WATER BALANCE OF A SMALL LAKE
IN THE HIGH ARCTIC

WATER BALANCE OF A SMALL LAKE IN THE HIGH ARCTIC

by

PETER J. STEER

A 4C6 Research Paper
Submitted to the Department of Geography
in Partial Fulfillment of the Requirements
for the Degree
Bachelor of Arts

McMaster University

April 1979

BACHELOR OF ARTS (1979)
(Geography)

McMASTER UNIVERSITY
Hamilton, Ontario

TITLE: Water Balance of a Small Lake in the High Arctic

AUTHOR: Peter James Steer

SUPERVISOR: Dr. M.K. Woo

NUMBER OF PAGES: viii 94

ABSTRACT:

In 1978, the water balance of a small lake near Resolute, N.W.T. was studied. Using measured water inputs and water outputs, the change in storage term was calculated. A positive net change in storage was partitioned between storage in the active layer and storage in the lake.

A comparison of the magnitudes of the various components of the water balance equation shows that i) for the snow-dammed lake, outflow is most important for the few days following the breakup of the channel, ii) evaporation is an important process, removing almost as much water as summer precipitation received by the basin, and iii) depending on the condition of the active layer during freeze-up, considerable amounts of water may be held in storage at the end of summer.

ACKNOWLEDGEMENTS

I wish to express my appreciation to my supervisor, Dr. M.K. Woo, for his guidance and encouragement during the course of this study. For the loan of equipment, I wish to thank Drs. J.A. Davies, F.G. Hannell, and W.R. Rouse, and Glaciology Division, Department of Environment. Thanks also to Richard Heron, Paul Tice, and Philip Marsh for their companionship and assistance in the field.

This study was supported by grants from the Water Resources Research Support Programme, Department of Fisheries and the Environment; and the McMaster University Presidential Committee for Northern Studies. Generous logistical support was received from Polar Continental Shelf Project, Department of Energy, Mines and Resources, and Glaciology Division, Department of Fisheries and the Environment.

TABLE OF CONTENTS

LIST OF FIGURES

LIST OF TABLES

CHAPTER 1

INTRODUCTION

1.1 Review of Literature

Various authors have noted inadequacies of hydrologic information in the Canadian Arctic (Ambler, 1974; Church, 1974) although much work has been done in Arctic Alaska and the Soviet Northlands (Mackay and Løken, 1974).

Inconsistencies in water balances for northern North America have been attributed to an underestimation of basin snow storage (Woo and Marsh, 1978) and the lack of trace precipitation measurements which can contribute significant amounts during some summers (Woo and Steer, 1979).

From studies conducted in a vegetated basin in Ellesmere Island, Woo (1976) concluded that evaporation is an important component of the water balance of small Arctic drainage basins with tundra vegetation. This conclusion was confirmed by a water balance study of a small pond in the same area (Marsh and Woo, 1977). Brown et al. (1961) also found that summer precipitation approximately equalled evaporation in a small drainage basin near Barrow, Alaska. In unvegetated areas, however, evaporation is likely to be less significant (Marsh, 1978).

Hartman and Carlson (1973) studied a small lake in a permafrost region near Fairbanks, Alaska. Their results indicated little or no hydraulic connection between the lake and a groundwater system. The lake routinely fills with water from spring runoff, after which time the level throughout the remainder of the season is essentially controlled by rainfall and evaporation. For lakes without surface channel inlets or outlets, the main hydrologic inputs are direct precipitation and overland flow occurring during the spring-melt period (Kane and Carlson, 1973).

All the above studies suggest a lack of complete agreement between authors regarding the role played by various components of the water balance. This is probably due to inadequate hydrologic data scattered over a wide range of environmental conditions.

1.2 Objective of the Study

To date few comprehensive attempts have been carried out to examine all the major components of basin water balance in a High Arctic environment. In particular, little attention has been paid to basins containing lakes, despite the abundance of lakes in the Canadian Arctic. The present study attempts to provide hydrologic information in these respects.

The objective of this study is, therefore, to determine the daily water balance of a basin containing a small lake in the High Arctic during the break-up and ice-free periods.

1.3 Presentation of Thesis

The water balance approach may be summarized as follows:

$$I - O = \Delta S \qquad (1.1)$$

where I is the input of water to a given area during a given time period; O is the output of water from the same area during the same time period; and ΔS is the change in storage in the same area during the specified time period (Gray, 1970).

Expanding and rearranging terms yields the following equation for an Arctic basin with a small lake:

$$P_S + P_R + P_T = E + Q + \Delta S_L + \Delta S_G \qquad (1.2)$$

where P_S, P_R, and P_T are snowfall, rainfall, and trace precipitation respectively, E is evaporation, Q is outflow from the basin, and ΔS_L and ΔS_G are lake storage and groundwater storage respectively. Groundwater storage is difficult to measure so will be calculated as a residual of equation (1.2).

The water balance equation for the lake itself is as follows:

$$P_L + Q_O + R_S = E_L + Q + \Delta S_L \qquad (1.3)$$

where P_L is precipitation falling directly onto the lake, Q_O is overland flow and channelled flow, R_S is subsurface flow, E_L is evaporation from the lake surface, Q is outflow from the lake, and ΔS_L is as previously defined.

Chapter Two of this thesis describes the study area and outlines the methods used for data collection. Chapter Three examines the water inputs to the basin and to the lake; the terms on the left-hand side of equation (1.2) and (1.3). Chapter Four treats water outputs and changes in storage. Because the discharge term (Q) is identical for both equations (ie. outflow from the lake is the same as outflow from the basin), it is analyzed only once in section 4.1.1. The fifth chapter presents an analysis of the water balance for Three Mile Lake. A separate discussion is presented for the entire basin and for the lake itself. How the input and output components discussed in the previous two chapters interrelate is examined and the water balance is computed. A summary of findings is presented in Chapter Six.

The name Three Mile Lake is not an official place name.

CHAPTER 2

STUDY AREA AND FIELD METHOD

2.1 Study Area

Between June and the beginning of September 1978, field work was carried out in a small coastal basin 5 km. northwest of Resolute Airport, Cornwallis Island, N.W.T. (Figure 2.1). The basin is occupied by a lake and has an area of 2 km^2. The surface area of the lake averages 192,700 m^2.

