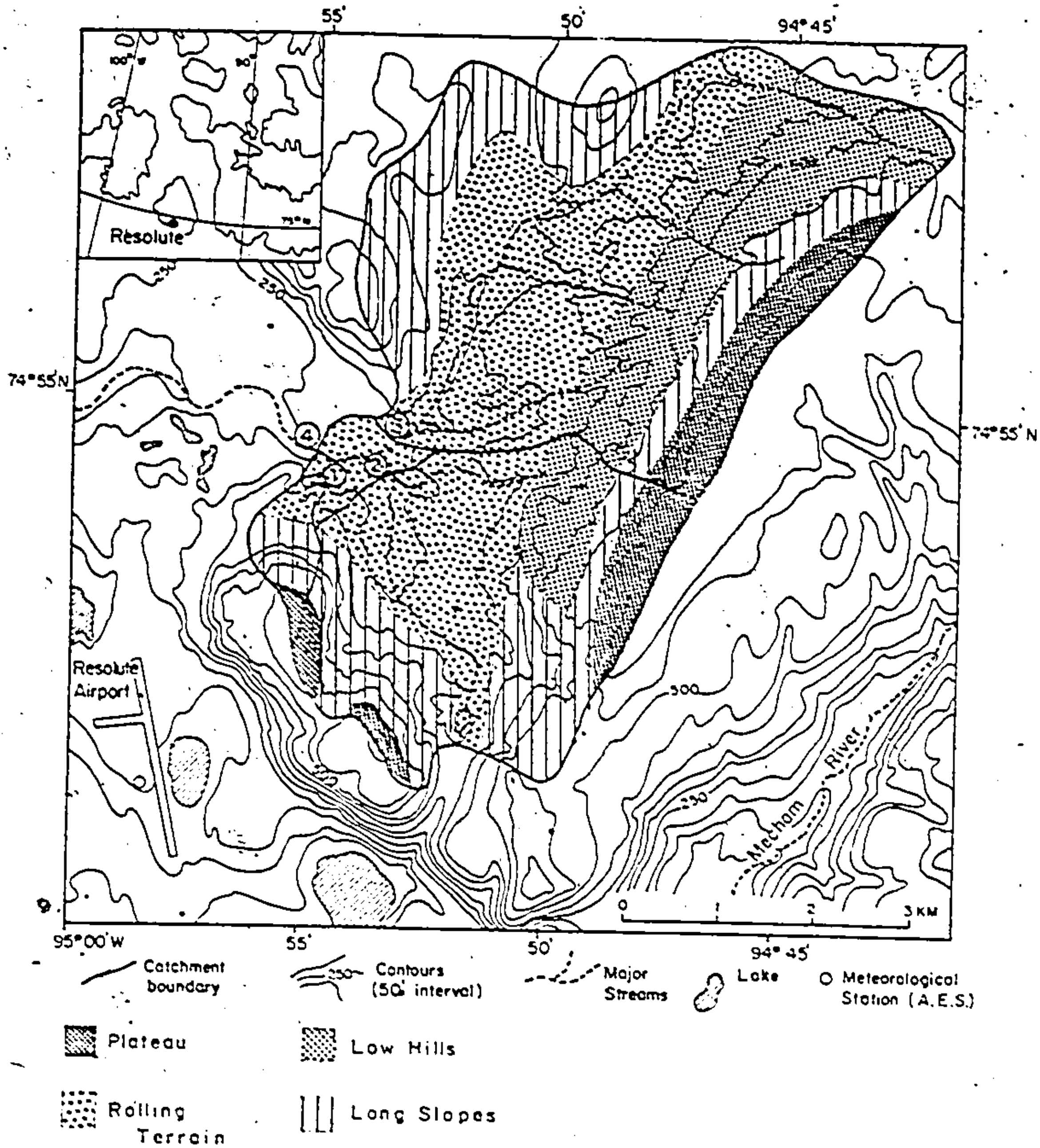


Fig. 2.3 The four major topographic units in the study basin

(a) Typical Plateau in Foreground

(b) Rolling Terrain

Fig. 2.4 Four topographic units found in the study basin

(c) Low hills

(d) Typical Long slope in background

Fig. 2.4(cont'd) Four topographic units found in study basin

(3) Low hills: an area between the eastern plateau and the rolling areas. This region is dissected by valleys up to 15 m deep (fig. 2.4).

(4) Long slopes: slopes extending from the plateau to the rolling areas but are not dissected by valleys (fig. 2.4).

The surface material is generally stoney and sandy with a sparse vegetation cover. Cruickshank (1971) described the following soil and terrain units found within the basin (fig. 2.5):

(1) bog soils: The material is a sandy loam which is continually wet and the surface is colonized by black lichens, mosses and vascular plants. They are located on flat and low-lying areas, or below semi-permanent snow banks (fig. 2.6).

(2) polar desert: It consists of small limestone chips in a sandy loam mantle with a negligible plant cover, and is generally found in elevated locations (fig. 2.6).

(3) lithosols: These consist of shattered limestone (fragments 0.1 to 0.25 m in diameter) occuring on level ground with a very sparse plant cover (fig. 2.6).

A thick layer of permafrost at Resolute (390± 3 m, Misener 1955) precludes the possibility of subpermafrost

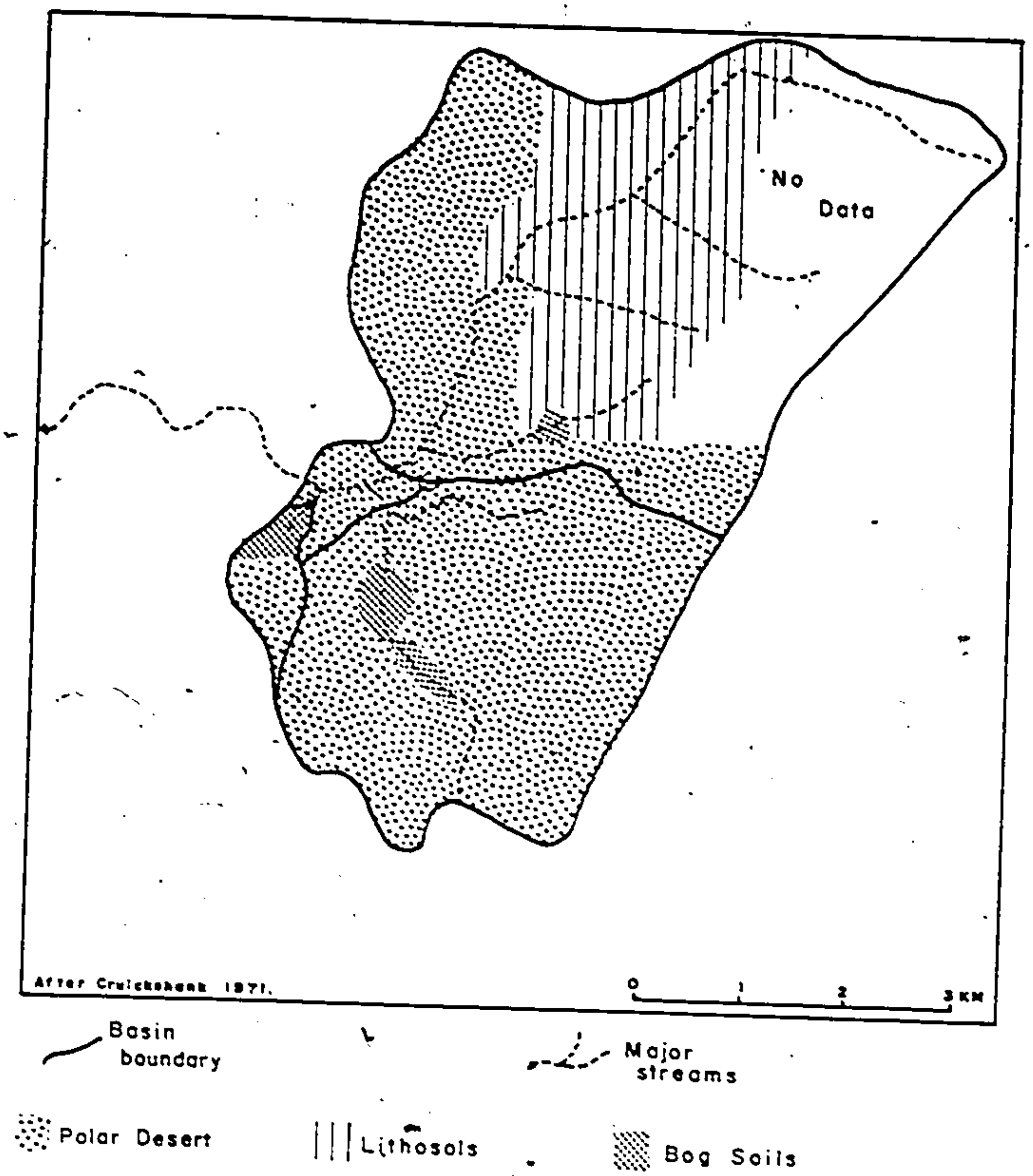
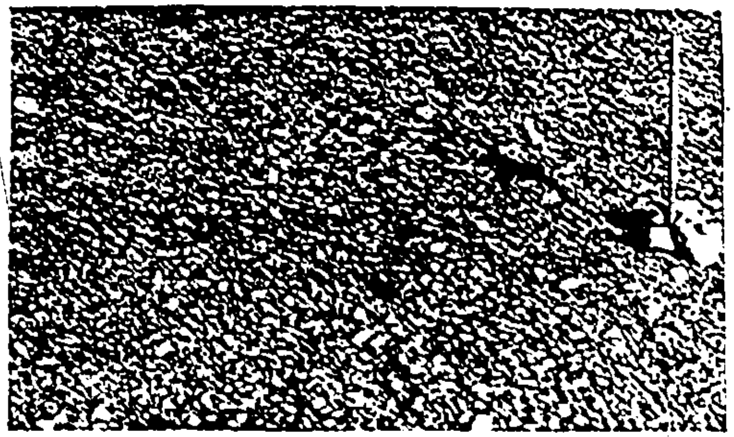


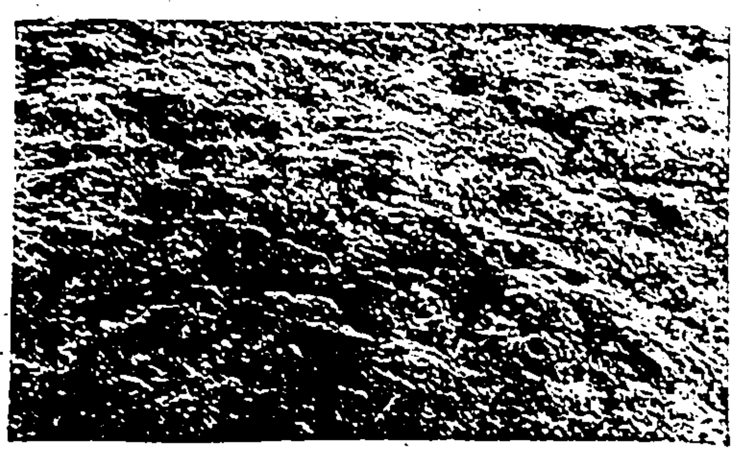
Fig. 2.5

Three terrain units, as defined by Cruickshank (1971), found in the Research Basin

(a) Polar Desert



(b) Lithosols



(c) Bog soils

Fig. 2.6 Basin Terrain Units defined by Cruickshank (1971) groundwater contribution to the surface hydrologic cycle. The occurrence of the permafrost table at a shallow depth (depth of frost table less than 1 m) ensures that most of the hydrologic activities remain close to the surface.

2.3 Climate

Climate exerts a strong influence on the hydrology of a basin. In the high Arctic, precipitation is low, but because of the low temperatures the snow is stored in the basin for approximately ten months each year. The mean daily temperature rises above 0°C only during late June, July and August when the precipitation of the entire year is available for surface runoff and evaporation.

An extensive review of the Arctic climate is given by Barry and Hare (1974) and the climatic conditions of Arctic Canada are discussed in Hare and Hay (1974). For Resolute, the climatic conditions can be summarized as follows (fig. 2.7).

(1) September to November: Temperature falls steadily, passing below the freezing point after mid-September. This is a period of relatively heavy snowfall, receiving 45% of the annual total.

(2) December to April: A significant change in the weather pattern occurs after November. Firstly the seas become frozen, cutting off the local moisture supply and secondly, the wave cyclones move further south. Mean daily air temperature reaches −28°C in February,

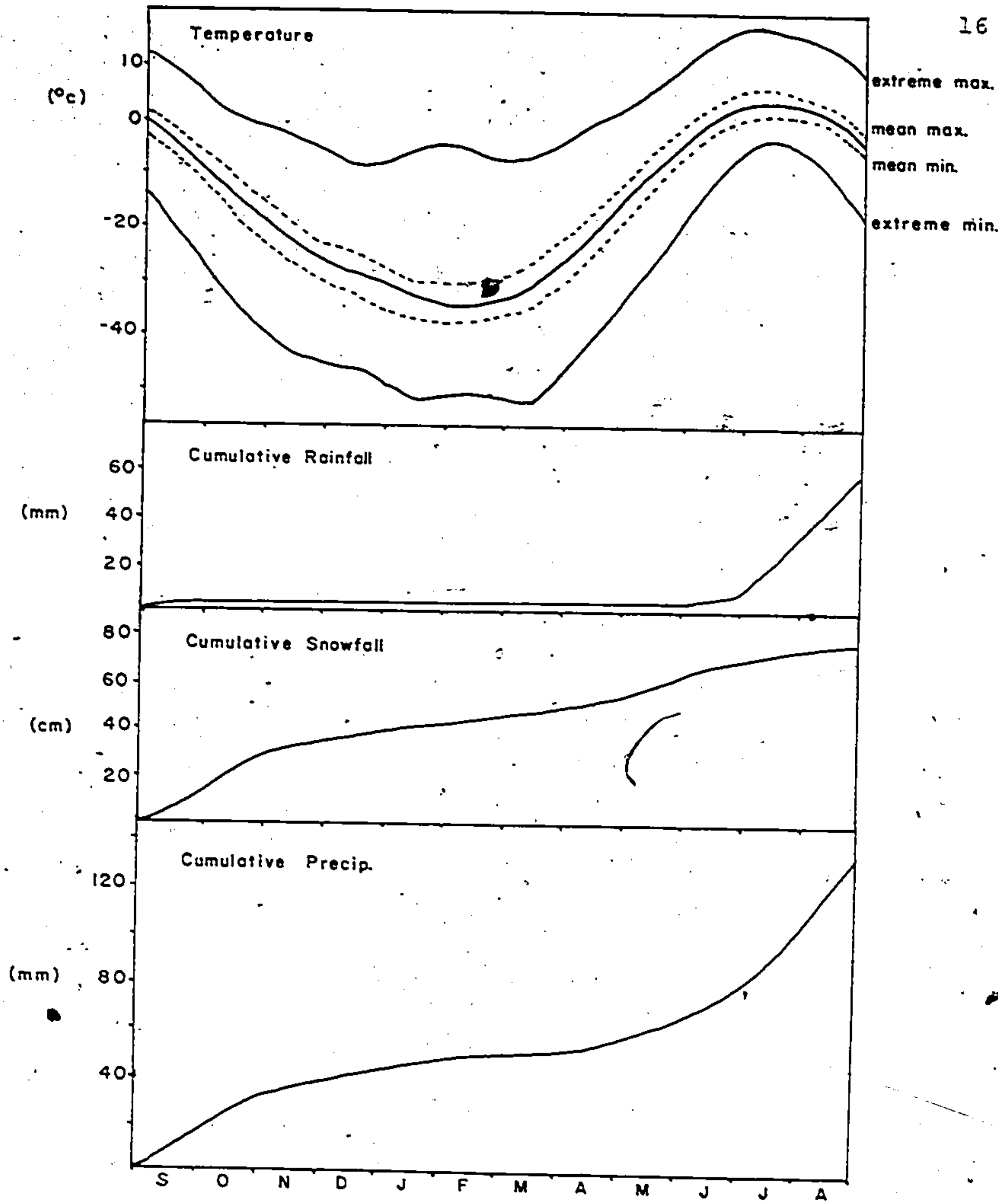


Fig. 2.7

Summary of Climatic Conditions at Resolute, N.W.T.

but the outstanding characteristic of winter temperature is the persistence rather than the extreme severity of the coldness. Snowfall during this period is light, but blowing snow is prevalent. Fraser (1964) found that for December, January and February, blowing snow constitutes 17, 24 and 16 percent of all the observations.

(3) March to June: Temperature climbs steadily throughout this period, but due to the high surface albedo of snow, net radiation remains low and little energy is available to heat the air. It is not until June 15 that mean daily temperature reaches 0°C. Precipitation increases gradually, all of which occurs as snowfall until late June when the first rainfall occurs.

(4) July and August: These are the only months when temperature is consistently above freezing with mean daily temperatures of 4.3 and 2.7°C. respectively. A combination of increased frontal activities and the opening of the sea ice results in an increase in precipitation (mostly rain). Fog is common, and it occurs during half of the entire period (Barry and Hare 1974). Cooling of the air begins in August and temperatures can dip below 0°C by the middle of the month. A snow cover can be established by mid-August but the average date is about September 1.

2.4 Data Collection

The data collection program was designed to allow all input and output components of the hydrologic system to be measured in the field or calculated from the field data. Most data were obtained from the research basin, but data from the Atmospheric Environment Service weather station were used as a supplement.

2.4.1 Meteorological Data

Air temperature was recorded by a Lambrecht thermograph housed in a Stevenson's screen whose location is shown in fig. 2.8. The accuracy of the thermograph was checked by a mercury thermometer. Net radiation data were obtained for three different surfaces. For the snow surface, a Swissteco net radiometer was used and the signals were recorded on a Rustrak recorder (fig. 2.8). The radiometer was kept desiccated and inflated by pumping air through a bottle of silica gel. Net radiation over polar desert surface was measured by the Atmospheric Environment Service (A.E.S.) during the month of August. To obtain data for June and July a regression relationship of the following form was used (fig 2.9):

$$Q^* = a + bk\downarrow \qquad (2.1)$$

where Q^* is net radiation and $k\downarrow$ is incoming solar radiation. The necessary $k\downarrow$ data were measured by A.E.S. for the entire season. Since the surface over which A.E.S. results were obtained was similar to most parts of the research basin, these data are considered representative of the basin. No radiation data were