Physiographically, the area lies at the southern edge of the Innuition Region (Bostock, 1970) although the local relief is more typical of the areas classified as the Arctic Lowlands. Topographically, most areas in the basin are gently undulating, but the eastern sector is dominated by a steep slope which rises to a plateau. Elevation ranges from 5 m. above sea level to approximately 90 m.

The basin is underlain by permafrost occurring at depths of 0.3 - 0.8 m. below the surface. By far the largest proportion of the basin ground surface is gravel, with rock outcrops occurring on the eastern slope. Some depressions and hollows on the slope are occupied by permanent snow patches. Below these snow patches, meltwater sustains bog conditions throughout the summer. In addition, a depression

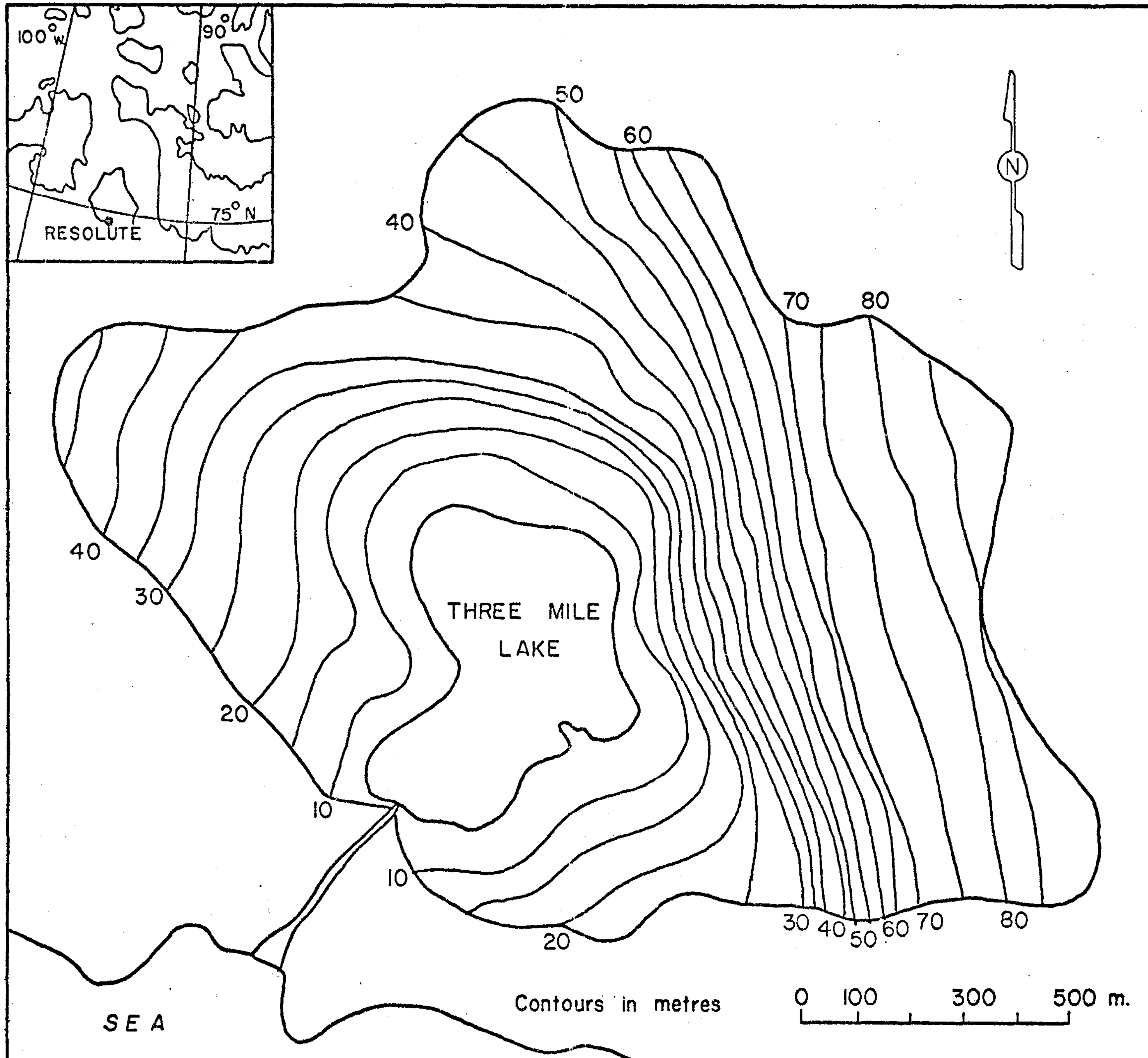


Figure 2.1. The study area on Cornwallis Island, N.W.T.

near the western boundry of the basin contains a small pond approximately 0.25 - 0.3 m. deep.

On the whole, the surface cover of the basin is as follows: 74.58 percent gravel; 13.29 percent bogs and ponds; 9.46 percent lake; and 2.67 percent permanent snow patches (Figure 2.2).

2.2 Field Method

2.2.1 Meteorological Data

Precipitation

Snow storage was determined by a late-winter (early June) snow survey conducted before the melt season. A total of nine transects were made across the basin. Along each transect, 15 to 40 snow depths were taken at regular intervals. For each transect at least four snow densities were determined using a Meteorological Services of Canada snow sampler.

Precipitation was collected from mid-July until the end of August using fifteen manual gauges with an orifice diameter of 145 mm. and at a height of approximately 25 cm. above the ground (Figure 2.3). These gauges were checked soon after each precipitation event so as to minimize evaporation of the catch.

A Weather-Measure tipping bucket rain gauge was attached to a Weather-Measure event recorder and placed adjacent to a manual gauge for comparative purposes.