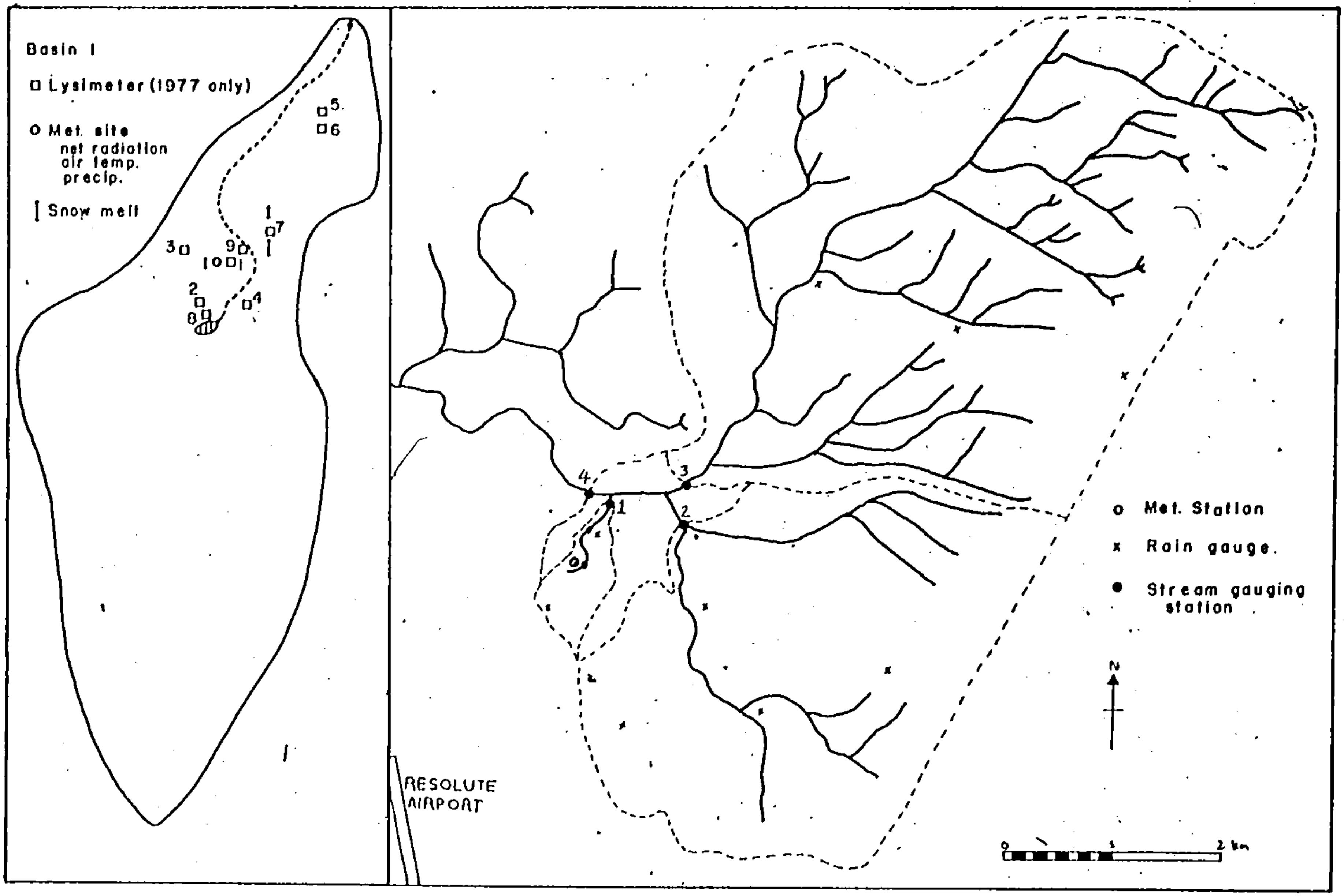


Fig. 2.8 Location of field instruments in the research basins

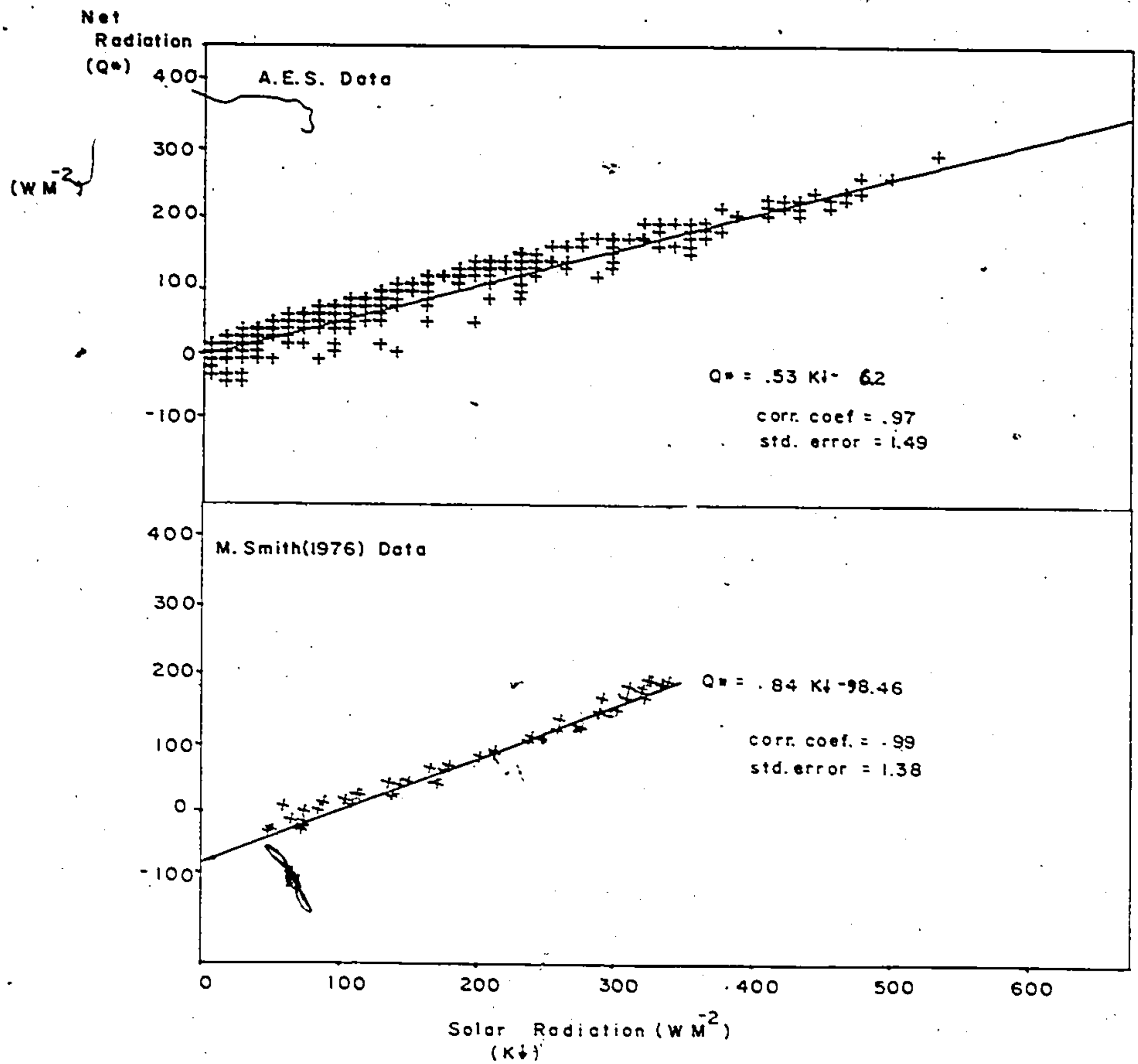


Fig. 2.9
Regression relationship between net radiation and solar radiation

available for boggy surfaces. However, Smith (1976) conducted microclimatic studies near Eureka, Ellesmere Island (80° 00'N, 85° 56'W), and there are similarities between the surfaces studied by Smith and the boggy surfaces of Resolute. Both sites were saturated, had similar vegetation cover and thermal properties and the surface albedo was 0.2 for the Eureka site and 0.18 for the Resolute site.

Smith's data were therefore used to obtain a regression relationship between $k\downarrow$ and Q^* (eq. 2.1) and the results are graphed in figure 2.9. This relationship then enables the computation of Q^* for the bog surfaces of Resolute for the 1976 field season.

Precipitation is measured by the government weather station at six hourly intervals. These data were available for the period September 1975 to September 1976. To check the representativeness of these data when applied to the research basin, additional data were collected.

(1) Ten non-recording rain gauges were deployed throughout the basin (fig. 2.8) to determine the spatial variability of rainfall. These gauges were measured at the end of each rainfall event.

(2) A snow survey was carried out to estimate total snow storage in the basin prior to the spring melt (see section 2.4.2).

Snowmelt was measured at three locations within the smallest research basin (basin 1). At each location the amount of surface lowering and surface snow density were measured several times each day and their product yielded the rate of melt at the snowpack surface.

2.4.2 Snow Survey

To estimate total snow storage in the basin, an extensive snow survey was carried out. This survey was based on two assumptions:

(1) no melting occurs during the long Arctic winter so that a snow survey carried out prior to the spring melt will provide data on the total amount of snow accumulated in winter.

(2) topography exerts a strong control on the distribution of snow so that various terrain units will have characteristic amounts of snow storage.

In a small basin on Axel Heiberg Island, Young (1969) observed extreme spatial variation in snow depth but noted that such variations could be related to various measures of the land surface geometry. In view of large variations in snow depth it is logistically impossible to accurately map the snow cover of a large area, but the relationship between snow storage and terrain type suggests that basin snow cover can be obtained as the areally weighed mean of the snow storages in various types of terrain. Following this second assumption, aerial photographs were used to divide the basin into various terrain types whose boundaries were confirmed in the field. Fifty-three snow survey traverses were carried out across various units. Each transect consisted of 10 to 50 sample points and the transect lines spanned the entire lengths of the terrain units. A 3 m steel pole was used to measure snow depth and a Meteorological Service of Canada snow sampler was used to determine snow density. Several densities thus determined were checked against the density obtained by taking a series of 250 cm^3 sample cores from vertical profiles exposed in snowpits.

The snow survey was conducted during the period May 16 to June 6, 1976. The snowpack did not undergo any significant melting during this period though several snow storms occured between May 19 and May 26. Fortunately, this new snow can be easily distinguished from the older snow (fig. 2.10). For consistency, the new snow was ignored and all snow data referred to the condition as of May 19, 1976. Due to the absence of melting in winter, the snow storage determined by the survey represents total snowfall since September 1, 1975, when the winter snow cover was first established.

2.4.3 Streamflow

Stream discharge was measured at six stations within the main basin (fig. 2.8). A major difficulty occured during the initial stages of runoff when water flowed along snow-lined channels that were constantly changing shape, resulting in an unstable stage-discharge relationship. During this period, discharge was directly obtained by the velocity-area method, with velocity determined by a Price-type current meter. Discharge was measured at least twice each day, approximately

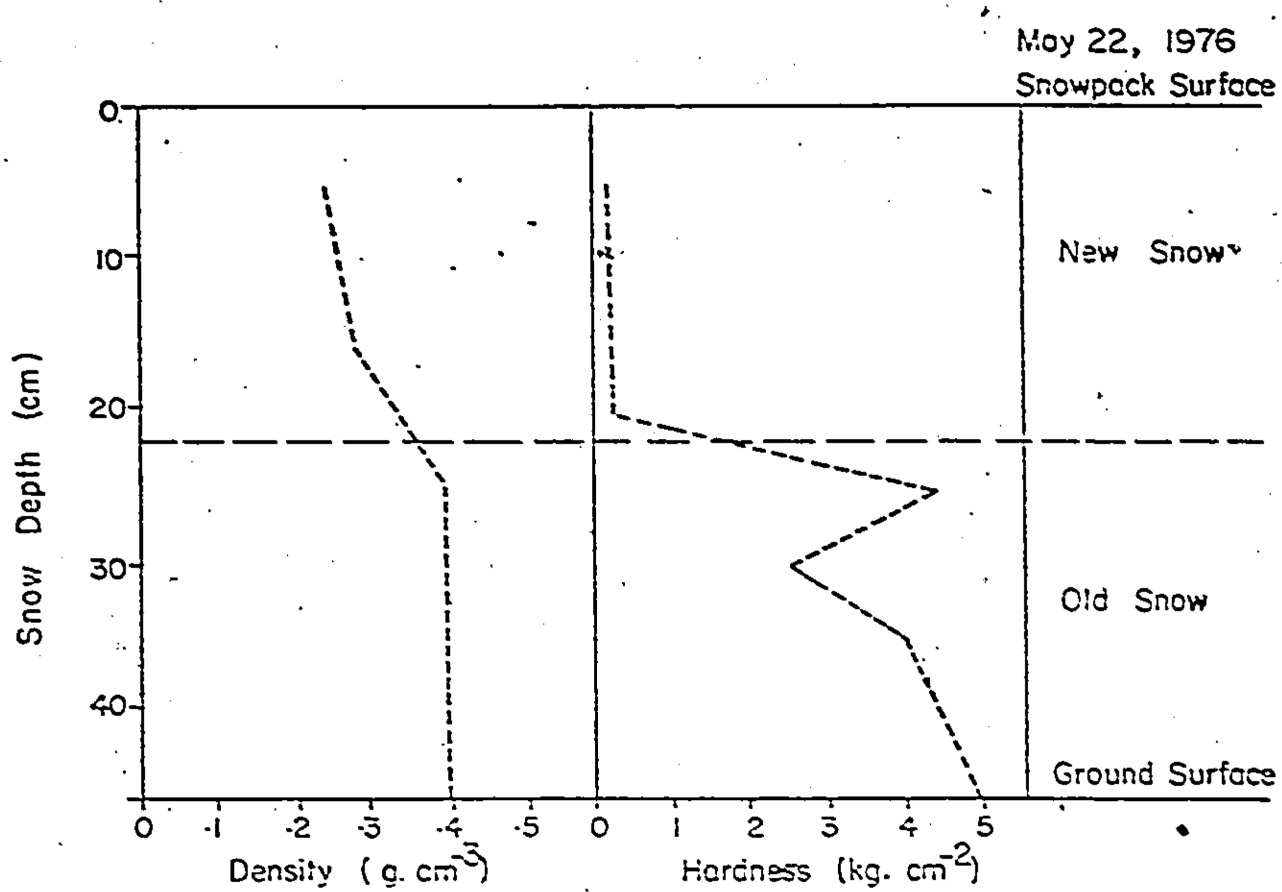


Fig. 2.10

Differentiation of new and old snow by snow hardness and density during the high flow and the low flow stages. As soon as streamflow was confined in a stable channel, a stage-discharge relationship was established for each gauging station.

Discharge was then obtained from the stage records which were read manually off the staff gauges at sites 1 upper, 1 middle and 3. Stages were recorded by an Ott and two Leopold-Stevens Type F water-level recorders at sites 1, 2 and 4.

CHAPTER 3

COMPONENTS OF THE WATER BALANCE

3.1 Components of the Water Balance

The water balance of any drainage basin can be written in the following simplified form

$$I - O \pm \Delta S = 0 \tag{3.1}$$

where: I is input

O is output

ΔS is change in storage.

This equation may be resolved into different components and written in a more comprehensive form. However, for different environmental conditions the relative importance of various components change. For high Arctic environments, the following components should be considered:

(1) Inputs - (a) snowfall (S)

(b) rainfall (R)

(c) blowing snow

(2) Outputs - (a) streamflow (Q)

(b) evaporation (E)

(c) groundwater flow

(3) Storage - (a) lakes and ponds

(b) soil moisture

(c) groundwater

(d) ground ice

The magnitude of some of these components is extremely small compared to the total flux of water, while some of the storage components change very little from year to year and/or have very small absolute magnitudes. The following components meet these criteria and therefore can be ignored in the water balance equation, thereby simplifying the use of the water balance approach without seriously affecting its accuracy:

(1) Blowing snow can be ignored in the present study because total basin snow storage was obtained by a snow survey carried out prior to the melt period. Any redistribution of snow during winter will be taken into consideration by the snow survey. In summer, the number of snowfall events is limited, and the residual snowpack is not prone to severe wind blown action.

(2) Groundwater flow is restricted to the very shallow active layer. Therefore, subsurface discharge from the basin is minimal.

(3) Lakes are absent in the basin and ponds contain a small volume of water compared to the total incoming precipitation. Therefore any variation in storage from year to year would constitute a very small change in basin storage.

(4) Soil moisture can vary greatly over the summer period, but by the end of the season most parts of the basin are dry. At that time, the amount of water stored in the non-saturated portion of the active layer would be very small and this condition varies little from year to year.

(5) Groundwater storage is also confined to the thin active layer. Soon after snowmelt, suprapermafrost groundwater exists over much of the basin, but by late summer, it is confined to poorly drained regions or below late lying snow patches. Therefore its total volume is also very small and varies little from year to year.

(6) Position of the permafrost table is stable, so that a change in ground ice compared to basin water balance will be insignificant.

The above discussion shows that the amount of water stored in the basin is very small at the end of summer and the change in storage from year to year is negligible. This conclusion is substantiated by the observation that by late summer, streamflow depends entirely on water supply from the active layer. At this time, streamflow becomes extremely low, if not completely ceased, indicating that the water storage components are also very low, and this is a late summer condition which is normally found in the study basin.

Since two of the input and output terms can be ignored and the change in water storage can be approximated by zero over a one year period, the water balance equation is simplified to

$$(S + R) - (Q + E) = 0$$

Of these terms snowfall, rainfall and stream discharge were measured, while evaporation will be calculated using an evaporation model.

3.2 Precipitation

3.2.1 Snow

3.2.1.1 Terrain Units

Granberg (1972) suggested a relationship between terrain types and snow storage characteristics for a subarctic environment near Schefferville. For the study basin, several terrain types were recognized from the aerial photographs and their boundaries confirmed in the field. The following terrain types were then used as the basis for the survey (fig. 3.1):

(1) hilltops: normally rounded or rolling areas which are fairly exposed and which occur as ridge crests or hilltops grading into valleys or long slopes

(2) high flats: extensive, exposed areas which correspond with the plateaus in the basin

(3) low flats: lowlying flat areas which are often less exposed than the high flats due to the presence of low hills or valley walls nearby

(4) gullies: troughs with a depth of less than 4 m, being usually broader than they are deep

(5) valleys: troughs which are larger than the gullies

(6) long slopes: slopes which dip in one general direction and which are sufficiently extensive both in length and in width

The percentage distribution of each terrain type in the basins is summarized in fig. 3.2.

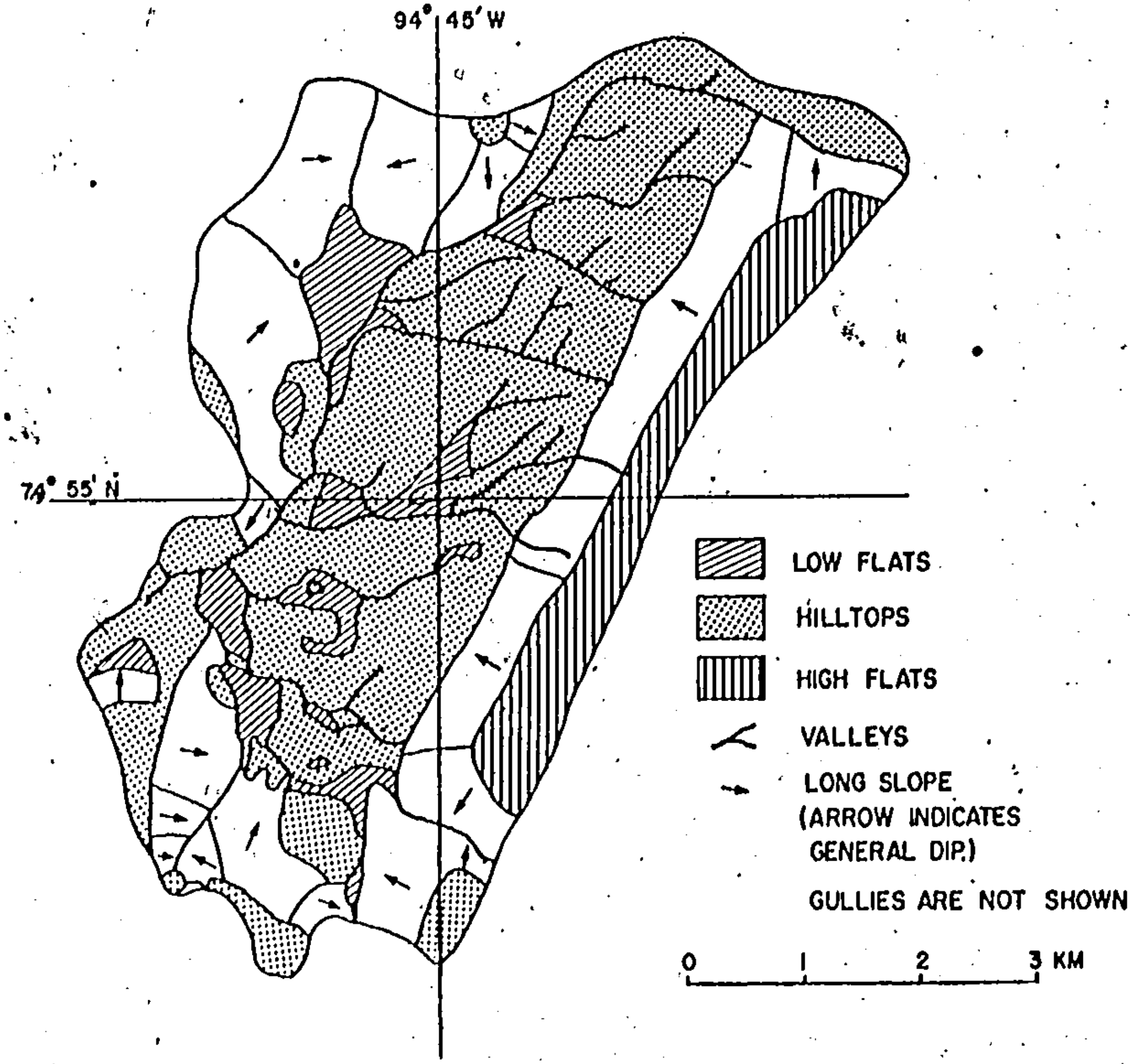


Fig. 3.1
Map showing terrain units used in snow survey

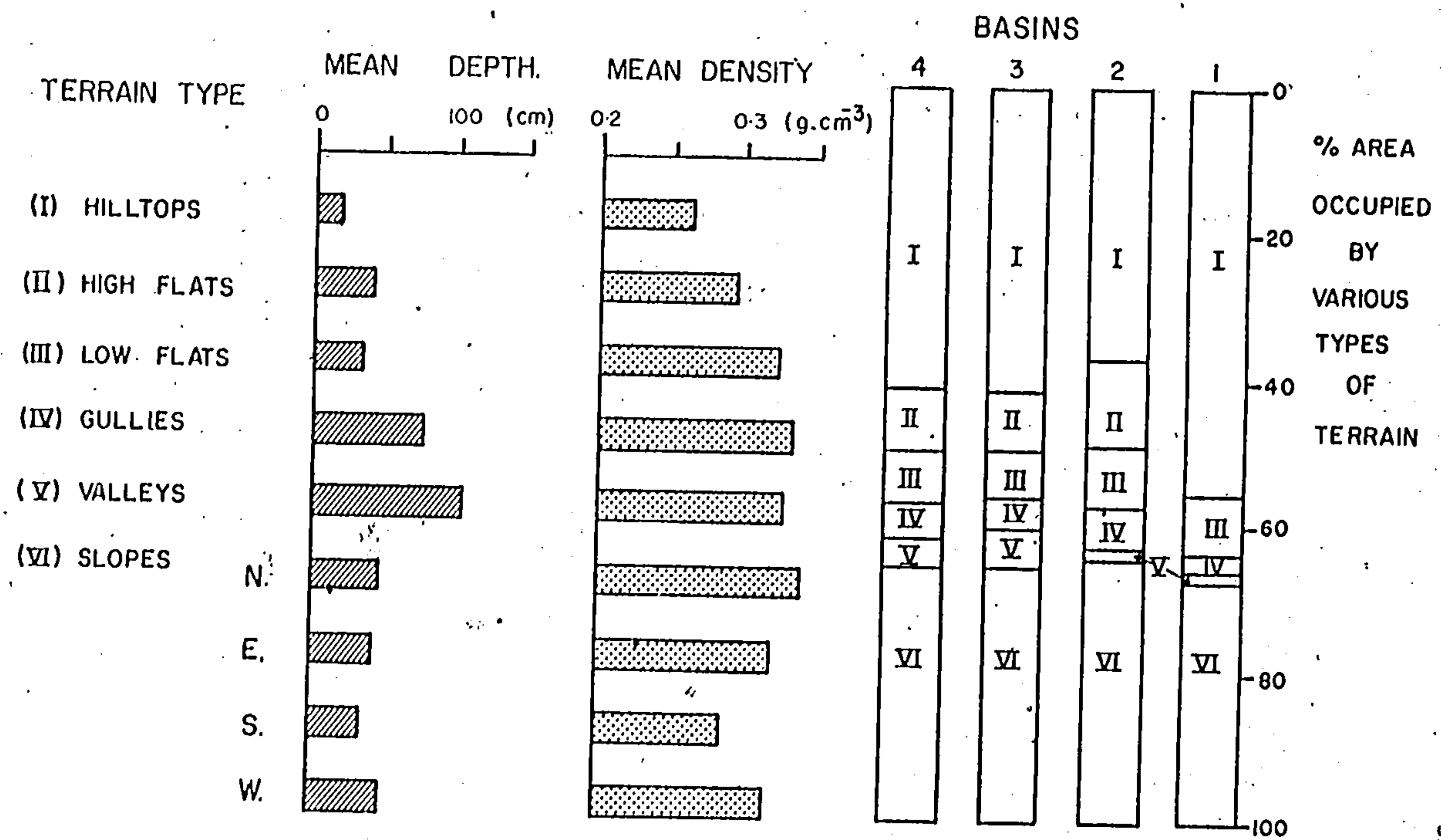


Fig. 3.2

Percent area of each terrain unit and their mean depths and densities as determined by the snow survey

3.2.1.2 Snow depth and density

The variability of snow depth within each terrain unit depends on the terrain type. Hilltops and flat areas, for instance, showed a lesser degree of variability (fig. 3.3) compared with the large variations in snow depth across river valleys (fig. 3.4). Individual slopes also had large variations, due to the presence of rock ledges and changes in slope. This source of variation was eliminated when only the average depth from each transect across a terrain unit was considered. Using the averages from all the transects, figure 3.2 shows that hilltops had the thinnest cover, followed by the flats and the slopes, while the deepest packs accured in the gullies and valleys.

The distribution of snow density also varies with terrain types (fig. 3.5). Densities increased from the exposed hilltops to the more sheltered lowlying flats. In the gullies and the valleys, the snow was more compacted and the densities increased. For snow lying on slopes, the lowest densities occured at the south facing aspect while the highest occured at the north-facing aspect. Intermediate densities occured on east and west-facing slopes. Mean densities for each terrain type are shown in figure 3.2.

Once average depths and densities for each terrain type were obtained, the following equation was used to compute basin snow storage.

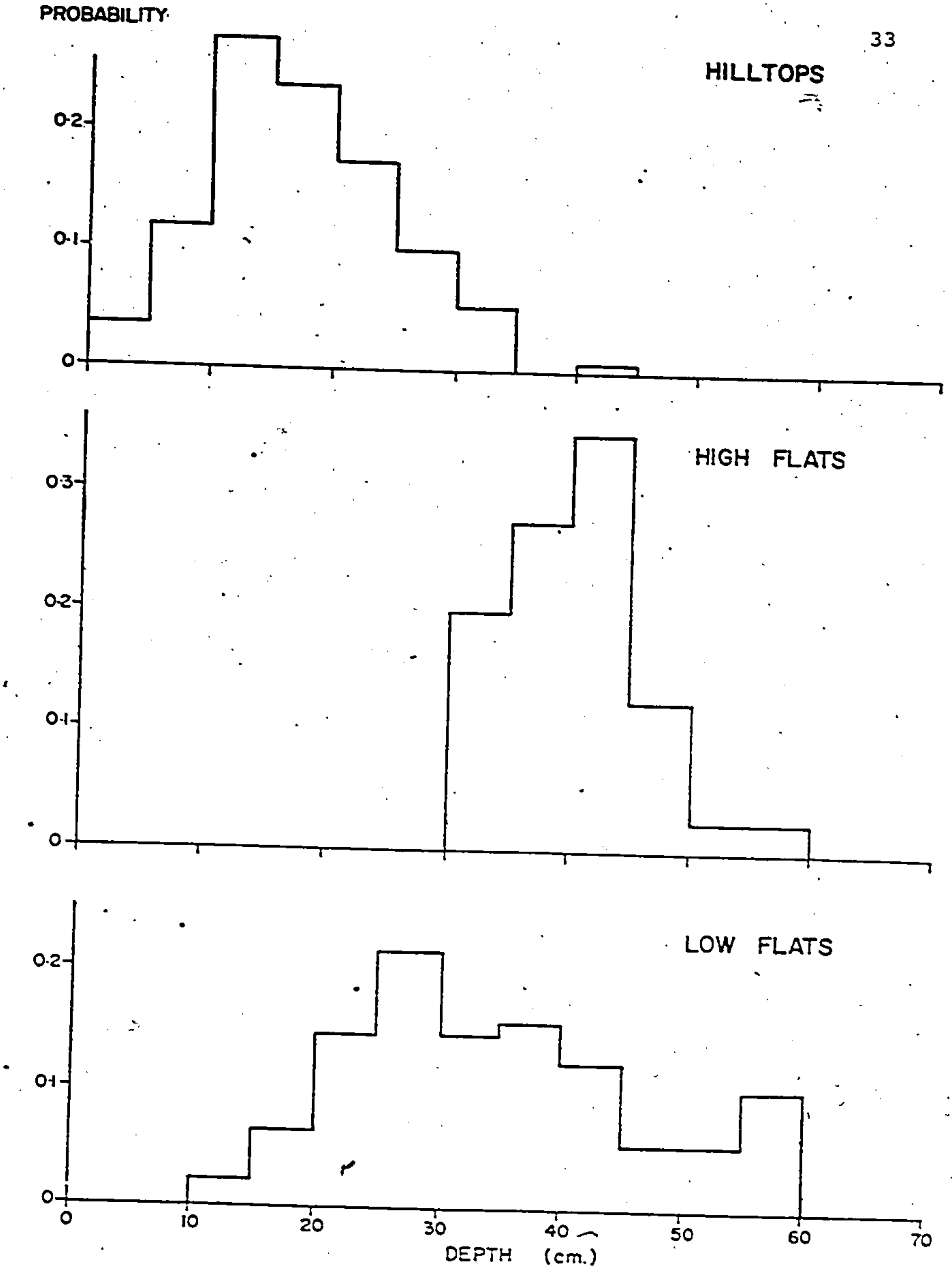
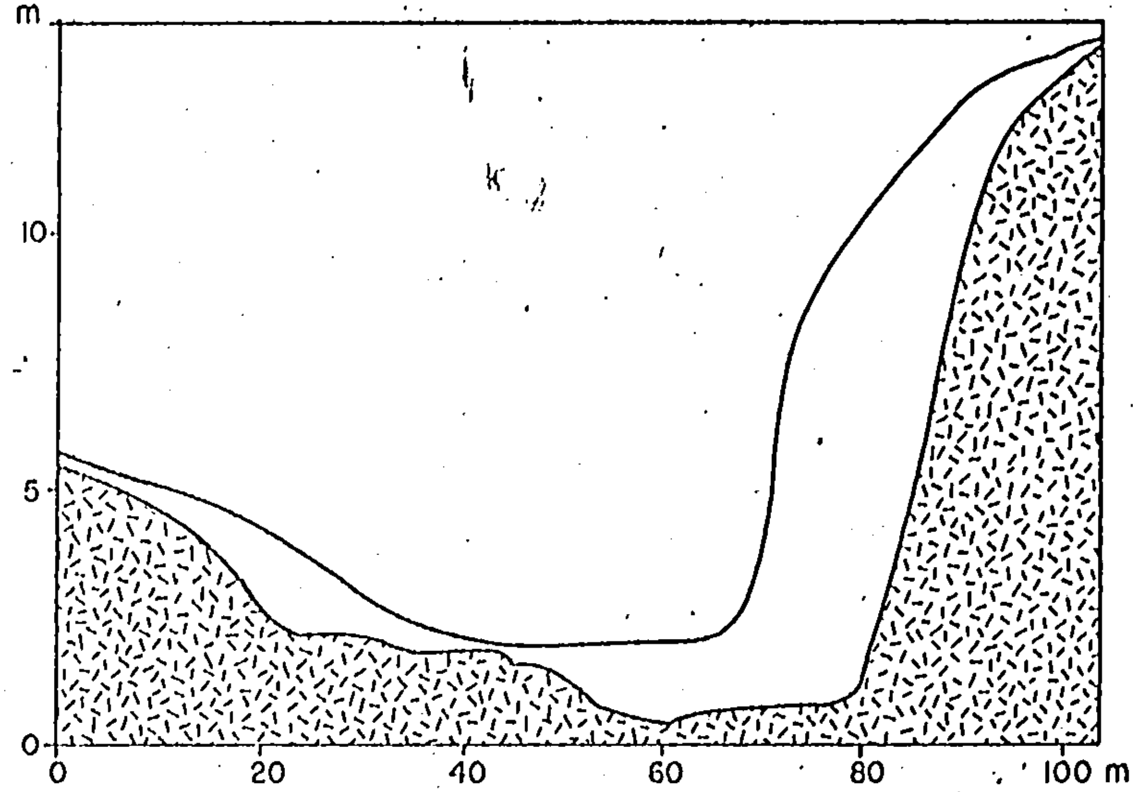
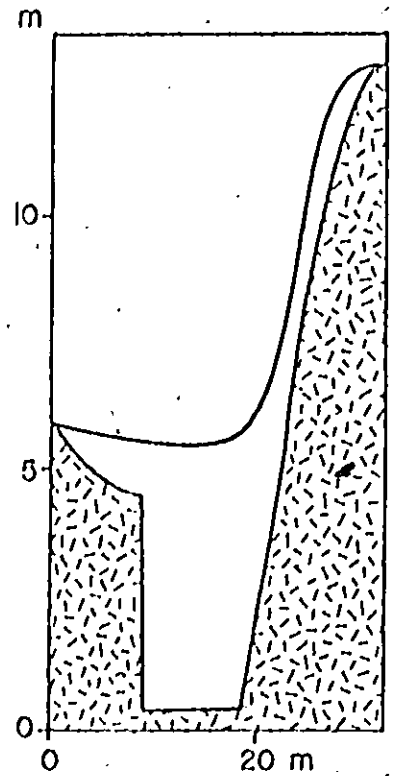


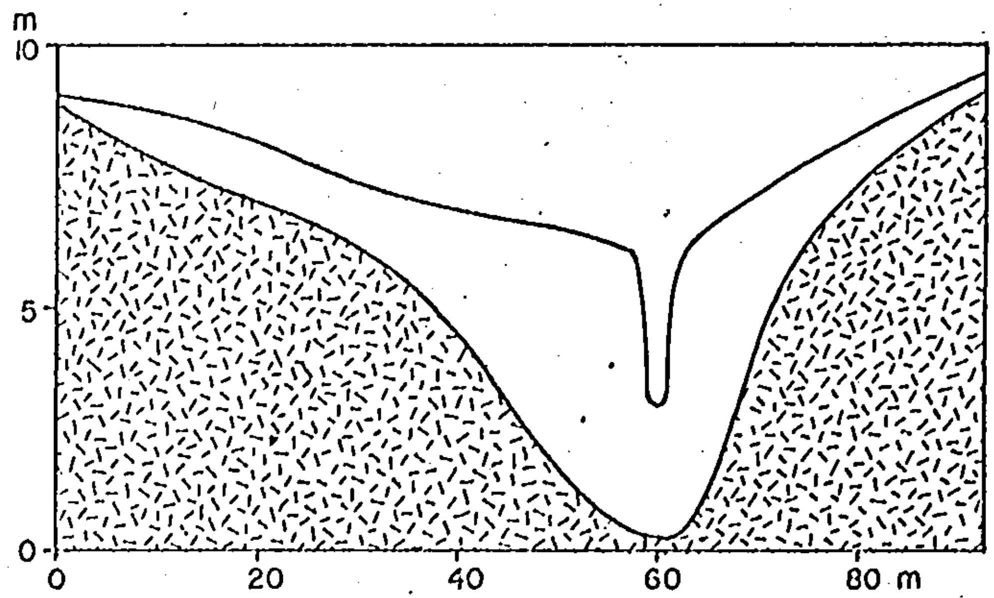
Fig. 3.3
Distribution of snow depths for hilltops, high flats & low flats

Fig. 3.4

Cross sections of three river valleys showing typical snow profiles







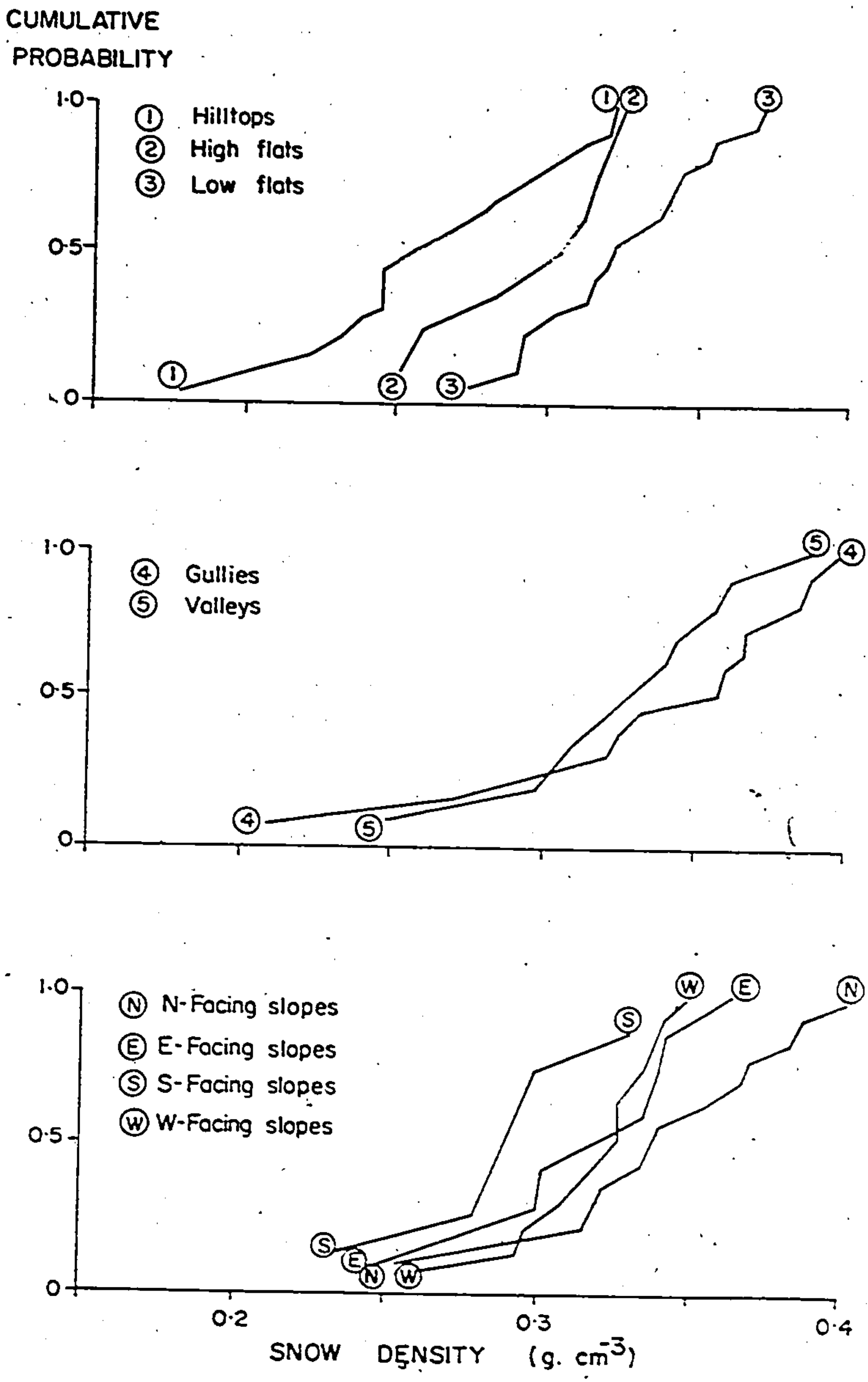


Fig. 3.5
Distribution of snow densities for each terrain unit

$$S = \sum_{i=1}^{m} \rho i \; di \; ai \qquad (3.2)$$

where S is snow storage expressed in water equivalent unit

i is the i^{th} terrain type, with a total of m terrain types in the basin

ρi is the mean snow density in terrain type i

di is the mean depth in terrain type i

ai is the area of terrain type i expressed as a fraction of total basin area.

Using this equation, mean snow storage for the four basins are found to be 122 mm for basin 1, 110 mm for basin 2, 127 mm for basin 3 and 122 mm for basin 4. These figures represent the snow accumulated between September 1, 1975 and May 19, 1976.

To obtain an estimate of the error involved in these values, two procedures were followed. Firstly, basin snow cover was mapped for the smallest basin (basin 1), and the result from the mapping was compared with the value computed by equation 3.2. At this scale, mapping is the most accurate method to determine basin snow cover. Secondly, an error analysis was performed to allow an estimate of the statistical error involved in the computed snow storage values and to determine if it is possible to reduce the number of terrain types without increasing the error significantly. The purpose of reducing the number of terrain types is to expediate future snow surveys.