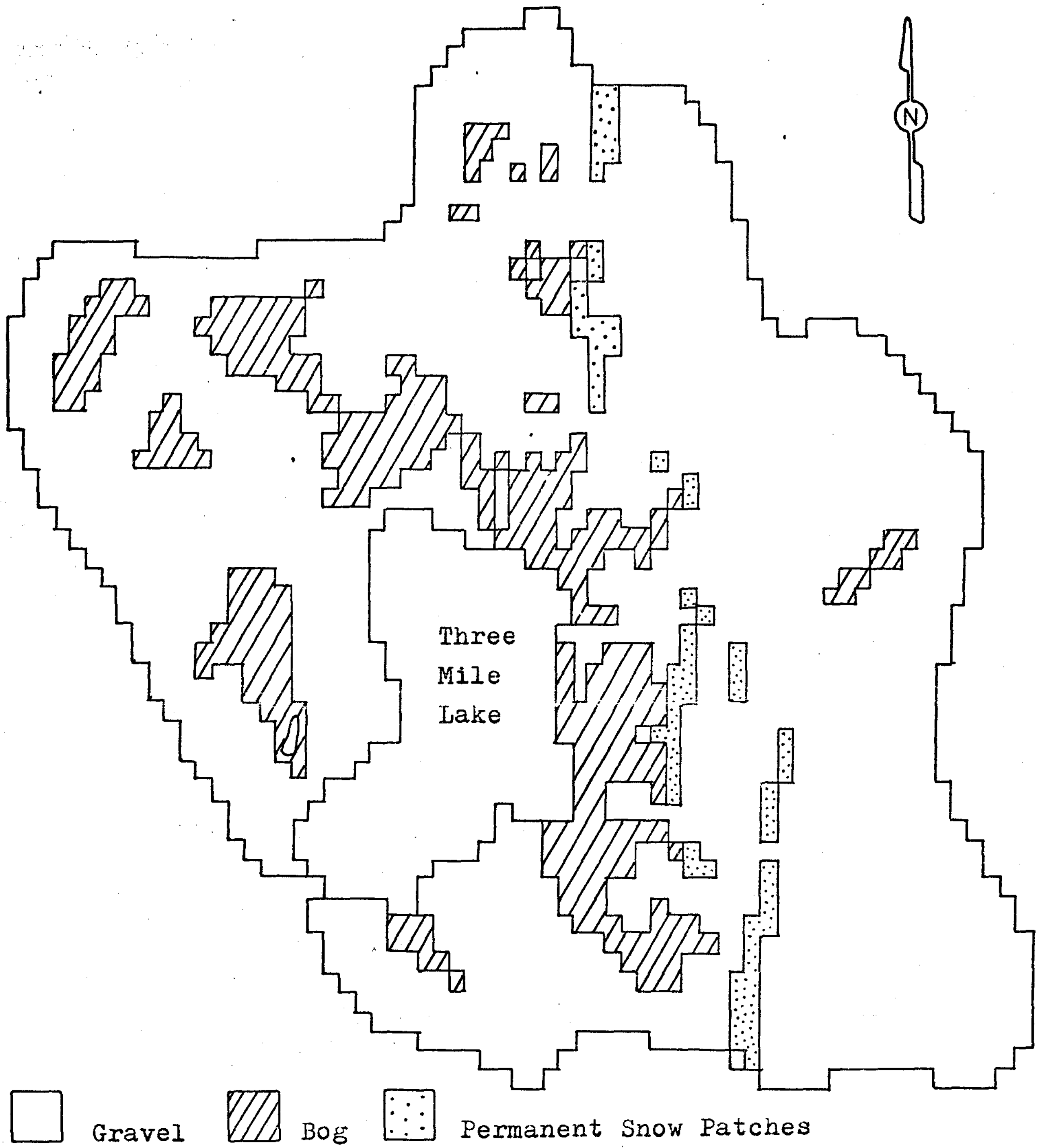
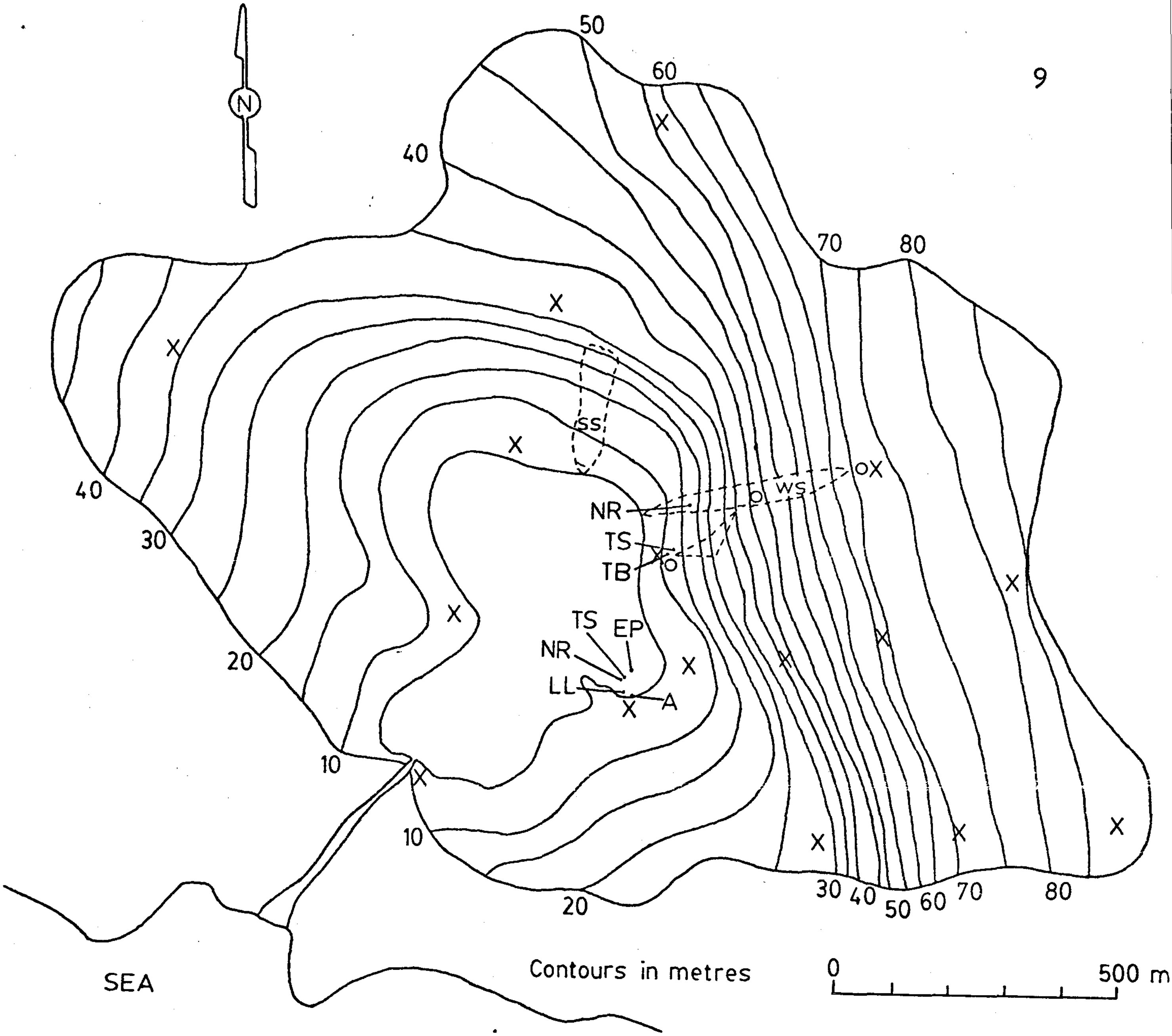


Figure 2.2. Computer-mapped surface cover of "Three Mile Lake" basin.



WS	west-facing slope	A	anemometer
SS	south-facing slope	EP	evaporation pan
TS	thermohygrograph screen	TB	tipping bucket rain-gauge
NR	net radiometer	X	manual rain-gauge
LL	lake level recorder	o	trace precipitation measurement station

Figure 2.3 Instrument location within the study basin. Weir 1 drains the area south of the west slope; weir 2 is slightly north of the shoreline on the south-facing slope; and weir 4 is located at the mouth of the west-facing slope.

Trace precipitation was measured using an impermeable platform covered with absorbent paper towels held firm by a frame strung with picture-framing wire (Figure 2.4). Three such devices, standing approximately 15 cm. high, were placed at 11 m, 42 m., and 72 m. above sea level up the west-facing slope. Dry paper towels were weighed on an Ohaus Triple-beam balance and then set out at the start of a trace precipitation event. At the end of the event, the wet towels were collected and re-weighed. The difference in weight divided by the catch area provides a measure of trace precipitation.

Figure 2.4 Trace Precipitation Measurement Device

Evaporation

Evaporation over both land and water was measured during the month of August. Lysimeters were used to measure evaporation from various surface types (Figure 2.5). An undisturbed soil sample was placed in a tin which was then set back into the ground inside another tin of slightly larger diameter. This prevented lateral flow of groundwater into the sample. The top of the sample was flush with the ground. Lysimeters were paired in gravel and bog; one 'wet' and one 'dry'. The dry lysimeter was allowed to evaporate naturally throughout the study period whereas the wet lysimeter had water added to maintain constant wetness. Each lysimeter was weighed daily on an Ohaus triple-beam balance. The weight differences from day to day divided by the evaporating area gives daily evaporation.

Evaporation from a free water surface (Three Mile Lake) was measured using an evaporation pan 0.45 m. in diameter placed in the lake. The lip of the pan was approximately 18 cm. above the water surface. A pointer gauge was placed in the pan and water added until the pointer just broke the water surface. The amount added was recorded and this, divided by the evaporating area, gives daily evaporation. If precipitation raised the water level above the pointer, the pan was bailed to the level of the pointer and the amount recorded.

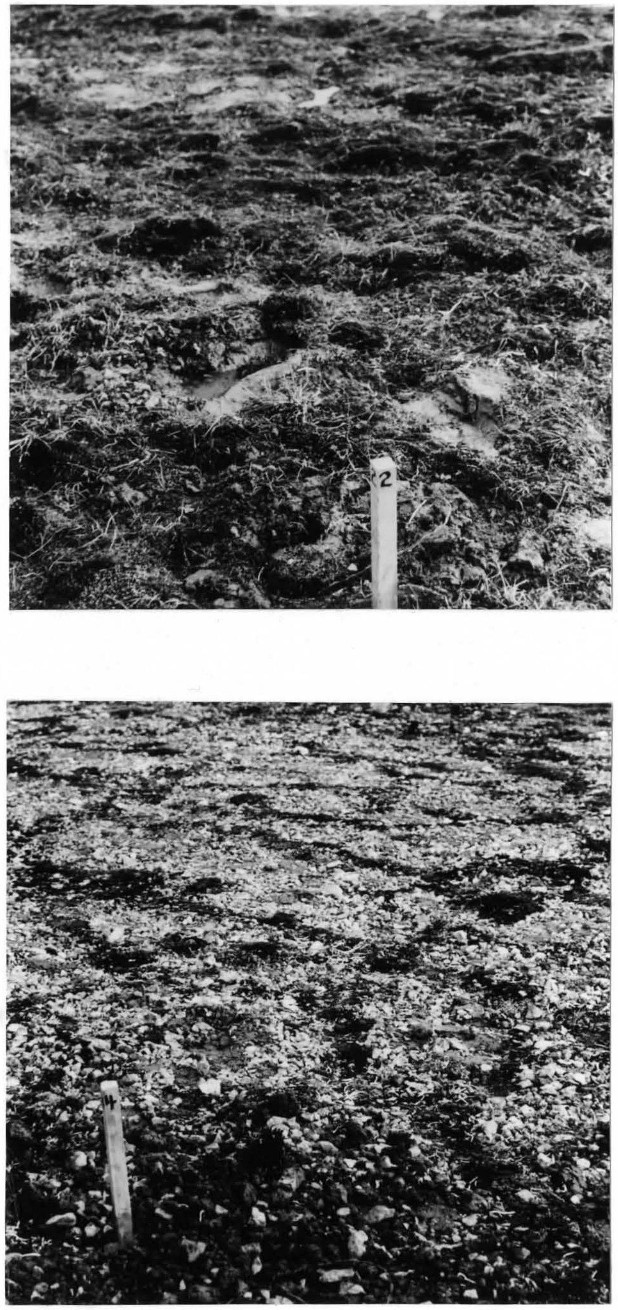


Figure 2.5 Evaporation lysimeters in an area east of Three Mile Lake : the top photo shows the 'dry bog' lysimeter; the bottom photo a 'wet gravel' lysimeter.

Pan-water and lake-water termperatures were periodically checked with a mercury thermometer to ensure that the pan was neither warming nor cooling at rates different from the lake.

Equilibrium evaporation may be estimated using the following equation (Priestley and Taylor, 1972):

$$LE_{EQ} = s/s + \gamma \quad (Q^* - G) \qquad (2.1)$$

where LE_{EQ} is the equilibrium evaporation rate, s is the slope of the saturation vapour pressure vs. temperature curve, γ is the psychometric constant, Q is net radiation, and G is the ground heat flux.

Since, in the Arctic, ground heat flux is small compared to net radiation (Weller and Holmgren, 1974) equation (2.1) becomes:

$$LE_{EQ} = s/s + \gamma \ (Q^*) \qquad (2.2)$$

Thus evaporation may be estimated using only net radiation (Q^*) and $s/s + \gamma$ which is temperature dependent (Woo, 1976).