The small basin snow cover was mapped from data collected from 8 profiles across the basin, involving 220 individual point samples. Using the map information (fig. 3.6) mean snow

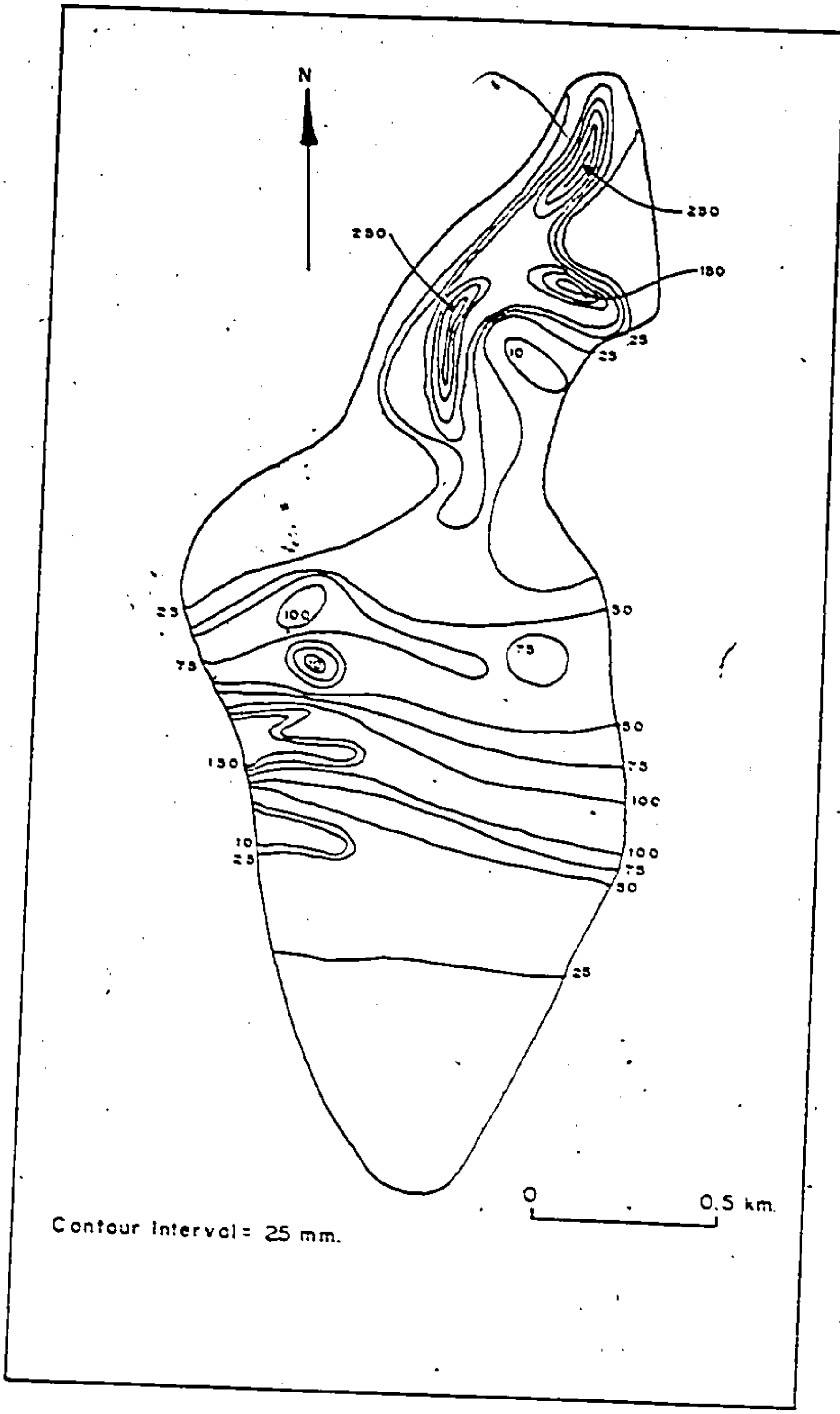


Fig. 3.6
Snow water equivalent (mm) for Basin 1 as of June 1, 1976.
Mean value was 136 mm

water equivalent for this basin was determined to be 136 mm, which is about 10 percent higher than the value of 122 mm obtained by equation 3.2. This close agreement between the two methods, indicates that the use of terrain units to determine basin snow cover produces results similar to the more accurate mapping technique.

For each basin, snow storage was determined using equation 3.2. Each of these independent values, ρi, di and ai has a possible error, therefore the computed snow storage must also have an error associated with it. The magnitude of this error is dependent on the magnitude of the errors in its independent components and the nature of the equation itself. The equation does not contribute directly to the error in the result, but modifies or propagates the errors already present (Fogel 1962).

The probable error R for any function of independent variables is (Scarborough 1960).

$$R^2 = \sum_{j=1}^{n} \left(\frac{dY}{dXj}\right)^2 \delta X_j^2 \tag{3.3}$$

where X_j is the j^{th} independent variable, $Y = f(X_1, X_2, \ldots, X_n)$ and δ_j^2 is the variance of X_j. For the snow storage information,

$$RS = \{ \sum_{i=1}^{m} [(\frac{ds}{d\rho i})^2 \delta_{\rho i}^2 + (\frac{ds}{ddi})^2 \delta d_i^2 + (\frac{ds}{dai})^2 \delta a_i^2] \}^{1/2} \tag{3.4}$$

$$= \{ \sum_{i=1}^{m} [(diai)^2 \delta \rho_i^2 + (\rho i\ ai)^2 \delta di + (\rho i\ di)^2\ \delta a_i^2] \}^{1/2}$$

The maximum error Rm in the calculated values of $Y = f(X_1, X_2, \ldots, X_n)$ is computed as (Fogel 1962)

$$Rm = \sum_{j=1}^{n} \left(\frac{dY}{dX_j}\right) \delta X_j , \tag{3.5}$$

In the determination of snow storage

$$Rmax = \sum_{i=1}^{m} (di\ ai\ \delta\rho_i + \rho i\ ai\ \delta di + \rho i\ di\ \delta ai) \tag{3.6}$$

Table 3.1 shows the calculated errors for all basins. Probable error varied from 10 to 22 mm which corresponds to an error of 9 to 15%, while maximum error varied from 41 to 56 mm. The distribution of maximum error among the various terrain types is shown in fig. 3.7. Attempts were made to further simplify the terrain types by pooling all slopes or by grouping the hilltops with the flats. The result was a rapid increase, both in percentage error and maximum error. If a percentage error of about 15% is desired, only the high and low flats may be combined. Hence for terrain similar to that of the study basins, a subdivision into hilltops, flats, slopes by various aspects, gullies and valleys is sufficient. This low level of error ensures that the snow survey results are sufficiently accurate for the purposes of this study.

During the period of snow accumulation prior to the snow survey, the Resolute meteorological station recorded 0.78 m of cumulative snowfall which converts to 63 mm in water equivalent unit (fig. 3.8). This value is less than the 29 year

Table 3.1 Computed snow storage and corresponding errors

Location	Snow storage(mm)	Probable error(mm)	Percentage error(%)	Maximum error(mm)
Basin 1	122	18	15	44
Basin 2	115	10	9	41
Basin 3	124	15	12	56
Basin 4	121	13	11	49
Resolute station	66			

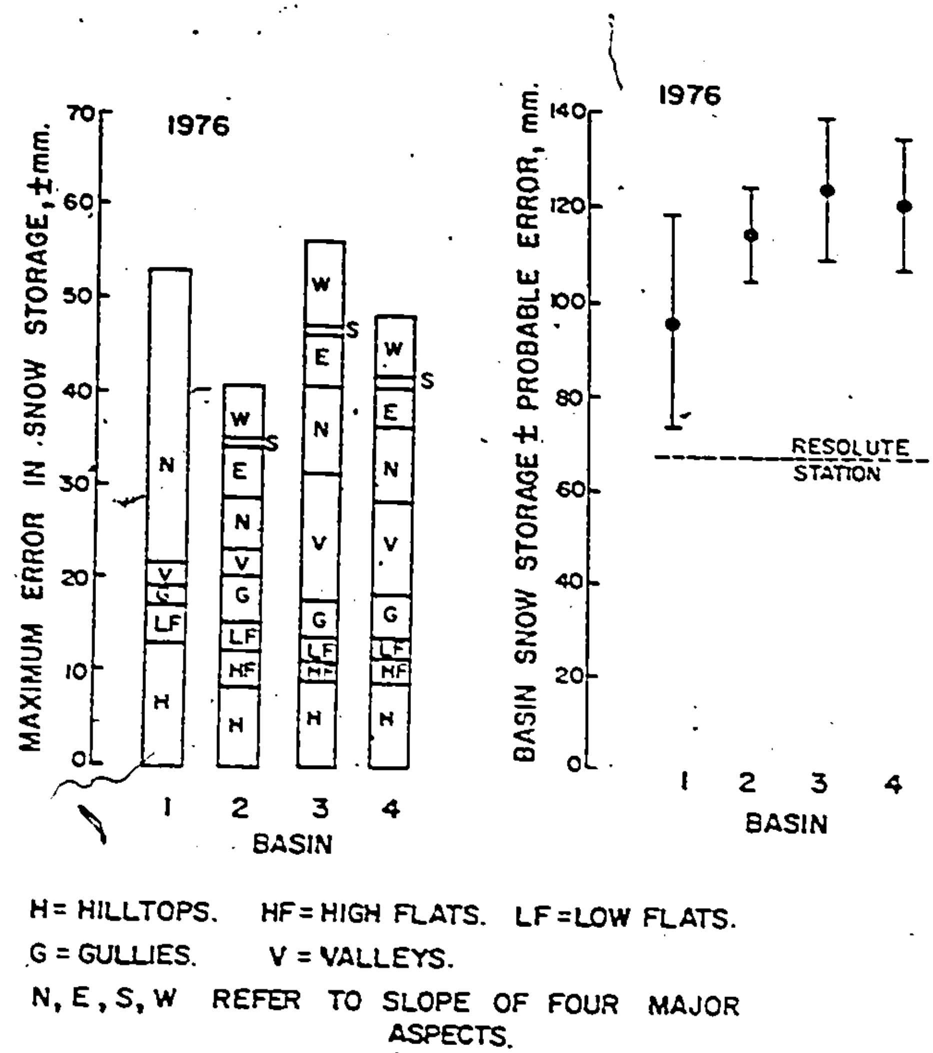


Fig. 3.7

Computed probable error for each basin and terrain unit

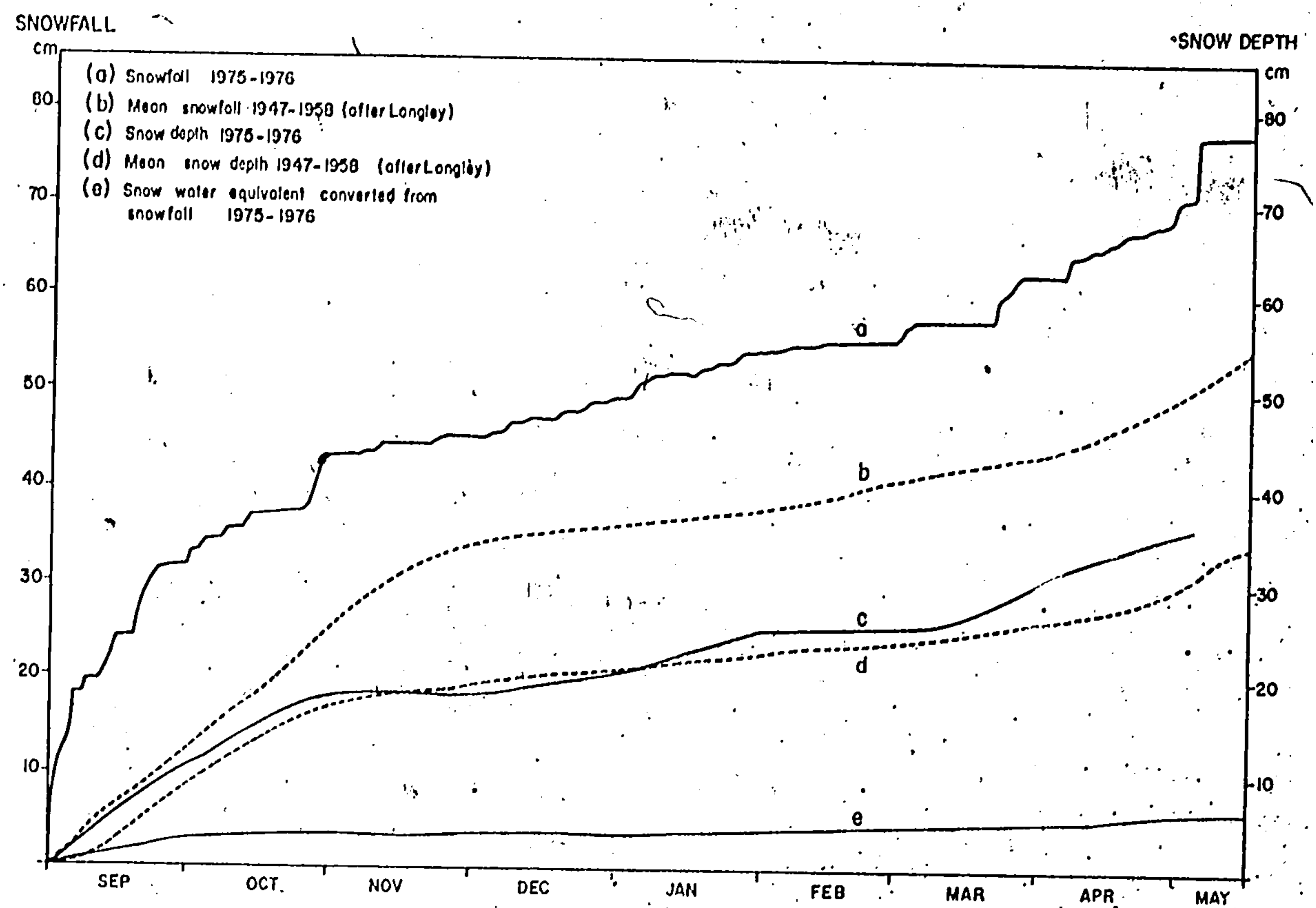


Fig. 3.8
Comparison of 1975-76 snowfall and snow depth recorded at Resolute Meteorological station with the ten-year station mean

average (1941 - 1970) reported by the Atmospheric Environment Service. A mid-May snow survey carried out by the Atmospheric Environment Service at the meterorological station yielded a value of 66 mm, indicating that cumulative snowfall and the snow survey data at the meterorological station were comparable.

Compared with the basin snow survey results, snowfall data obtained by the government weather station shows an under-estimation by over 50 percent. It is therefore difficult to estimate basin snow storage directly with weather station records. To overcome this difficulty weighting factors were applied to this station data to obtain an estimate of basin snowfall for the period May 20 to August 20, 1976 and also to calculate the amount of new snow storage in each terrain unit for the same period. This last calculation will be used in calculating snow storage throughout the melt season (see section 3.2.1.3). In the first case total basin snowfall over the period May 20 to August 20, 1976 was calculated using the following equation.

$$S2 = P2\left(\frac{S}{P1}\right) \tag{3.7}$$

where $S2$ = basin snowfall in water equivalent units

$P2$ = snowfall measured by the Atmospheric Environment Service from May 20 to August 20, 1976

S = basin snowstorage on May 19, 1976, as obtained from equation 3.2

$P1$ = snowfall measured by the Atmospheric Environment Service over the period Sept. 1, 1975 to May 19, 1976.

To estimate daily snow storage changes in each terrain type, after the snow survey of May 19, the following procedure was used

$$S3i = Si + P3\left(\frac{Si}{P1}\right) \qquad (3.8)$$

where i is the i^{th} terrain unit

S is the snow storage expressed in water equivalent units, as of May 19, 1976

S3 is the new snow storage value

P3 is daily snowfall measured at the government weather station over the Period May 20 to August 20, 1976

P1 is snowfall measured at the government weather station over the period Sept. 1, 1976 to May 19, 1976.

Although cumulative snowfall for the winter of 1975-76 was lower than the average, snow depth at the weather station remained similar to the long term average reported by Longley (1960). This reinforces Longley's observation of a snowpack near Resolute which built up to a characteristic depth in early winter, but remained little changed until the melt season. This observation agrees with Tabler's (1975) finding that for various topographic snow accumulation areas snow builds up to a maximum depth and these maxima cannot be exceeded regardless of the amount of blowing snow. At the maximum capacity, the snow surface reflects an equilibrium profile which may not be attained when the amount of snow drift is limited.

3.2.1.3 Snowmelt

The amount of snowmelt at any given point depends on the energy available. The various sources of energy include (US Army 1956)

$$M_E = M_{Rad} + M_{conv} + M_{cond} + M_G + M_R \quad (3.9)$$

where M_E is total energy available for snowmelt; which can be partitioned into components due to radiation (M_{Rad}), convection (M_{conv}), Conduction (M_{cond}), ground heat flux (M_G) and rain (M_R).

During days without rainfall, most of the energy available for snowmelt comes from radiation (Petzold 1974, Woo 1976) so that snowmelt can be estimated by a statistical relationship with net radiation. A regression relationship was therefore obtained between the measured melt rates and net radiation measured over a snowpack. Owing to a large error involved in the measurement of melt at the snow surface, the melt data were closely scrutinized. Data were rejected:

(1) when the wire above the snow was not kept at a constant tension

(2) when melt occured at the snow surface as well as at a zone below the surface, causing very low or very high readings over short time periods

(3) when the cold content of the snow was increased in the early mornings, requiring energy to raise the snow temperature to 0°C.

The final regression analysis made use of thirty-five data points based on snowmelt for periods ranging from 1 to 3 hours. The regression relationship (fig. 3.9) is statistically significant at 99 percent probability (correlation coefficient were 0.88) with a standard error of 0.6 mm.

$$M = Q^*/.31 - 0.2 \tag{3.10}$$

where M is snowmelt in mm, Q* is Net Radiation in J/mm^2. This may be compared to the theoretical conversion of Q* to melt heat flux

$$M = (Q^*/333) / \rho \text{ ice} \tag{3.11}$$

where 333 is the energy necessary to melt 1 gram of ice (J/gram). ρ ice is the density of ice (ρ ice = .001 $gm\ mm^{-3}$).

Equation 3.10 will be used in the estimation of snowmelt for the entire drainage basin.

3.2.1.4 Snow distribution during the melt season

Spatial distribution of snow in a drainage basin during the snowmelt period has important applications. The snow covered areas contribute meltwater to streamflow while the snow-free areas are subject to evaporation (see section 3.4.3).

The areal distribution of snow was obtained by applying equation 3.10 to estimate daily snowmelt from a snow cover whose initial distribution was determined by a mid-May snow survey. For basins 2, 3 and 4 the entire basin was divided into grid

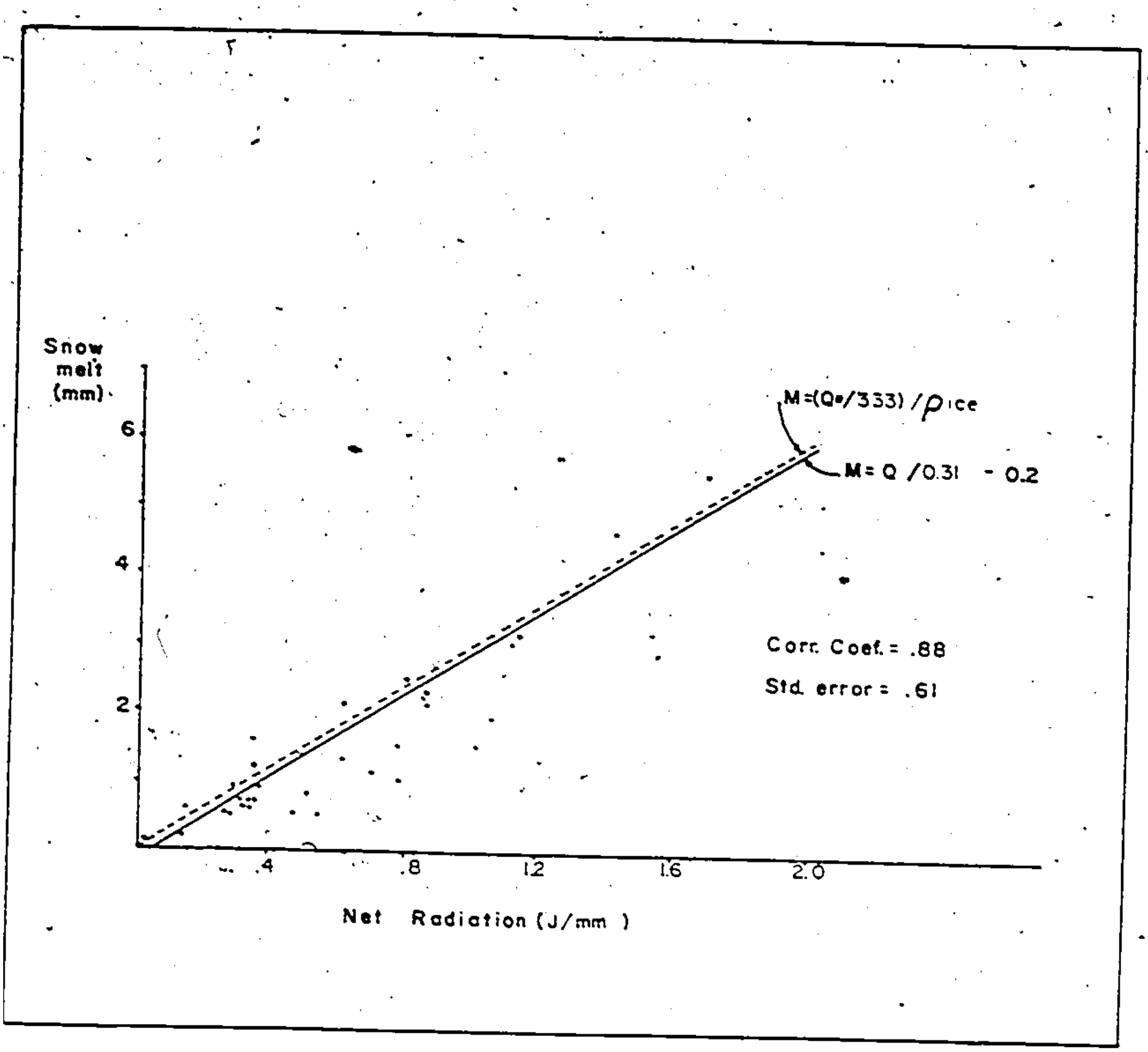


Fig. 3.9

Relationship between net radiation measured over a snow surface and snowmelt.

squares (125 m × 125 m) and the terrain type of the squares was determined from aerial photographs. Using a computer program, this information was mapped (fig. 3.10) and each terrain type was assigned a representative snow water equivalent value based on the snow data obtained for May 19, 1976 (fig. 3.11). For each day after this date, total snowmelt was calculated using equation 3.10, with the additional assumption that no melt occured if the mean hourly air temperature was below 0°C. Daily snowfall was added to each terrain type after the weather station snowfall data were adjusted by appropriate weighting factors (see Section 3.2.1.2). The snow budgetting procedure then allows the total basin snow storage and the proportion of basin snowfree area to be calculated for each day (fig. 3.11). Similar procedures were applied to basin 1, with the modifications that initial snow cover was based on the snow survey map (fig. 3.6) and a smaller grid size was used.

The present approach is prone to inaccuracies because of errors arising from the statistical snowmelt equation and because each terrain unit was assigned an average value for snow water equivalent. However, favourable results were obtained when the snow-free areas determined by the computer program was compared with the snow-free areas shown on panoramic photographs of the basin taken several times during the field season. A further check on snow coverage at the end of summer season was made by comparing the predicted values

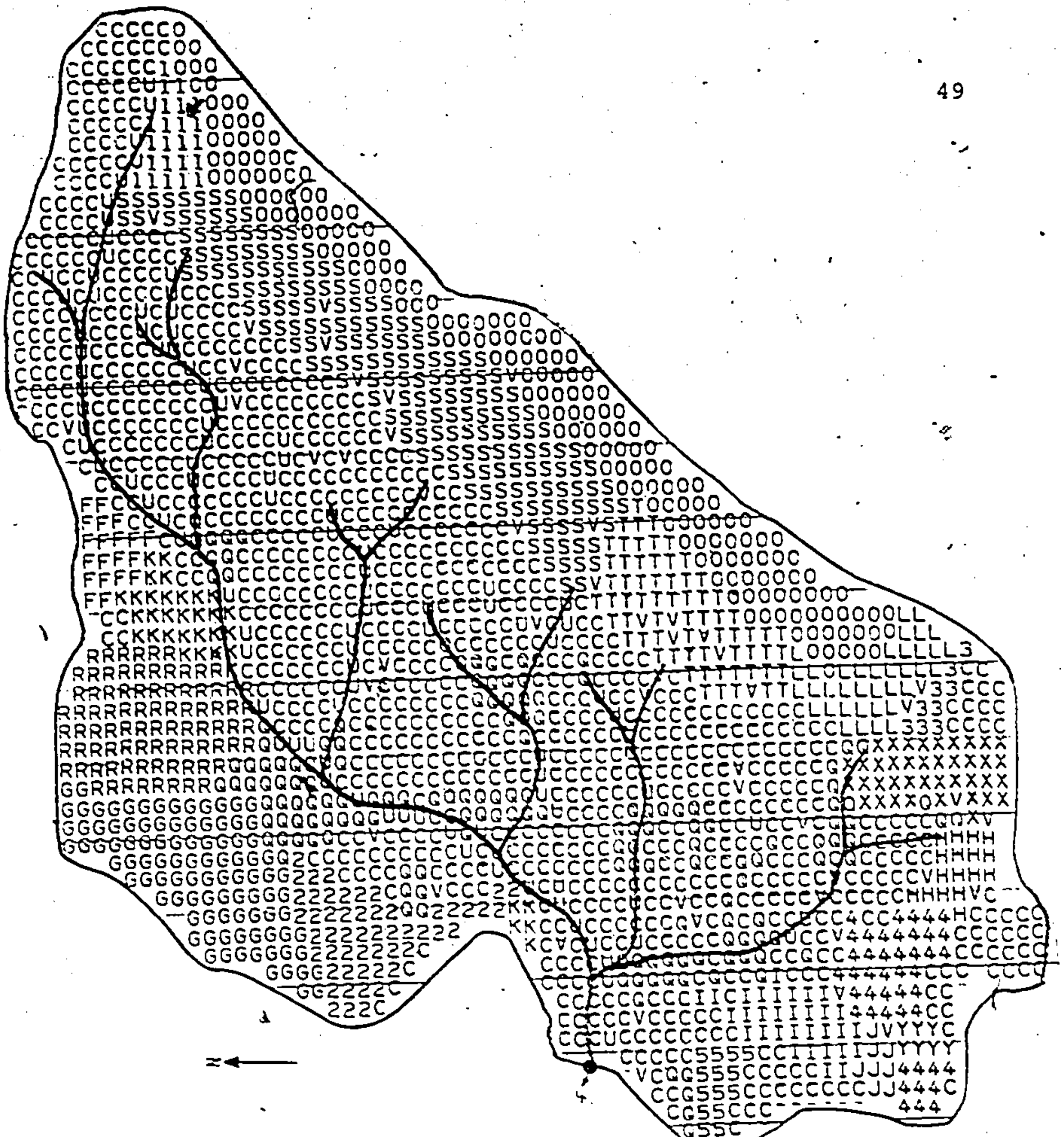
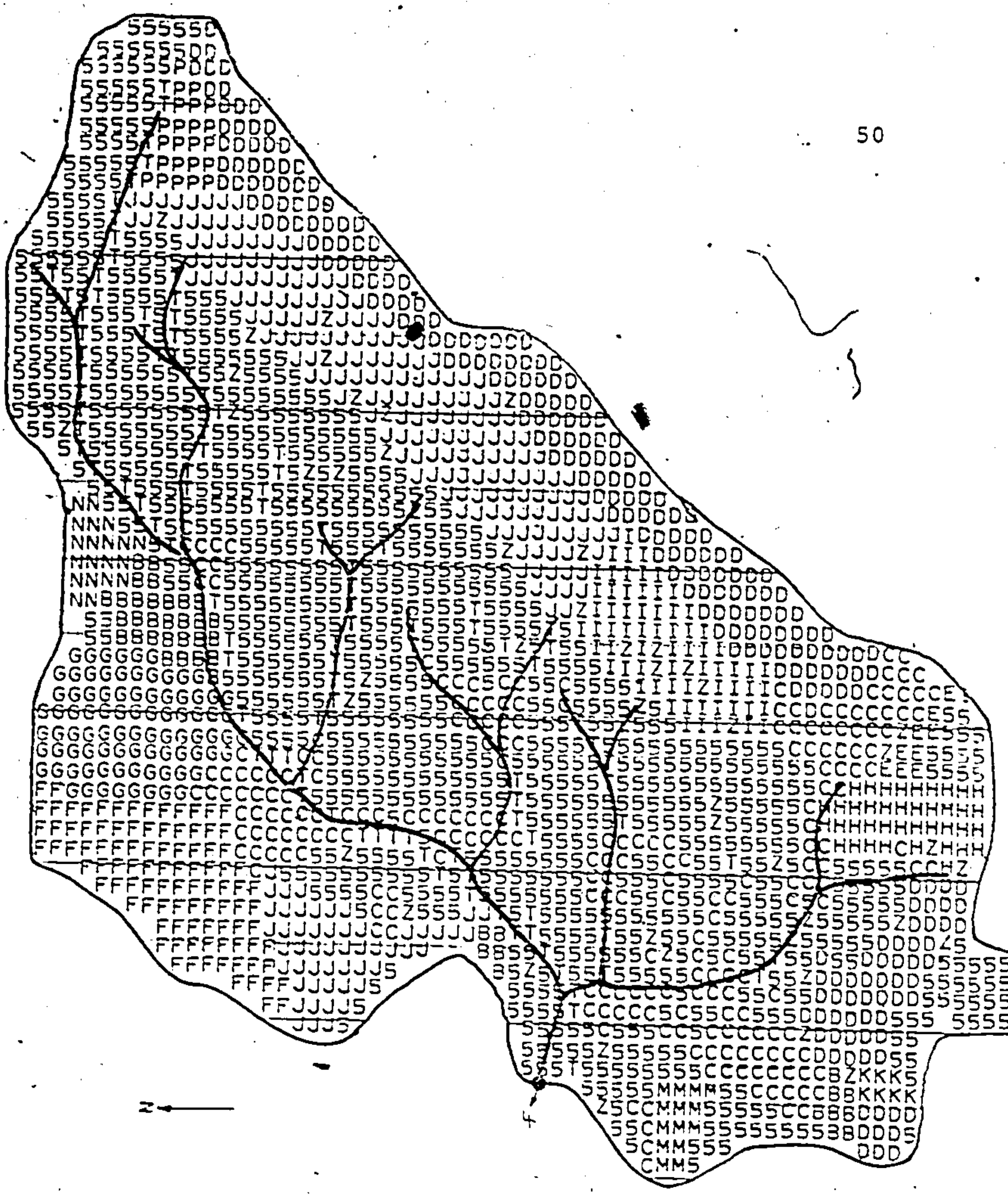


Fig. 3.10
Example of computerized terrain unit map used in snow survey. Each symbol represents a different terrain unit. Each terrain unit is given a characteristic snow storage value.



Day 006 210576

Mean W.E. = 121.5

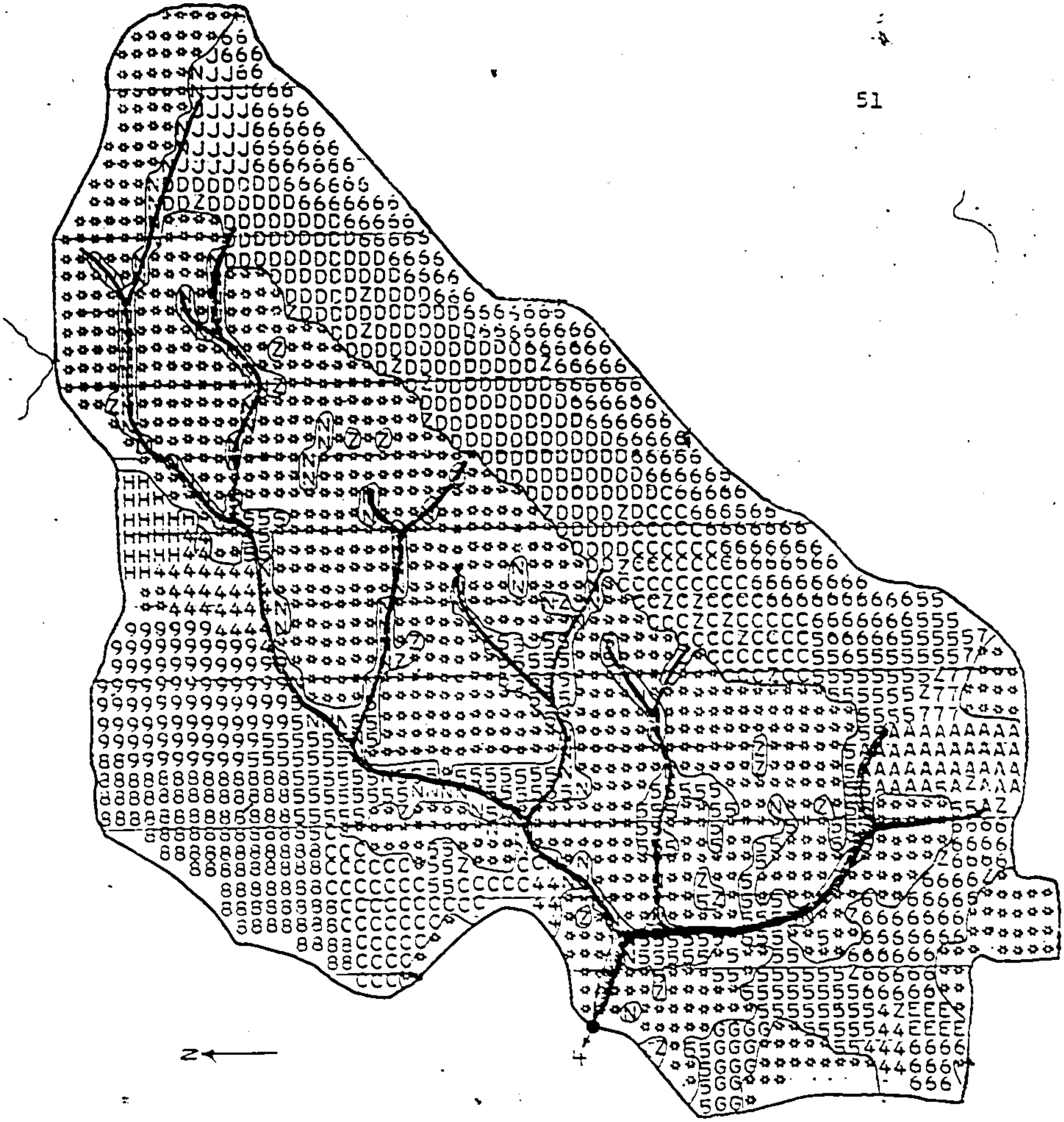
Proportion Bare = 0.000

New Snowfall at Station = 3.8

Daily Snowmelt = 0.00

Fig. 3.11

Example of computerized snow distribution map on two days (May 21 and July 5, 1976). Each symbol represents a different snow water equivalent. Star pattern on July 5 map represents bare ground.



Day 051 050776

Mean W.E. = 64.8

Proportion Bare = .419

New Snowfall at Station = 0.0

Daily Snowmelt = 9.03

Fig. 3.11

Example of computerized snow distribution map on two days (May 21 and July 5, 1976). Each symbol represents a different snow water equivalent. Star pattern on July 5 map represents bare ground.

to the snow pattern appearing in aerial photographs taken of the basin at the end of summer 1969. Both values agree closely. The procedure adopted by the present study is therefore considered to be adequate for determining daily changes in the percentage of snowfree areas in the drainage basins.

3.2.2 Rainfall

During the summer of 1976, the Resolute weather station (fig. 3.12) recorded 31 mm of rainfall, a value considerably lower than the long term mean of 59 mm (Dept. of Envir. 1972). June was wetter than average (9.4 mm compared to 5.8 mm), while July and August were drier (0.8 mm compared to a mean of 23.4 mm and 21.1 mm compared to 25.7 mm).

Ten manual raingauges were monitored throughout the study basin to check if the weather station data were representative of the basin rainfall. Six rain-storms were recorded all of which occured during the first three weeks of August. Westerly or northwesterly winds prevailed during four of the six storms, while the other remaining storms were accompanied by southeasterly winds. These limited data indicate ~~the~~ influence of basin topography (fig. 3.13) and wind direction. When the prevailing wind was westerly or northwesterly, orographic effect caused rainfall to increase with elevation. When southeasterly winds prevailed, however, the topographically higher parts of the basin received lower rainfall. This is the result of winds blowing downslope and therefore being unable to release as much rainfall.

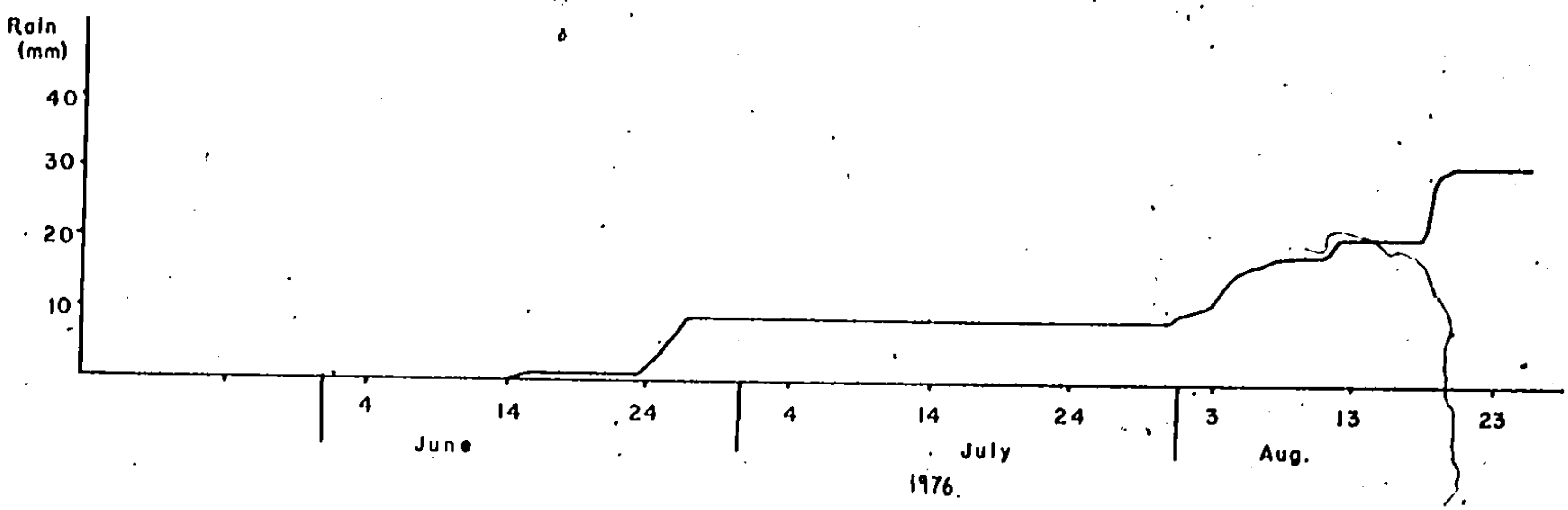


Fig. 3.12
Cumulative rainfall as measured by Atmospheric Environment Service

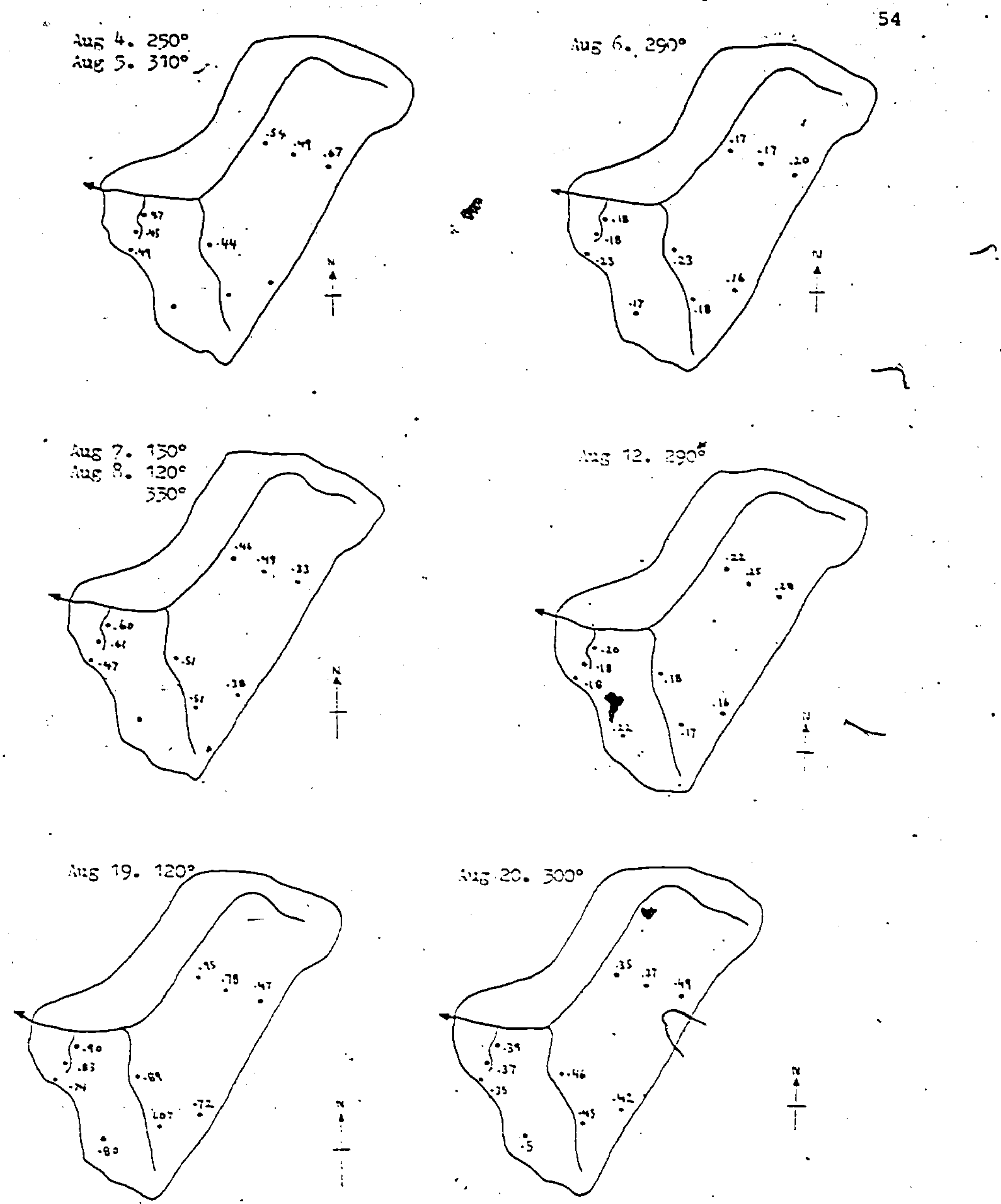


Fig. 3.13
Precipitation distribution for 6 storms in August, 1976 (wind direction in degrees follows date)

Mean basin rainfall was obtained using the Thiessen polygon method (fig. 3.14). For individual storms, the weather station data can differ substantially from the basin rainfall. For the entire month of August, however, basin rainfall totalled 26 mm which is similar to the 21 mm reported by the weather station. This analysis therefore indicates that the weather station rainfall was representative of the drainage basin over the entire summer.