To obtain actual evaporation, Priestley and Taylor (1972) proposed:

$$LE = \alpha \ LE_{EQ} \qquad (2.3)$$

where α is a coefficient which ranges from 0 for non-evaporating surfaces, to some value exceeding 1.0. For saturated conditions, Davies and Allen (1973) and Stewart and

Rouse (1977) obtained an empirical value of $\alpha = 1.26$. For the present study, α is allowed to vary as a function of soil moisture (Marsh and Woo, 1979).

Net Radiation

Net radiation (Q^*) over bog and open water was measured for the month of August, using Swissteco net radiometers whose signals were recorded by Rustrak recorders, after appropriate amplification. Missing values were estimated by regressions of the form

$$Q^* = a + b\,K\downarrow \qquad (2.4)$$

where $K\downarrow$ is shortwave radiation measured by the Atmospheric Environment Service (A.E.S.) and a and b are empirically determined coefficients. Net radiation over gravel was measured by A.E.S.

Air Temperature

Air temperature at screen height (1 m.) was measured over a gravel site for the entire study period using a Lambrecht thermohygrograph. A Lambrecht thermohygrograph housed in a Stevenson's screen was used to measure temperature over water. This screen was approximately 0.6 m. above the water surface to facilitate the checking of the instrument. Both thermohygrographs were checked using a mercury thermometer. It was found that the temperature above the lake was quite similar to the air temperature at the shore (Figure 2.6).

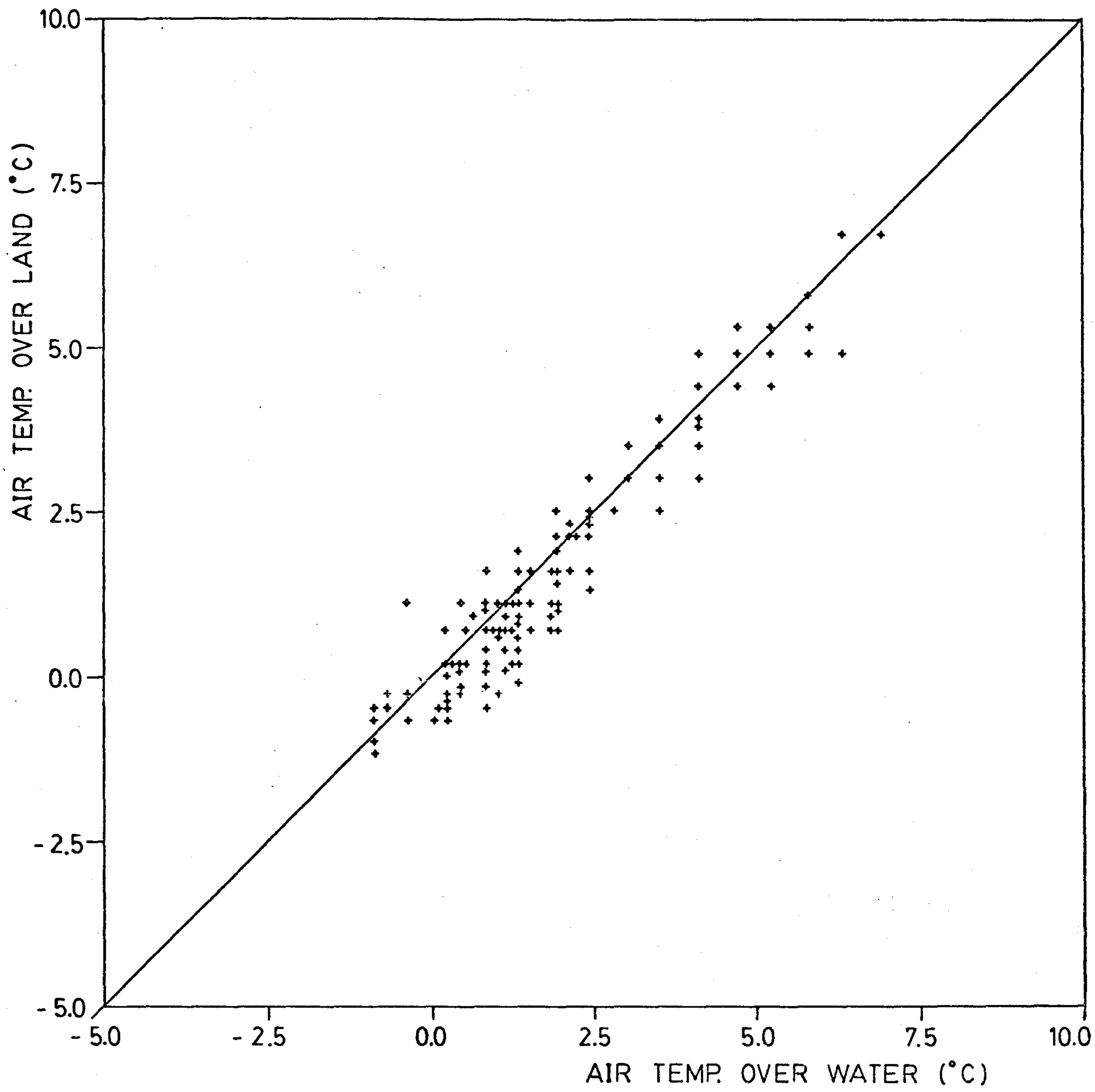


Figure 2.6 Relationship between screen air temperature over Three Mile Lake and screen air temperature over land. A regression coefficient of 0.97 was achieved, thus, for practical purposes, a 1:1 line was assumed.

2.2.2 Hydrologic Data

Lake Water Level

The water level of Three Mile Lake was recorded continuously from July 5th using a Leupold-Stevens Type F water level recorder sitting on a stilling well set in the lake (Figure 2.3). Before outflow began the recorder had to be moved due to excessive lake level rise caused by snow-melt.

Discharge

Three Mile Lake drains to the Arctic Ocean through one outflow channel located at the southwestern end of the lake (Figure 2.7). This channel remains choked with snow until sometime after melt begins. After the channel opened, discharge was current-metered twice daily for a week, by which time the outflow channel stabilized and a stage-discharge relationship was established to compute streamflow from the lake level (Figure 2.8). Discharge was obtained by the velocity-area method, with velocity determined by a Price-type current meter.