For use in the water balance, it is necessary to estimate the probable errors involved with the above estimates of rainfall. One method to do this is to use the relationship between catch deficiency and wind speed determined from previous studies (Linsley et al 1975). An analysis of wind and precipitation during the summer of 1976 showed that 2% of rain occured while winds were between 0-8 km/h, 34% with winds between 8-16 km/h, 44% with winds between 16-24 km/h and 8% with winds between 24-32 km/h. Using these percentages as a weighting factor, a mean catch deficiency was obtained. It must be remembered that the relationship is based on wind speed at orifice height, while the wind speeds used in the analysis were taken by the Atmospheric Environment Service are at 10 m. Therefore the estimated catch deficiency of 12% is probably an overestimate.

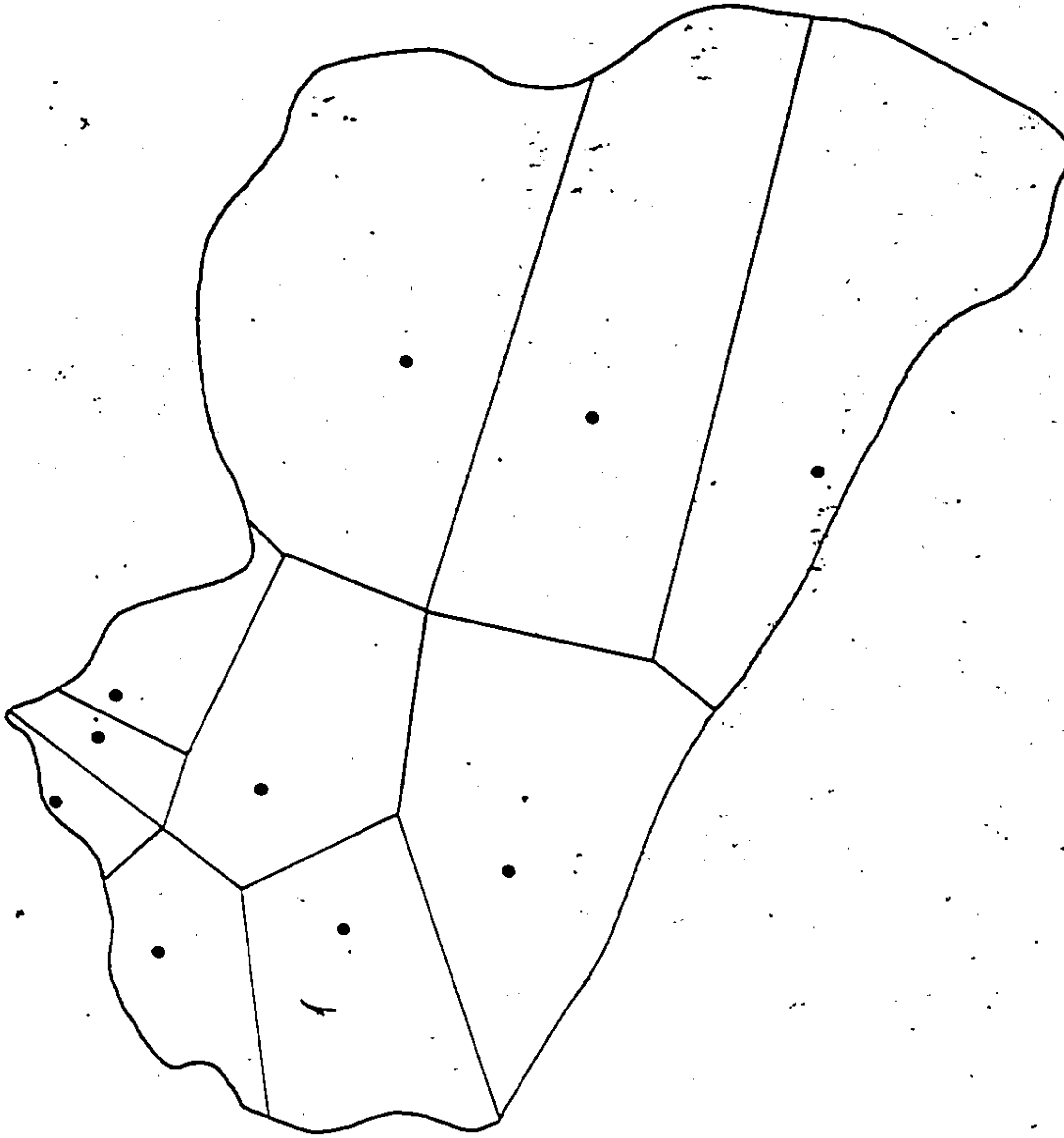


Fig. 3.14

Thiessen polygons for computing mean basin rainfall from the 10 manual rain gauges.

3.3 Streamflow

Streamflow in the study area typically follows the Arctic nival regime discussed by Chuch (1974). Nival-regime streams are dominated by large spring floods, but since there is no connection between the stream and the subpermafrost groundwater sources, streamflow ceases soon after the active layer freezes. The discharge pattern during the snowmelt period can be examined in terms of the variable source area concept. This concept was first applied to soils (Freeze 1972) but is equally applicable to snow hydrology where a variable snow cover and snow depth control the source area of runoff (Woo and Slaymaker 1975 and Woo 1976).

An application of the variable source area concept can be demonstrated by the streamflow records from the study basin (fig. 3.15). Streamflow commenced on June 30 at station 4, but a snow storm on July 2 arrested this flow, not to restart until July 5. At sites 2 and 3, flow began on July 9, and at site 1, on July 14. One general observation is that further up the basin, the first date of flow becomes progressively delayed. This implies that there is a continual upstream expansion of the basin area which supplies water to the stream. One phenomenon which affects meltwater flow is the pondage of water behind large snowdrifts across the valley. A pond drains rapidly as the snow blockage is broken, and for a downstream station, there is a sudden increase in its streamflow contributing area. An example of pond drainage occured on July 9, 1976.

5.8

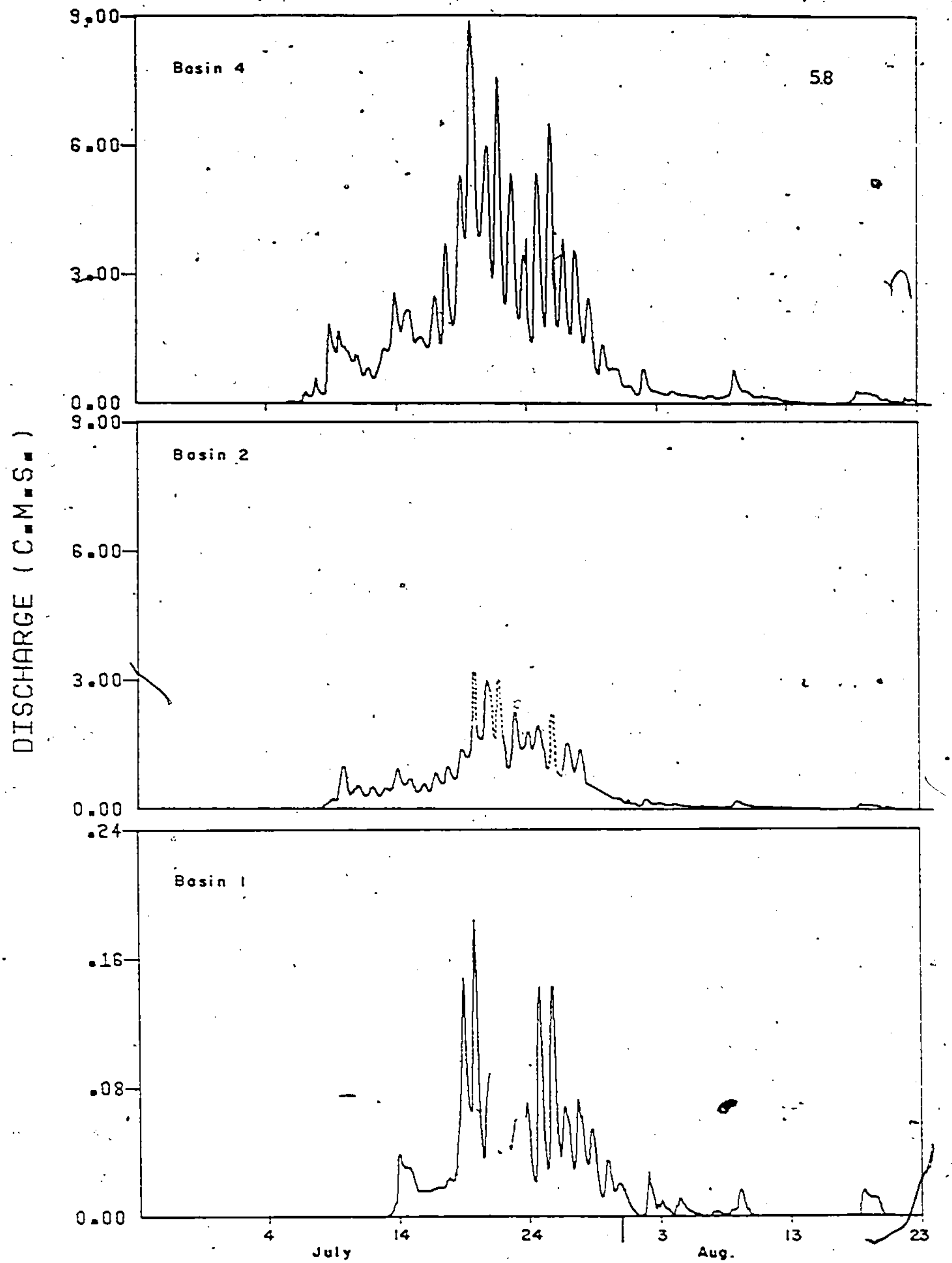


Fig. 3.15
Hydrographs for Basins 1,2,4 July & Aug. 1976.

At site 4, discharge was 0.3 m^3 s^{-1} at 14:00 h, but by 15:00 h, the flow was 1.0 m^3 s^{-1}. During the morning, a large amount of water was impounded behind a large snowdrift above site 3, allowing a flow of 0.01 m^3 s^{-1} to overflow the snow dam and to reach station 3. By 14:30 h downcutting through the snow dam led to pond drainage accompanied by a rapid flow increase downstream, thus abruptly expanding the streamflow contributing area for site 4. This extension of streamflow contributing area continued during the early snowmelt period, until streamflow channels were well established and integrated throughout the basin. By this time the potential contributing area corresponded with the total basin area.

During any time of the year, only a portion of the potential contributing area contributes meltwater to the stream. The reasons are: (1) areas far from the stream channel may not have developed flow connections with the channel or (2) some areas of the basin are snowfree. The 1976 flow records of the basin illustrates the discrepancy between the potential and the actual contributing areas. By July 9, streamflow occured at sites 2, 3 and 4, implying that the potential contributing area for site 4 extended above sites 2 and 3. However, streamflow did not begin at site 1 until July 14, so that basin 1 was not contributing streamflow to site 4 until after July 14. It can be seen therefore that the actual contributing area is that portion of the potential contributing area which supplies meltwater to the stream. Then given a steady melt rate, the highest spring melt discharge would occur when the actual contributing area is at a maximum.

As a consequence of the varying extents of the melt-water scource area several streamflow characteristics are expected: (1) for a given point along the stream, the major source of water will extend upstream progressively, thus increasing the lag time between peak melt and peak discharge and (2) the ratio between the discharges at any two points on the stream system should vary over the melt season, depending on the changing contributing area of each point (Woo 1976). An examination of the discharge ratios can therefore be used to study changes in the source area during the melt period.

Fig. 3.16 shows the discharge ratios between sites 2, 3 and 4. Note that discharge at sites 2 and 3 began on the same day, approximately 4 days after flow began at site 4. The ratios of the basin areas are 0.31 for basins 2:4 and 0.65 for basins 3:4. If the entire basins or a similar proportion of basins 2 and 3 were contributing to streamflow, the discharge ratios should equal the ratios of the meltwater contributing areas of each basin.

For the first 6 days after streamflow began, the discharge ratios for sites 2:4 and 3:4 increased rapidly, but both had equal values. This indicates that the actual contributing areas in both basins were similar, but were increasing at the same rate. Here an assumption is made that the melt rates were identical in both basins. By day 7, the discharge ratios

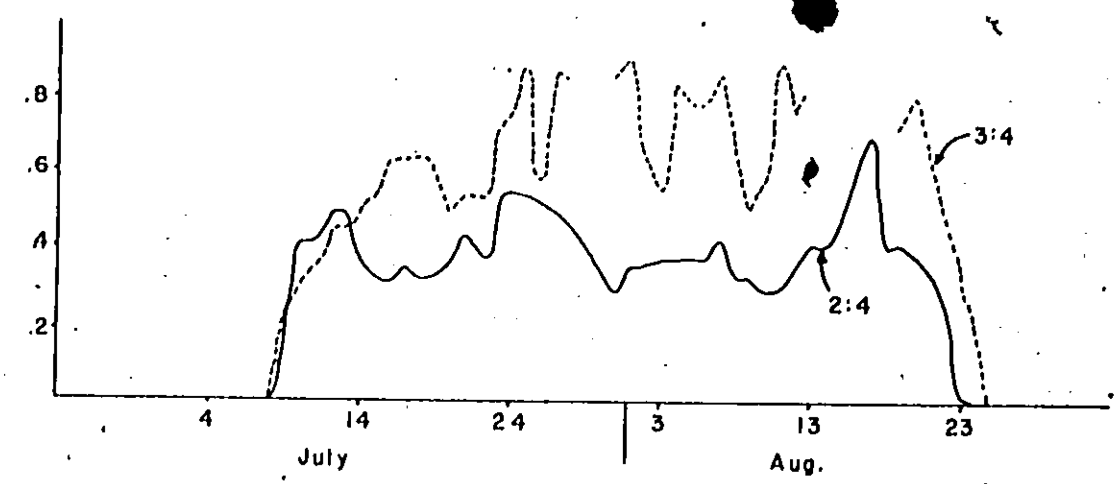


Fig. 3.16
Discharge ratios for streams 3:4 and 2:4. July & Aug. 1976.

for basins 3:4 continued to increase, but there was a decline in the ratio for basin 2:4, suggesting that the actual contributing area of basin 2 had reached a maximum but was still increasing for basin 3. Note that the discharge ratio for basin 3 gradually increased to a value approximately equal to that of its area ratio, and the discharge ratio for basin 2 gradually decreased until it was close to its area ratio. This indicates that by late in the melt period, the actual contributing areas of each basin are an equal percentage of the basin area.

For the water balance study, total stream discharge for the entire season was calculated by summing the mean discharge computed at two hourly intervals. The results are given in table 3.2.

Table 3.2 Total discharge for the four study basins.

Basin	Discharge (mm)
1	137 ± 15
2	157 ± 17
3	160 ± 18
4	161 ± 18

Total discharge for basins 2,3 and 4 are similar but the total for basin 1 is considerably less. The discrepancy cannot be due to measurement error because streamflow for all stations was obtained by identical means. The difference is attributed to the presence of a large, flat marshy area in basin 1, thus retarding streamflow but enhancing evaporation. The result was a reduction in the amount of water available for streamflow.

It is important to determine some error values for the above discharge figures, particularly when the water balance method is used to determine basin evaporation. Error involved in discharge calculations are introduced by instrumentation and technique in the initial stream gauging observation, by use of a stage-discharge relationship, by stilling well and stage recorder errors and from the method of calculating discharge values (Dickinson 1967).

Church and Kellerhals (1970) estimated the total error for discharge measurements by calculating the total pooled error based on systematic errors due to meter performance, errors due to velocity pulses, errors in determining depth and width and errors in estimating mean velocity for a vertical. They found that for the six-tenths method, the total error (95% confidence level) was 11% if 10 verticals were used, 8% if 15 were used and if 20 verticals were used, the error had dropped to 7%. Dickinson (1967) stated that the errors involved in the determination of mean stage were minimal and therefore the error in a single mean daily discharge estimate may be assumed to be equivalent to the error in any single discharge estimate for which the stage is given. From this fact, it can be seen that the above error values may be assumed to be representative of the error involved in estimating discharge from a stage record. Therefore a conservative estimate of the error involved on the given discharge estimates for the study basins would be in the 11% range (more than 10 verticals were normally used), and confidence limits can then be placed on the basin discharge estimates (table 3.2).

3.4 Evaporation

3.4.1 Evaporation Calculations

For the Arctic areas, direct measurement of evaporation is cumbersome. While most evaporation models require microclimatological data which can only be acquired at great expense, the equilibrium form of the combination model (Priestly and Taylor 1972, Davies and Allen 1973, Stewart and Rouse 1976) offers the possibility of evaporation estimation using readily obtained meterorological information. The combination model of evaporation was first developed by Penman (1948). One of the modified versions was presented by Slatyer and McIlroy (1961)

$$LE = \frac{S}{S+r}(Q^* - G) + \frac{\rho Cp}{ra}(Dz - Do) \qquad (3.12)$$

where: LE = latent heat flux

S = slope of the saturation vapour pressure versus temperature curve

r = psychrometric constant

Q^* = net radiation

G = ground heat flux

ρ = air density

Cp = specific heat of air at constant pressure

Dz, Do = wet bulb depressions in the overlying air and at the surface

ra = the aerodynamic resistance to the diffusion of water vapour between the surface and height z

Instead of evaluating all the terms on the right hand side of equation 3.12, recent works (Priestly & Taylor 1972, Stewart & Rouse 1976, Davies & Allen 1973) have shown that evaporation under all conditions can be expressed as a function of equilibium evaporation (LE eq)

$$LE = \alpha \; LE \; eq \tag{3.13}$$

where $LE \; eq = \frac{S}{S+r} (Q^* - G)$, and α is generally found to range between 0.0 and 1.26 (table 3.3). During a typical Arctic summer, the value of α is expected to vary according to changes in the soil moisture conditions. Although it is difficult to model short term changes in α, it is possible to assign average α values to provide seasonal estimates of evaporation for various types of surfaces. For an initial estimate of evaporation, two values of α were used.

(1) Under saturated conditions, previous studies have shown that α = 1.26. Therefore calculations based on this value will give an upper limit to evaporation.

(2) Under moderate moisture conditions, it is appropriate to

TABLE 3.3 Characteristic α values

Authors	α	
Priestly and Taylor (1972)	1.26	saturated surfaces
Stewart and Rouse (1976)	1.26	saturated sedge meadow
Stewart and Rouse (1976)	1.26	shallow pond
Davies and Allen (1973)	1.27	wet bare soils
Stewart and Rouse (1976)	0.95	upland ridge - wet soil covered with non tran-piring lichens which exhibit a strong resis-tance to vapour diffusion.
Wilson and Rouse (1972)	1.00	moderately dry soils
Davies (1972)	1.00	moderately dry soils
Denmead and McIlroy (1970)	1.00	moderately dry soils (wheat)

set $\alpha = 1.0$. Unless there are very dry areas in the basin this should provide a lower limit for the evaporation rate.

3.4.2 Evaporation at a site

To compute equilibrium evaporation using equation 3.13, S/(S+r) can be obtained as a function of air temperature (Dilley 1968) and Q* can be measured at representative sites. For the present study, air temperature was measured in the research basin, and net radiation was measured or estimated using regression relationships with solar radiation (Davies 1967). Evaporation was calculated as a function of equilibrium evaporation for Polar desert-lithosol surfaces and bog surfaces assuming both saturated (α=1.26) and moderately dry (α=1.0) conditions. Evaporation for the period June 15 to August 19 are summarized in table 3.4.

Table 3.4 Calculated evaporation from polar desert and bog surfaces.

Site	α	E (mm)
Polar desert-lithosols	1.26	130
	1.00	103
Bog	1.26	162

These are point estimates of evaporation, assuming that the site is snow-free for the entire period. These point estimates will be used to calculate average evaporation by weighting with the proportion of snow free areas in the basin.