Overland Flow

Overland flow was measured at several representative sites in the study basin (Figure 2.3). V-notch weirs were constructed and flow was channelled through the weir (Figure 3.8). Water height behind the weir was recorded using a

Figure 2.7 The outflow channel at the southwestern end of Three Mile Lake . Top: During the snowmelt period before the channel-break. Bottom: After the channel-break but still early in the season. Note the rope knotted at 0.5 m. intervals across the channel. A wading rod and current meter are at the lower right of the photo.

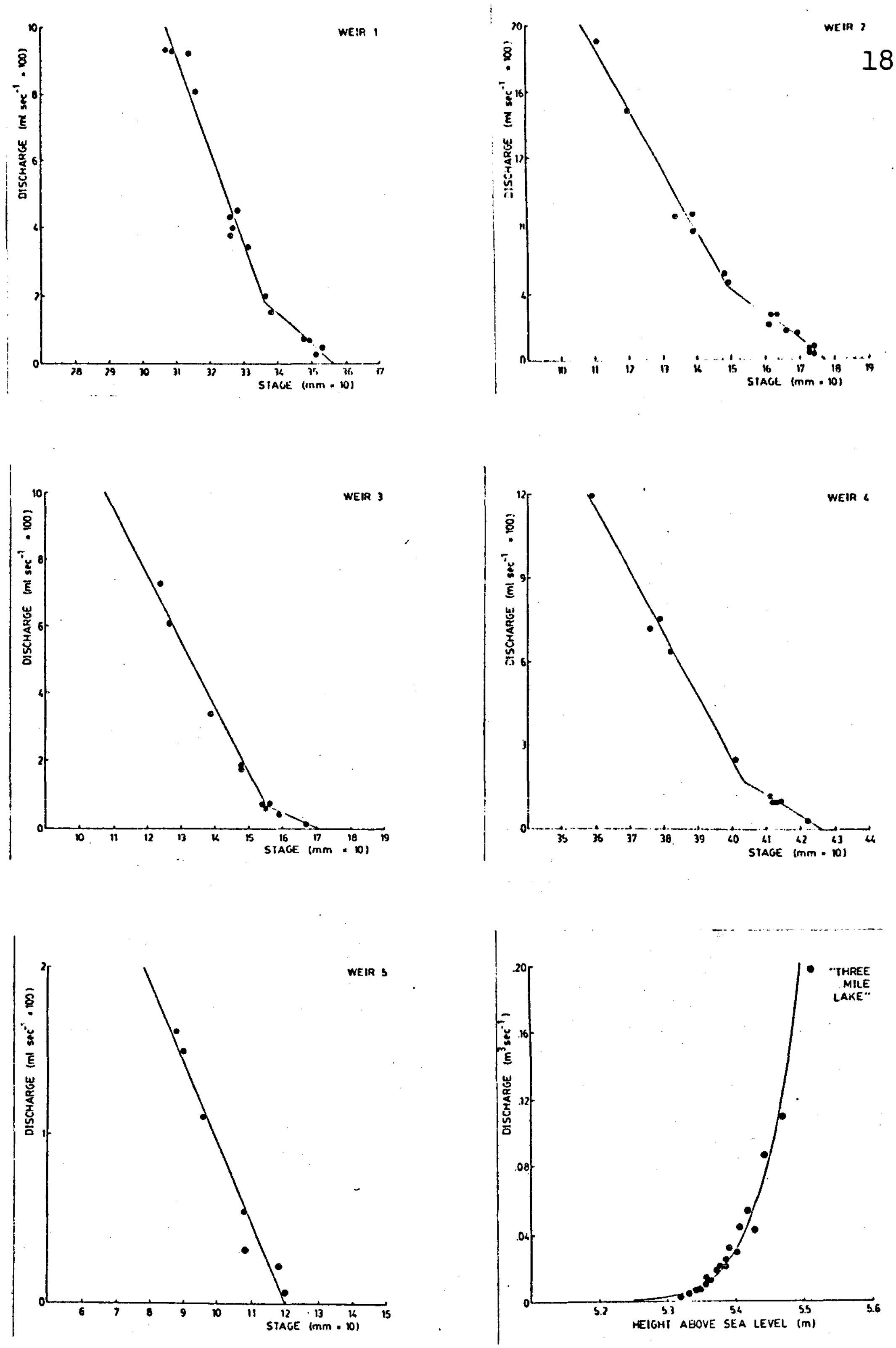


Figure 2.8 Stage-discharge relationships for five weirs and Three Mile Lake. Stage is distance down from an arbitrary datum. Equations for the lines are given in Appendix Two.

Leupold-Stevens Type F water level recorder. Discharge through the v-notch was determined by timing with a stopwatch the rate of fill of a container of known volume. Water level was then converted to discharge using rating curves shown in Figure 2.8.

Groundwater Table

Changes in supra-permafrost groundwater table were measured on two different slopes; one facing west and one facing south (Figure 2.3). Pits were dug at irregular intervals up each slope and the water table was measured from a reference point established across the top of the pit. At three of the pits on the west-facing slope the water table was continuously recorded using Leupold-Stevens Type F water level recorders.

Frost Table

Whenever possible, daily changes in frost table depth were measured at each groundwater pit by driving a steel rod into the ground until the frost table prevented further penetration.

Soil Moisture

Soil samples were taken from areas adjacent to the lysimeters using a sampler constructed from a steel pipe. Samples were weighed on an Ohaus triple-beam balance and then dried in a portable Coleman oven on top of a hot plate set at a temperature of 100° C $\pm$ 5° C. The dried samples were then re-weighed to determine the gravimetric soil moisture content. Soil samples were also taken near three groundwater pits up the south-facing slope. To obtain representative values of the bulk density, the contents of the lysimeters were dried at the end of the season and the dry weights were then divided by the volume of the containers.

Hydraulic Conductivity

Hydraulic conductivity was measured periodically during the month of August. 'ABS' plastic pipes punctured with holes approximately 10 mm. in diameter were dug into the ground until they reached the frost table. The pipes and the soil were allowed to settle for several days before measurements were taken. The original water table was noted before water was pumped out of the pipe. Then the rate at which the water table rose back to the original level was timed to enable hydraulic conductivity computation.

Specific Yield

Specific yield was determined by saturating an undisturbed soil sample for 48 hours, decanting any water ponded on top and then letting the sample drain through a mesh screen for 48 hours. The amount of water drained was recorded, and the specific yield (Sy) was calculated as follows;

$$Sy = V_w/V_s \quad (2.5)$$

where V_w is volume of water drained, and V_s is volume of the sample.