3.4.3 Basin Evaporation

To calculate total seasonal evaporation for water balance studies, it is important to consider evaporation over all the dominant surface types existing in the basin, as well as to consider the effect of a variable snow cover. The rate of evaporation from snow surfaces is low. In Alaska, Weller and Holmgren (1974) found that evaporation from snow surfaces contributed to only 2 percent of the total ablation. They note that this value corresponded closely with other findings from Greenland and at Fairbanks, Alaska, where estimates varied from 1 to 6 percent. In view of its small magnitude, evaporation from snow surfaces will be ignored in the present study. Basin evaporation was then obtained at daily time intervals by weighting the site evaporation estimates with the proportion of snowfree areas in the basin. For basins 2,3 and 4, the areal extent of bog surfaces is limited and the evaporation rates for the polar desert-lithosol surfaces are considered to be representative of all the snowfree areas. For basin 1, however, the bog soils cover approximately 50 percent of its surfaces. Table 3.5 summarizes the computed results for the entire summer period.

Calculated evaporation from the four study basins.

Table 3.5

Basin	α	E (mm)
4	1.26	80
	1.00	63
3	1.26	79
	1.00	63
2	1.26	81
	1.00	64
1	1.26	82
	1.00	65

CHAPTER 4

WATER BALANCE DISCUSSION

4.1 Water Balance

In the previous chapter, the magnitude of the individual components of the water balance were described. In this chapter, the relationship between the components will be discussed. A comparison of the present findings with those from other water balance studies in the Arctic will enable some generalization upon the nature of the water balance in high Arctic basins.

4.1.1 Water Balance Calculations

Due to the lack of microclimatologic measurements of evaporation, total evaporation from the basin will be calculated from the water balance. This value will then be used as the control, against which evaporation calculated as a function of equilibrium evaporation will be compared.

Seasonal water balance for each basin is given by:

$$S_1 + S_2 + R - Q \pm \Delta S = E \tag{4.1}$$

where: S_1 = amount of snow measured by snow survey

S_2 = snowfall after snow survey

= AES snowfall × weighting factor

R = rainfall

E = evaporation

Q = stream discharge

ΔS = change in storage = 0 over a season.

The water balance for the period Sept. 1, 1975 to Aug. 20, 1976 is given for each basin in Table 4.1.

Table 4.1 Magnitude of the various components of the Water Balance/in mm for four Basins Sept. 1, 1975 to Aug. 20, 1976

Basin	S_1	S_2	R	Q	E
1	122	50	31	137	66
2	111	50	31	157	35
3	127	50	31	160	48
4	122	50	31	161	42

Calculations of this type allow a comparison of the relative importance of the various input and output components and their variation in importance between basins.

In the four study basins winter snowfall was the dominant input during the study year, accounting for approximately 60% of the total yearly precipitation. Snow which fell after the snow survey accounted for another 25%. Therefore 85% of the 1975-76 annual precipitation fell as snow, a value considerably higher than the mean of 58% for the Resolute weather station (Dept. of the Envir. 1972). Of this incoming precipitation, approximately 80% was removed from basins 2,3 and 4 by streamflow and 20% by evaporation. Due to its different physiographic characteristics, basin 1 released only 67% of its incoming precipitation as streamflow, the other 33% evaporated.

4.1.2 Comparison of the two estimates of evaporation

In the preceding sections, evaporation from the four study basins has been calculated in two different ways: by a form of the combination model, and as the residual of the water balance equation. Table 4.2 shows that these two estimates agree closely only for basin 1. For the other 3 basins, the estimates obtained from the combination equation were substantially higher.

Table 4.2 A comparison of evaporation determined by the water balance approach and by the equilibrium model (all values in mm).

Basin	Water Balance E	E eq
1	66	65
2	35	64
3	48	63
4	42	63

These differences may be due to any of the following factors: (1) a basic error in the form of the combination model being used, (2) use of the wrong α value, (3) error in the measurement of the water balance terms or (4) a major term in the water balance equation has been excluded.

Of these possibilities, the first is unimportant since a number of studies (see section 3.4.1) have shown that this form of the combination model is accurate if the correct value of α is used. The last possibility is also unimportant, since as mentioned earlier, no components which have a significant affect on the total water balance have been excluded from the analysis.

Although there is error involved in the measurements of the water balance components, the magnitude is considered to be insufficient to explain the large differences. The probable errors for each term of the water balance equation are given in Table 4.3.

Table 4.3 Probable and percentage errors in the determination of various components of the water balance

Basin	S_1			S_2			R			Q			E
	Value	% Error	Error	Value	% Error	Error	Value	% Error	Error	Value	% Error	Error	Value
1	122	15	18	50	15	8	31	12	4	137	11	15	66
2	111	9	10	50	9	5	31	12	4	157	11	17	35
3	127	12	15	50	12	6	31	12	4	160	11	18	48
4	122	11	13	50	11	6	31	12	4	161	11	18	42

Except percentage error, all values are in mm.

Of the four factors listed, the most likely cause of error is the choice of a wrong α value. In the combination model approach, α was set to 1.0. This was based on a number of studies (Rouse, Mills, Stewart 1978) which showed that if the supply of water was moderate or if the surface resistance was moderate, then an $\alpha = 1.0$ can produce an accurate estimate of evaporation. However, in the study basin or in the high Arctic where gravelly soils dominate, the value of α is likely to be much lower. For basin 1, both estimates are in agreement but are considerably higher than the evaporation estimate for the other basins. This is possibly attributed to the differences in the type of basin surfaces. In basins 2,3 and 4, a large part of these basins (approximately 80 to 90%) are covered with lithosols or polar desert soils and a high proportion (41%) of the three basins consists of ridge tops or crests. Both factors are expected to contribute to lower evaporation. On the other hand, approximately 50% of Basin 1 is covered by saturated, bog soils and a smaller percentage of its area is covered by ridge tops. Basin 1 is expected to evaporate at close to the potential rate ($\alpha = 1.26$) and Basins 2,3 and 4 at some unknown, but lower rate. The effect of terrain type on basin evaporation can be seen if a rough water balance of the crest regions is carried out to obtain an indication of the amount of water available for evaporation. For the crests located in Basin 4, the following data are available:

(1) Snow storage = 45 mm

(2) Summer precipitation = 50 mm

(3) Total precipitation = 95 mm

(4) Evaporation - the crests became free of snow on June 25. From then until Aug. 19, the following amounts of evaporation were calculated

$\alpha = 1.26$ $E = 108$ mm

$\alpha = 1.00$ $E = 86$ mm

$\alpha = 0.50$ $E = 43$ mm

Part of this precipitation must runoff as overland flow during the initial melt period and as saturated and non-saturated flow in the active layer throughout the summer. During the snow-melt period, it is estimated that 90% of the meltwater runs off as overland and subsurface flow. During the summer however, a thick active layer provides storage and therefore only a small percentage of rainfall will runoff, say 30%. Applying these estimates to the above precipitation amounts, only 39 mm are available for evaporation, a value which is considerably below

the estimate of evaporation if an $\alpha = 1.0$ is used. Another possible factor limiting the evaporation rate is the stoniness of the soil, a condition prevalent in the basins. The effect of this type of material on evaporation has not as yet been studied. If the bulk of evaporation occurs very close to the surface, then the large number of stones on the surface would lower the evaporation rate.

Concerning the two estimates of evaporation, several conclusions can be drawn:

(1) the water balance estimates of evaporation are the most accurate

(2) the equilibrium estimate of evaporation ($\alpha = 1.0$) is a good approximation of evaporation in basin 1. This is due to the large percentage of the basin which is covered with a saturated surface.

(3) equilibrium evaporation ($\alpha = 1.0$) overestimates evaporation in Basins 2,3 and 4. This is due to the stoniness of the soil and the large percentage of the basin occupying hill crests which have a low water supply.

4.1.3 Calculation of a mean α value

In the preceding section, it has been shown that evaporation is overestimated if a seasonal average of $\alpha = 1.0$ is used. To overcome this problem, it is possible to calculate a seasonal average α value using water balance and heat balance data. To determine an average α value for the basin, equation 4.10 will be used:

$$E = P - R \tag{4.2}$$

$$T = H + LE \tag{4.3}$$

$$H = T - LE \tag{4.4}$$

$$T \approx Q^* \quad \text{from Thom (1975)} \tag{4.5}$$

$$H = Q^* - LE \quad \text{combining (4.4) \& (4.5)} \tag{4.6}$$

$$B = H/LE \tag{4.7}$$

$$LE = \alpha \frac{S}{S+r} \quad Q^* \tag{4.8}$$

$$LE = T/(1+B) \qquad \text{combining (4.3) \& (4.7)} \tag{4.9}$$

$$\alpha = [(\frac{S}{S+r})\cdot(1+B)]^{-1} \quad \text{combining (4.8), (4.9) and (4.5)} \tag{4.10}$$

where: E = evaporation

L = latent heat

P = precipitation

R = runoff

H = sensible heat flux

Q^* = net radiation

B = Bowen Ratio

T = Total heat flux

The data required are evaporation calculated from the water balance, mean air temperature to allow the calculation of a seasonal value of S, and net radiation which is already weighted daily by daily estimates of the snowfree proportion of the basin. The areal weighting ensures that the snow covered areas are not included in the evaporation calculation.

The results of this analysis are shown in Table 4.4.

If these values of α are used to calculate evaporation, the resulting seasonal evaporation will obviously be equal to evaporation determined from the water balance. It must be remembered however, that these α values are averages for the entire basin and over a period of two months. The basin itself

Table 4.4 Calculation of Mean seasonal α values

Basin	Total seasonal Q^* over Snowfree areas (J)	E(mm)	LE(J)	H(J)	$B=\frac{H}{LE}$	α
1	35102	66	16391	18707	1.14	1.07
2	34834	35	8629	26209	3.04	0.57
3	34022	48	12017	22048	1.83	0.81
4	34395	42	10471	23923	2.28	0.70

Note: (1) $\frac{S}{S+r}$ was calculated in two ways. The mean bi-hourly value and from the mean bi-hourly air-temperature. Both results were nearly identical.

(2) The latent heat of vaporization was calculated the same as S/S+r above. Both results were identical.

is composed of different terrain units each of which has different values of α and each value of α will vary over time. Therefore, calculations of evaporation using these α values can be used only for an entire season. Another limitation is that these values are based on the data from one year. The seasonal average value of α is expected to vary from year to year, depending on the moisture conditions of the soil. Therefore these values can not be expected to provide accurate estimates of evaporation in years when the soil moisture conditions (dependent on precipitation) are different.

However, the above estimates provide a first estimate of the value of α in the high Arctic. Since the precipitation of 1976 was only slightly below normal, these α values are likely to be representative of the long term mean.

4.1.4 Short term changes in α

One major problem in short-term estimation of evaporation is to determine the decreasing value of α as the soil dries out. A number of studies have related α to soil moisture. Priestly and Taylor (1972) showed curves for three different soils, each exhibiting the same basic shape. As the soil dried from saturation, the value of α remained close to 1.26 until a specific limit was reached. Below that limit, α declined linearly until evaporation ceased. Davies and Allen (1973) show a non-linear drying curve which is similar to those shown by Priestly and Taylor (1972). Rouse, Mills and Stewart (1978) show the relationship between α and soil moisture for three different surfaces, postulating a sudden drop in α as the old and the new burn soils dry out. For a lichen surface, a constant α is used for all surface moisture conditions.

A method such as that used above is useful for allowing hourly calculations of evaporation. The major disadvantage with the method is the need of soil moisture data. A simplified alternative is to relate the changes in α with the number of days since a rainstorm (fig. 4.1).

To study the effects of varying soil moisture conditions on evaporation, the field program was extended to the summer of 1977. All field measurements were the same as in 1976. In addition, 9 lysimeters were installed to allow evaporation to be measured directly. These were installed in early July and were weighed daily until Mid-August. Five were placed in dry polar desert soils typical of hill crests. Two were placed in wet polar desert soils typical of lower lying flats, slopes and high flats. The other two were located in wet vegetated soils. Soil moisture was taken daily at each lysimeter site. To allow an estimation of the actual value of α, the lysimeter evaporation was compared with equilibrium evaporation.

The values of α determined by this method fall into three distinct groups. The wet soils consistently had values between 1.0 and 1.3, the dry soils between .80 and 1.33 immediately after a rain and between .07 and .4 after an extended drying period. Since α is dependent on the soil moisture char-

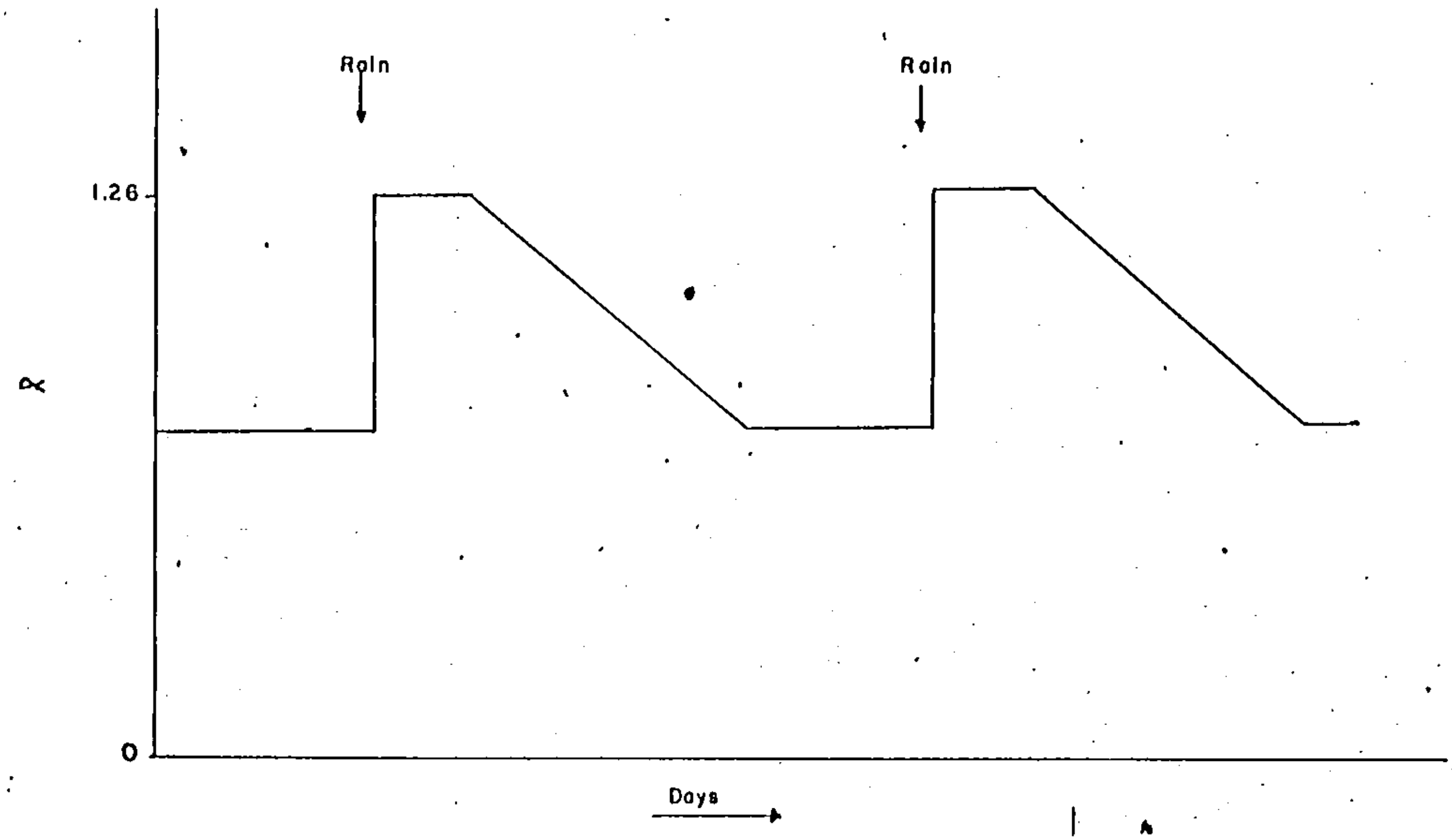


Fig. 4.1
General relationship between α and number of days since a rainfall

acteristics of the soil, it is appropriate to plot α against soil moisture. The results are shown in fig. 4.2. Even though there is a fairly wide scatter of points and a distinct lack of data for moisture values between .14 and .18, (moisture content to dry soil ratio by weight) the relationship is similar to those reported elsewhere. When the soil moisture exceeds .18, the mean value of α remains constant at 1.20.

Below a soil moisture level of 0.11, α assumes a minimum value of 0.12 and this value remains at least until the soil moisture reaches .04.

The drying curve presented above is very similar in shape to that presented by Rouse et al (1978). However, the values defining the lower limits of the curves are considerably different. Rouse et alia(1978) found this to vary between .91 and .97 depending on the surface type, a value considerably greater than that determined in this study,

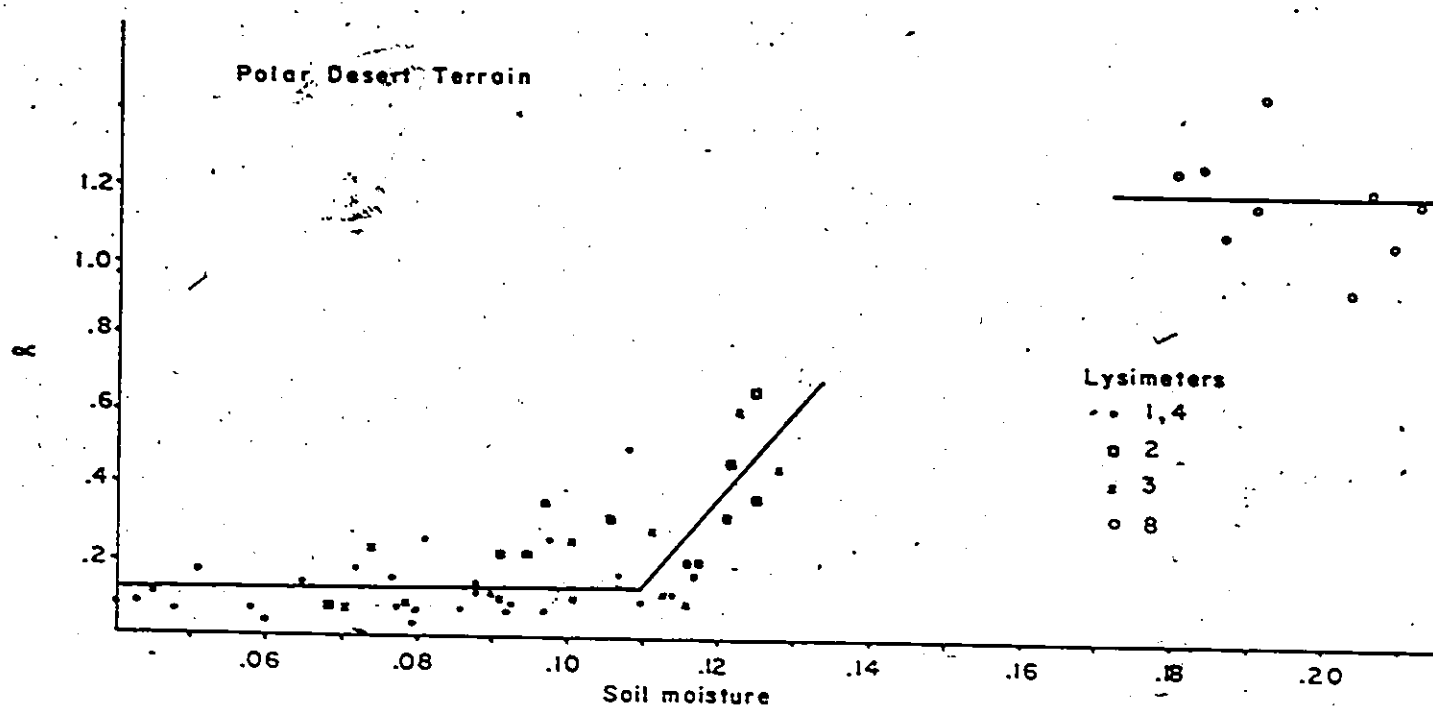


Fig. 4.2
Relationship between α and soil moisture as determined from lysimeters used in 1977.