2.3 Instrument Accuracy

All clocks on the recorders were checked daily when the instruments were checked and adjustments were made if necessary. The Ohaus triple-beam balance was checked by weighing a wad of paper towels and an empty soil sampler tin in the field. These two articles were re-weighed at McMaster University using a Mettler P162 electronic balance accurate to 0.1 gm. which did not produce significant error in the determination of soil moisture nor trace rainfall.

CHAPTER 3

WATER INPUTS

3.1 Inputs to the Basin

The input components of the basin water balance equation (1.2) are snowfall, rainfall, and trace precipitation. The winter snowfall component will be examined in conjunction with snowmelt. Rainfall will be discussed on the basis of isohyetal analysis of ten storms which occurred during July and August.

It is possible that during dry summers trace precipitation represents a significant proportion of total recorded rainfall, however, the 1978 field season was relatively wet and trace precipitation need not be considered as an important component of the water balance equation (Woo and Steer, 1979).

3.1.1 Snowmelt

In High Arctic areas snowfall can be stored in drainage basins for as long as ten months of the year (Cook, 1967; Woo and Marsh, 1978). This fact enables accurate determination of basin snow storage by a late winter snow survey before the first day of melt. Marsh (1978) has discussed the importance of spatial distribution of snow within a drainage basin during the snowmelt period. Snow-covered areas are sources of meltwater which contribute to overland flow and, for the present study area, subsequent lake level rises.

To determine total winter storage, the basin under study was divided into grid squares measuring 29.3 m. X 29.3 m. Each square was assigned a mean snow water equivalent value based upon field data obtained by a late winter snow survey. These values were used in conjunction with measured melt-rate obtained during the first two weeks of snowmelt to map the percent bare area of the basin as snowmelt progresses.

The computer-generated map in Figure 3.1 shows basin snow storage before the first day of melt. The map has been adjusted to account for new snowfall in the basin since the day of the survey. Depressions and hollows around the lake and on the west-facing slope generally contain the thickest pack and the most uneven distribution. The lake itself and the northwestern sector of the basin are covered by a thinner and more uniform snowpack.

During the first four days after initial melting (June 30th) the snowpack opened so that on the fifth day 6.5 percent of the basin became snowfree. Dissipation of the snowpack proceeded rapidly, baring 61.5 percent of the basin just three days later (eight days after melt began). By the end of the second week (July 13th) 93.2 percent of the basin was snowfree (Table 3.1). The bulk of the remaining 6.8 percent was stored on the west-facing slope. By the end of

DAY RESET SNOW

MEAN W.E. = 198.1

PROPORTION BARE = 0.000

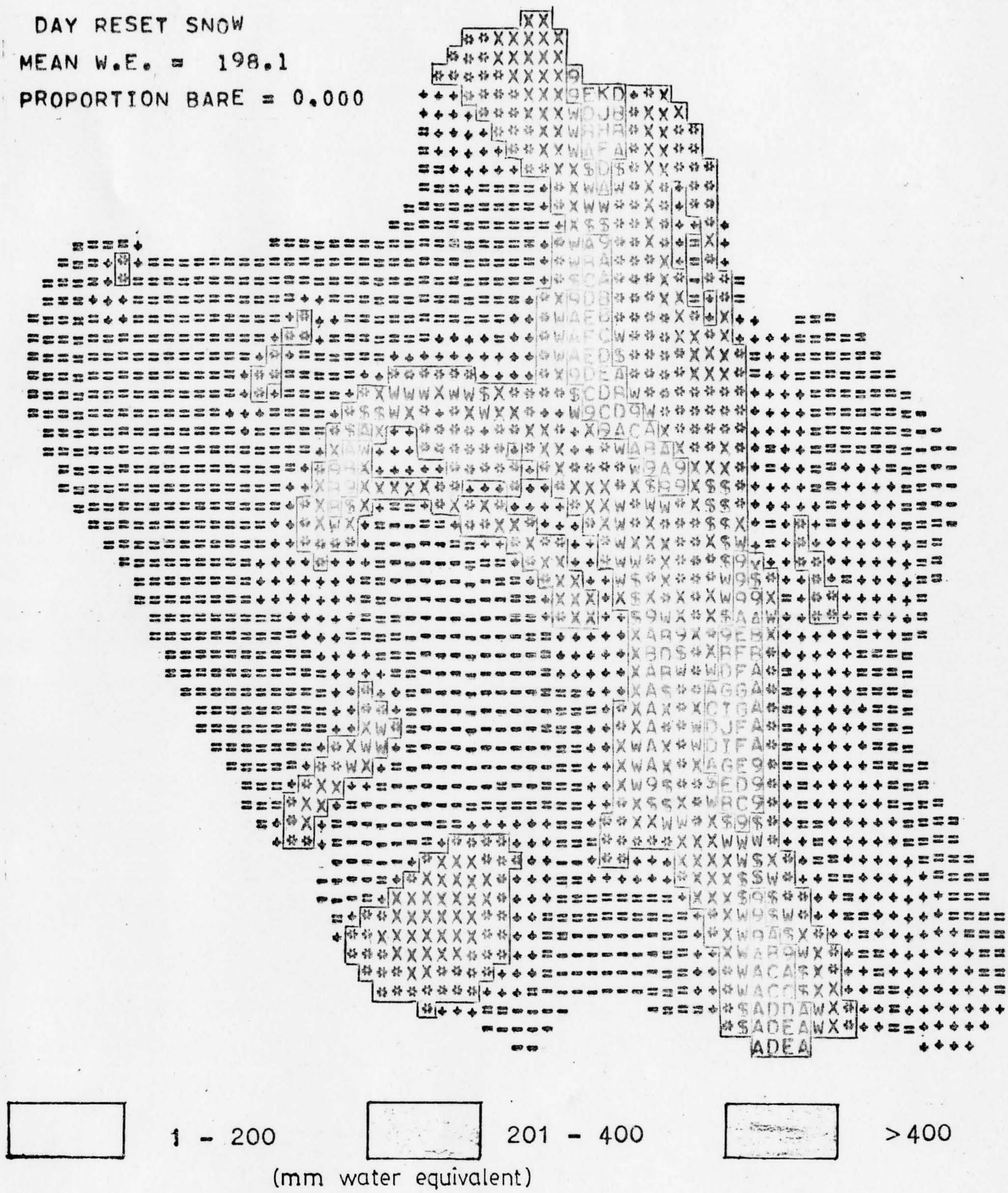


Figure 3.1. Computer-mapped snow storage at the end of the winter season.