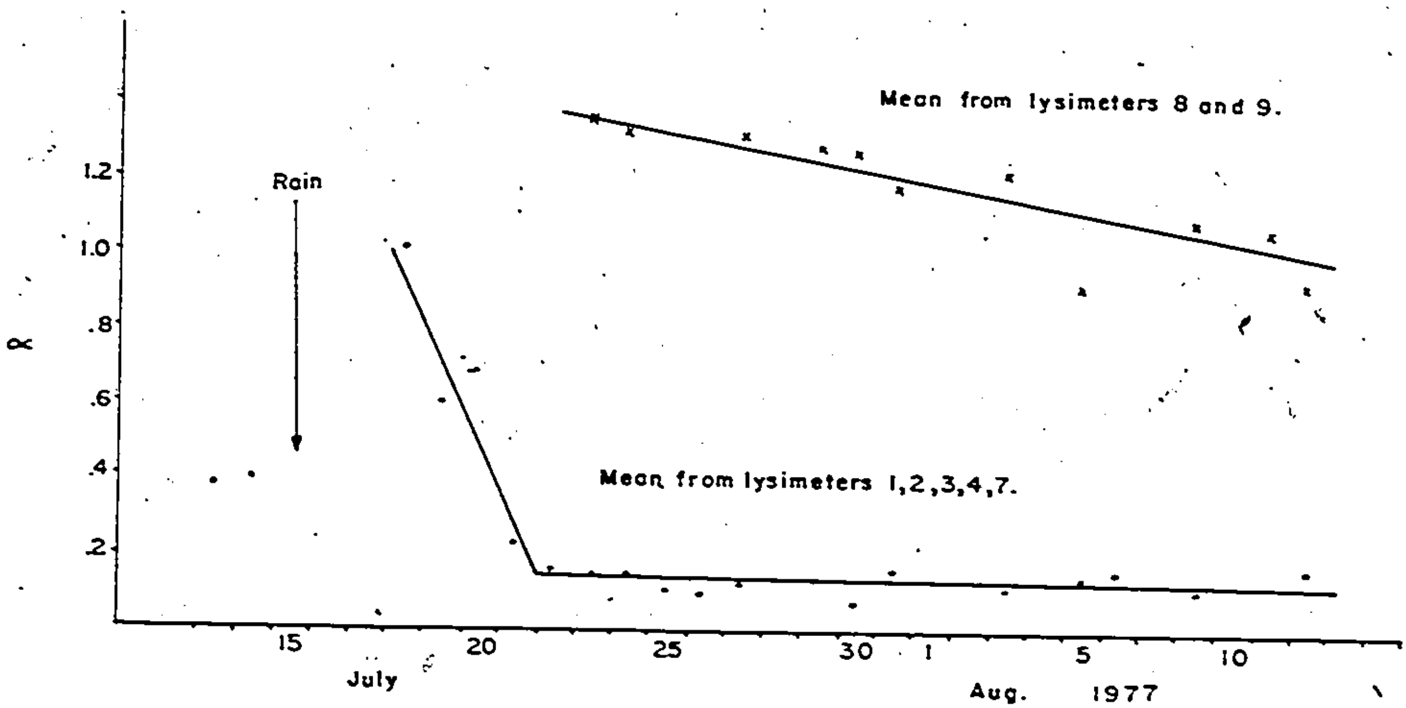


Fig. 4.3
Changes in α after a rainstorm. Determined from lysimeters in 1977.

An important factor in determining the limit of α is the formation of a thin layer of very dry soil at the surface. This results in a strong resistance to the upward movement of water from the wetter soil below. And since these soils are non-vegetated, no transpiration occurs. In contrast to this, the sub-arctic soils studied by Rouse et al are vegetated and would not form such a resistant layer.

A drying curve which plots α against time since wetting is more useful than the α vs. soil moisture relationship shown above. This is due to the fact that soil moisture data are difficult to obtain and therefore a model using them as an input parameter is not conducive to routine operation. Using two groups of lysimeter data, daily changes in α after the July 16, 17 rainstorm were plotted. As fig. 4.3 shows, the rates of drying are different. At the dry sites, α rose to a value of 1.0 soon after the storm, but then quickly fell to a minimum value of .12 four days later. Due to the fact that no other precipitation event was sufficiently large to wet the soil, α remained at this low value for the remainder of the summer. In contrast to this, the wet gravels site dried slowly due to the flow of groundwater at the low lying site. Therefore α declined very slowly and in fact still retained a moderately high value at the end of summer. An interesting feature of both drying curves is that the declining portion is linear, confirming the shape of the time dependent drying curve proposed earlier.

It is highly probable that the shape of the drying curve is affected by storm size and by the initial moisture condition of the soil, but the scarcity of summer storms in 1977 offered a limited range of moisture conditions. In the present study, therefore only a single drying curve can be used. The necessary variables which are known are:

(1) $\alpha = 1.26$ when evaporation is occuring at the potential rate and

(2) $\alpha = 0.10$ when evaporation is occuring at its lowest rate due to low soil moisture

A medium size storm (2.5 mm) was taken as the storm size necessary to raise α to 1.26. After the storm, α was allowed to remain at this value for three days, after which it was allowed to decline to its lowest value for twenty days. The choice of twenty days is based on the depletion of moisture after a mid July storm of 1976. Since moisture conditions are known to vary within the basin, a further improvement in the calculation was made by dividing the basin into separate units, each with a different minimum value of α. The values chosen were: (1) 0.10 for crests, (2) 1.26 for low flats, because they correspond closely to the low wet regions which do not dry out during the summer, (3) 0.9 for gullies, slopes, high flats, valleys because these regions stay moderately wet and so a value similar to moderately wet, subarctic surfaces (Rouse et al 1978) was used. Table 4.5 shows the results of this computation of evaporation. Seasonal evaporation was calculated for each terrain unit. This value was then weighted by the proportion of basin area to obtain a value for total basin evaporation. As shown in Table 4.5, the calculated values agree quite well with the water balance evaporation for basins 3 and 4 but not for 2.

Table 4.5 Estimation of basin evaporation taking into consideration various terrain types

Basin	Crest		Gullies		Slopes		High Flats		Low Flats		Basin	Water Balance
	E(mm)	Ex P.A.	E	Ex P.A.	E	Ex P.A.	E	Ex P.A.	E	Ex P.A.	E	E
2	42.2	16.1	17.4	0.25	51.5	19.6	53.4	5.7	63.5	5.6	47.3	35
3	42.2	17.9	22.2	1.4	49.6	17.4	57.7	4.7	66.6	4.2	45.6	48
4	41.9	17.5	8.3	0.4	49.6	17.4	53.4	4.6	66.6	4.8	44.3	42

E = evaporation in mm

P.A. = proportion of total basin area occupied by a given terrain type

4.1.5 Water Balance of Terrain Units

Available data on the snow storage, rainfall and evaporation for each terrain unit, allows the discharge from each terrain unit to be calculated as the residual of the following water balance equation.

$$Q = (S_1 + S_2 + R) - E \tag{4.11}$$

where the terms are as defined earlier.

This method allows the spatial variability of the water balance components to be shown. Then it is possible to infer upon the areas which supply most of the water to the streams.

Two important aspects of the hydrologic system are demonstrated in Table 4.6. Firstly, the input and output components of the water balance are highly variable over the study basin. This is due primarily to the variations in snow input, but also to variations in evaporation. The percentage of incoming precipitation that leaves the terrain unit as streamflow varies from a low of 53% for crests to a high of 98 and 100% for gullies and valleys respectively. The second important aspect shown in Table 4.6 concerns the percentage contribution of each terrain type to total basin streamflow. As can be seen in the table, the percentage contribution varies from a low of 5% for low flats to a high of 51% for slopes. These numbers become even more striking if the area of each terrain unit is considered. For example, crests occupy 42% of the basin but contributes only 14% of the streamflow. Slopes on the other hand occupy only 35% of the basin area, yet produce 51% of the total streamflow. This observation shows the importance of the variable source area of streamflow on a seasonal scale, and gives an indication of the magnitude of the variations to be expected.

Table 4.6: Components of terrain unit water balance for Basin 4

Basin	Area (% of total)	Terrain unit	S_1 mm	S_2 mm	R mm	E mm	Q mm	% of total precip	Q weighted mm	% of total basin discharge
4	42	crest	45	18	31	42	52	55	22	14
	5	gullies	264	105	31	8	392	98	20	13
	35	slopes	176	70	31	50	227	82	79	51
	9	h.flats	119	47	31	63	144	73	13	8
	7	l.flats	108	43	31	67	115	63	8	5
	2	valleys	496	197	31	0	724	100	14	9

4.2 Comparison with Previous Studies

With an increasing amount of hydrologic information it is pertinent to compare the results of the present study with those of other workers in Arctic hydrology. Water balance results may be compared in terms of the actual magnitude or the relative importance of the individual components. Table 4.7 lists the results of all comprehensive water balance studies carried out in high Arctic nival basins and a representative sample of similar studies carried out in lower latitudes (both vegetated tundra and wooded regions of the subarctic). Table 4.8 shows the results of a number of energy balance studies carried out in the high Arctic.

Precipitation controls the amount of water available to the nival-regime hydrologic system. Two precipitation characteristics are important in the Arctic: its absolute magnitude and the proportion which falls as snow. Since 1954, when Black determined that precipitation at Barrow, Alaska was 2 to 4 times the measured, it has been well known (Church 1974) that precipitation throughout the Arctic was underestimated. Two studies in the sub arctic (Adams et al 1966, Findlay 1966) have shown that between 25 and 37% of the total snowfall was not caught by official gauges. Two recent studies in the high Arctic (Holecek and Vosahlo 1975, Wedel 1977) have reported snowfalls between 92 and 120 mm. These values agree closely with the amount of snow measured by the present study (120 mm). Accepting this as the correct snowfall value, the Atmospheric

Table 4.7: Summary of Previous Arctic water balance studies

Study	Precipitation Snow	Precipitation Rain	Discharge	Evaporation	Sublimation	Location	Surface Type
Wedel (1977)	135(88%)	19(12%)	156 (99%)	2(1%)	-	Bathurst Island	non-vegetated
Holecek & Vosahlo (1975)	92(66%)	48(34%)	23 (17%)	90(64%)	27(19%)	Devon Island	lowland-sedge meadows
	174(Total)		116 (66%)	24(14%)	35(20%)	Devon Island	upland-barren plateau
Cogley (1975)	440(Total)		180 (41%)	241(55%)	-	Cornwallis Is.	non-vegetated
Church (1974)	616(Total)		462 (75%)	154(25%)	-	Baffin Island	non-vegetated, bedrock and shallow drift
Walker et al (1973)	143(Total)		146 (100%)	≈0 (0%)	-	Ellesmere Island	non-vegetated
present study	172(85%)	31(15%)	137 (67%)	66(33%)	-	Cornwallis Is.	50% vegetated
	161(84%)	31(10%)	157 (82%)	35(18%)	-	Cornwallis Is.	non-vegetated
	177(85%)	31(15%)	160 (77%)	48(23%)	-	Cornwallis Is.	non-vegetated
	172(85%)	31(15%)	161 (79%)	42(21%)	-	Cornwallis Is.	non-vegetated
Kane and Carlson (1973)	54(57%)	40(43%)	75 (80%)	27(29%)	-	Prudhoe bay, Alaska	muskeg, and heather, tundra

Study	Precipitation Snow	Rain	Discharge	Evaporation	Sublimation	Location	Surface Type
Brown et al (1968)	180 (Total)		90(50%)	90(50%)	-	Pt.Barrow Alaska	marshy tundra, many ponds
Findlay (1966)	915 (Total)		635(69%)	280(31%)	-	Knob Lake Quebec	spruce-dwarf birch lichen woodland
Dingman (1971)	313 (Total)		150(48%)	150(48%)	-	Glen Cr. Alaska	spruce-moss and birch-moss woodland
Anderson (1974)	285 (Total)		210(74%)	75(26%)	-	Boot Cr., Inuvik	spruce-birch lichen heath upland

Table 4.8: Summary of High Arctic Energy Balance Studies

Author	Location	Surface Material	LE(%)	H(%)	G(%)	B
Smith(1976)	Eureka, Ellesmere Island	Dry-vegetated	43	39	11	.97
		Wet-impeded drainage	57	33	11	.56
Addison(1975)	King Christian Island	Dry meadow (after rain)	39	55	6	
		Dry meadow (after a dry period)	10	75	15	
		Moist meadow	41	55	4	
Addison(1972)	Devon Island	Beach ridge Dryas Integ.	12	77	11	6.59
		Lichens	70	29	1	.40
		Complete	54	46	.8	.86
		Meadow	31	64	4	2.05
Weller and Holmgren(1974)	Barrow, Alaska	mid-summer	66	32	2	

Author	Location	Surface Material	LE(%)	H(%)	G(%)	B
Cogley(1975)	Cornwallis Island	polar desert				
		June	39	46	17	
		July	49	37	15	
		Aug.	36	47	18	
Present Study	Cornwallis Island	polar desert	25	75 (H+G)		
		bogs	53	47 (H+G)		

Environment Service records have underestimated snowfall by over 50%. Using this underestimation as a weighting factor, actual mean snowfall at Resolute would be 156 mm rather than the 78 mm reported by the government weather station.

Rainfall in the Arctic is considered to be underestimated, but the amount of underestimation is less than for snowfall. Black (1954), Cook (1960) and Cogley (1975) believed that trace events were responsible for large errors in measurement. On the other hand, studies by Wedel (1977) and Holecek and Vosahlo (1975) indicate that actual rainfall is very similar to that measured by the weather station at Resolute. Despite spatial variability of rainfall, the weather station data was a good approximation of actual basin rainfall when the data are summed over the summer season. Since rainfall is not likely to be greatly underestimated, the total error involved in estimating northern precipitation is mainly attributed to the measurement of snowfall.

In light of the above analysis, snowfall plays an important role in its contribution to total precipitation. Wedel (1977) and the present study both showed that for the 1975-76 study year, snowfall constituted 84 to 88% of total precipitation. Holecek and Vosahlo (1975) used the data for several years from Truelove Inlet, Devon Island and estimated that the long term mean precipitation consisted of 66% snow. Applying 1975-76 correction factors to the long term mean snowfall of Resolute, the new snowfall estimate of 156 mm represents 73% of total precipitation. This can be compared with a previous estimate of 57%.

The streamflow data outlined in Table 4.7 shows that most Arctic streams have an annual runoff in the 115 to 180 mm range. Of the studies listed, only Church's Baffin Island study and Holecek's sedge meadow basin, fall outside the range. The Baffin Island study shows a much large discharge in response to higher precipitation induced by a more maritime climate. The extremely low discharge reported by Holecek and Vosahlo is due to the low lying basin topography causing impeded drainage and therefore large amounts of water storage. In most of the basins, discharge exceeds 66% of total basin precipitation. The only exceptions being Holocek and Vosahlo's sedge meadow and Cogley (1975). The reason for the low percentage in the sedge meadow is given above, while the low value given by Cogley is due to an extremely high computed evaporation rate.

Evaporation studies can be divided into two groups, those concerned with site specific energy balance and those regarding basin wide evaporation as a part of the water balance.

The energy balance studies listed in Table 4.8 show a considerable range in the proportion of the total energy used for evaporation. Addison (1972, 1975) found that evaporation was dependent on vegetation and that it could vary from 10 to 70% of the total. One of Addison's sites, the dry meadow, was similar to the study basin. He found that LE accounted for only 10% of the total energy during a dry period, but 39% immediately after a rainfall. These values are similar to the basin values obtained in section 4.1.3 where 25% of the energy went to LE for basins primarily covered with polar desert and lithosol soils. Values for a wet site ranged from 41% (Addison 1975) and 57% (Smith 1976) to 66% (Weller and Holmgren 1974). These values agree closely with the value of 53% found for a basin covered by 50% vegetation (wet) in the present study.

Three other studies (Wedel 1977, Holecek and Vosahlo 1975 and Cogley 1975) have provided estimates of total basin evaporation in high Arctic basins. Wedel calculated evaporation as the residual of the water balance, Holecek and Vosahlo used lysimeters, while Cogley calculated evaporation from the combination model.

Wedel found that evaporation was extremely small, while Holecek and Vosahlo found a large variation in evaporation over different surfaces. They found a value of 90 mm over sedge meadows and 24 mm over barren plateaus. Cogley reported an evaporation value of 241 mm, a value greater than most results reported for subarctic regions. The results of Holecek and Vosahlo are in close agreement with those in the present study. Their result for a sedge meadow (90 mm) is comparable with the 66 mm found for a basin covered by 50% wet vegetation. And their value of 24 mm for the barren upland is comparable with the 42 mm found for non-vegetated basins in the present study. In light of the agreement between these two studies, it is generalized that for the high Arctic: (1) where water is abundant, evaporation ranges between 70-90 mm accounting for

35-65% of total precipitation, (2) where water is limiting (i.e. most of the non-vegetated high Arctic) evaporation is in the 20-35 mm range, accounting for only 14-25% of total precipitation.

A comparison of the water balance of high Arctic basins with those of the subarctic regions shows that the dominant feature is the amount of water in a basin. Although precipitation in the high Arctic is greater than normally reported, the high Arctic is basically a dry region. An annual precipitation of 150 to 200 mm is considerably less than that reported by Findlay (1966), Dingman (1971) and Anderson (1974) for more southerly latitudes. The result of low precipitation is a correspondingly low discharge and evaporation. Only in isolated regions of saturated ground, are evaporation rates close to those reported further south. However, due to the greater importance of snowfall, the bulk of discharge is released within a short, snowmelt period. Since over 70% of total annual flow occurs within 10 to 15 days and because of the lack of basin storage, little water remains in the basin for evaporation during the summer period. In comparison with subarctic basins, discharge in high Arctic accounts for a greater percentage of total basin precipitation.

CHAPTER 5

CONCLUSIONS

Previous works in the high Arctic have provided a limited amount of data on the drainage systems, but the magnitude and the importance of several components of basin water balance remain unknown. It was therefore the objective of this dissertation to study all components of the water balance of a small high Arctic basin and to compare the results to previous work to enable a general description of the high Arctic water balance.

Underestimation of basin snow storage has long been a major problem in the study of northern hydrologic systems. A late winter snow survey near Resolute confirmed this underestimation and determined that the Atmospheric Environment Service data was approximately 50% of the total basin snow storage. Using the survey information, it was possible to adjust the mean snowfall at Resolute from 78 mm to 156 mm. Previous studies have indicated that rainfall may also be grossly underestimated, but the present study shows that over an entire summer period, weather station data is representative of the basin. However, for individual storms, spatial variation in rainfall can be considerable, and is dependent on wind direction. This analysis indicates that total error in precipitation measurement is mainly related to snowfall measurement, and that near Resolute, snowfall can account for 73% of total precipitation.

Streamflow accounts for a large proportion of basin water losses, varying from 67 to 82%. Basins without significant storage capacities consistently release 80% of their precipitation by surface runoff. Only a small basin with impeded drainage had a lower value (67%). The importance of the variable source area concept was demonstrated in the Arctic basin. Using the streamflow records it was evident that further up the basin, the first date of flow was progressively delayed, implying a continual expansion of the basin area which supplies water to the stream. In addition, only a portion of this potential contributing area actually contributed water to streamflow depending on the areal extent of the basin snow cover, and the presence of flow connections between different parts of the basin. Variations in the actual contributing area is inferred by the ratio between the discharges at two points on the stream system. It was found that initially the contributing area increased in size at the same rate, regardless of basin size. Proportionally, therefore, the smaller basin contributes more water than expected during the early melt stages. As melt progresses, discharge ratios become similar to the basin area ratios, suggesting more extensive contributing areas for larger basins. The variable source area was also demonstrated by the water balance of various terrain units, each of which contributes a different amount of water to streamflow.

Evaporation varied greatly within the study areas. Evaporation was 40 mm for most basins, but 66 mm was evaporated from a small basin with impeded drainage. These values correspond to 21 and 33% of total precipitation. It was found that for the small basin, equilibrium evaporation ($\alpha = 1.00$) gives a good estimate of basin evaporation, but for the larger basins, equilibrium evaporation overestimates actual evaporation. Water balance and heat balance calculations indicate that a mean α of 0.60 can be expected for the larger basins, indicating that evaporation over most of the study area is limited by the availability of water. Lysimeter studies showed that α is dependent on soil moisture conditions. When the soil is wet after a rainfall, $\alpha = 1.26$, but as the soil dries, α decreases to a minimum value of 0.10.

In general, the water balance of a high Arctic basin can be summarized as follows. 70% of the incoming precipitation is snow, the remaining 30% rain. Over most of the study basin 80% of this water leaves the basin as streamflow and 20% as evaporation. However, for small areas that have impeded drainage, these values may be 67% and 33% respectively. These are basin wide values. Within each basin the importance of each output component varies greatly. Results of the water balance study are compared with similar studies carried out in subarctic regions. It was found that precipitation is generally lighter in the high Arctic, and that the percentage contribution of snowfall is more important. Streamflow accounts for a larger proportion of total outflow than in more southerly regions, leaving less water available for evaporation.

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