TABLE 3.1

Dissipation of Snowpack in Three Mile Lake Basin

During the First Two Weeks of Snowmelt 1978

Day	Date	% Bare	Daily Snowmelt	Residual Snow Storage In Water Equivalent Units (mm.)
180	29 06 78	0.0	0.0	198.1
181	30 06 78	0.0	10.56	187.5
182	01 07 78	0.0	13.80	173.7
183	02 07 78	0.0	19.30	154.4
184	03 07 78	0.0	22.70	131.7
185	04 07 78	6.5	21.59	110.1
186	05 07 78	12.1	26.31	83.8
187	06 07 78	48.3	28.79	55.0
188	07 07 78	61.5	12.46	42.6
189	08 07 78	73.3	12.97	29.6
190	09 07 78	88.0	10.91	18.7
191	10 07 78	89.6	2.03	16.7
192	11 07 78	90.5	1.90	14.8
193	12 07 78	92.2	2.76	12.0
194	13 07 78	93.2	2.14	9.9

July an additional 4.13 percent of the basin had become snow-free; the residual 2.67 percent remained as permanent snow patches on the west-facing slope (Figure 2.2). Figure 3.2 shows snowmelt from the last day of snowpack ripening (184) to the end of the second week of snowmelt at which time 93.2 percent of the basin was snowfree (194). This series of figures confirms the rapid rate at which melt occurs in the Arctic; a fact noted by many other authors (eg. Woo, 1976; Anderson, 1974; Brown et al., 1968; Cook, 1967; Cruikshank, 1971).

3.1.2 Summer Precipitation

During July and August 1978, the Resolute weather station recorded 66.4 mm. of precipitation, including snowfall events recorded in August. This summer precipitation total is slightly higher than the long term station mean of 59 mm. (Dept. of Envir.,1971), but is similar to the 62.6 mm. recorded using 16 manual gauges scattered in the drainage basin (Figure 3.3). The latter figure was obtained from isohyetal analysis of ten rainstorms during July and August (Table 3.2) and it incorporated two A.E.S. snowfall readings because the manual gauges were clogged by snow. The major disparity between the two curves shown in Figure 3.3 is due to the August 2-3 storm, the largest of the season. The Atmospheric Environment Service recorded 17.2 mm. over the two day period whereas 24.3 mm. was obtained in the Three Mile

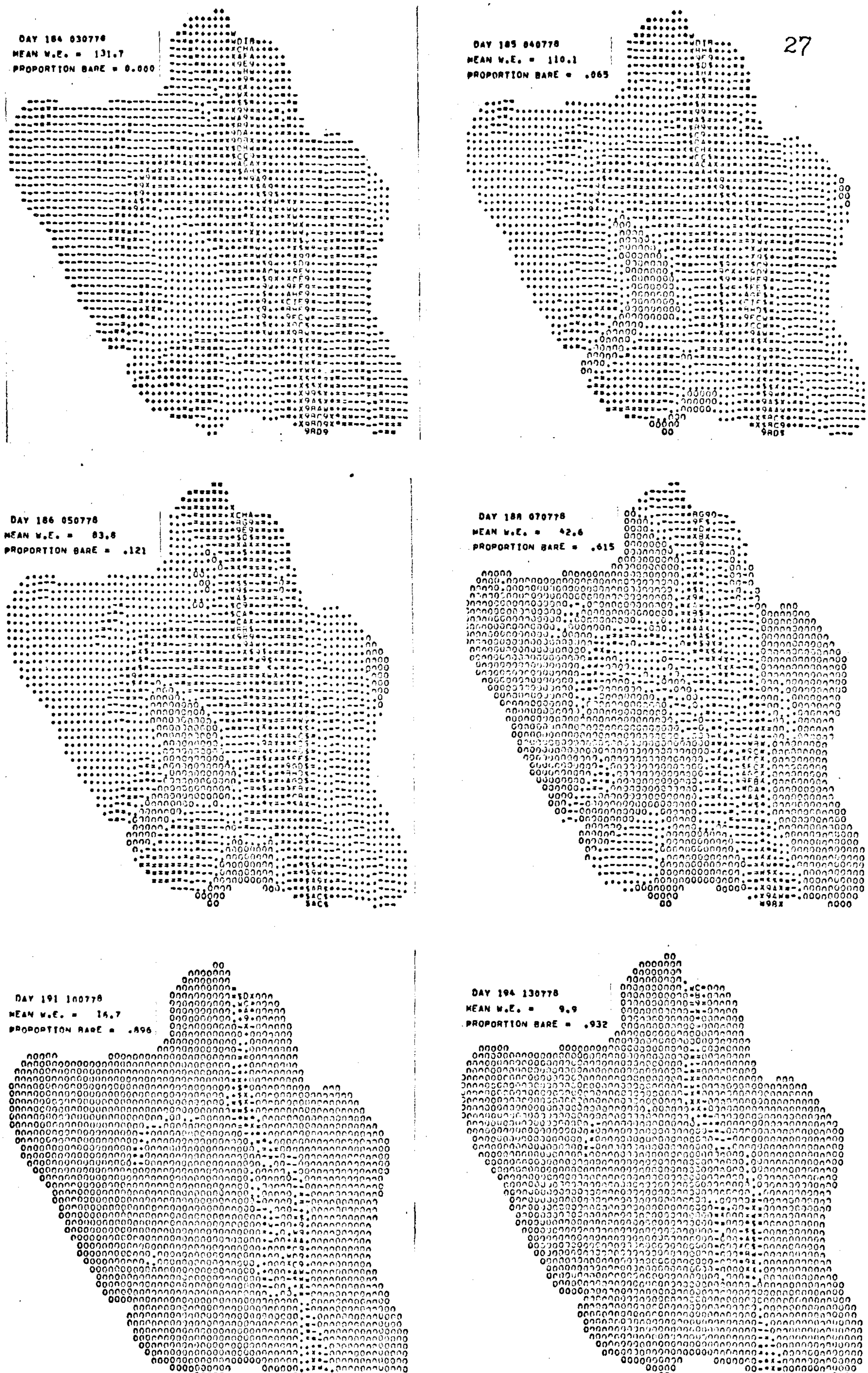


Figure 3.2 Dissipation of the snowpack during a ten-day period after the pack has ripened. Zero indicates bare ground. North is to the top.

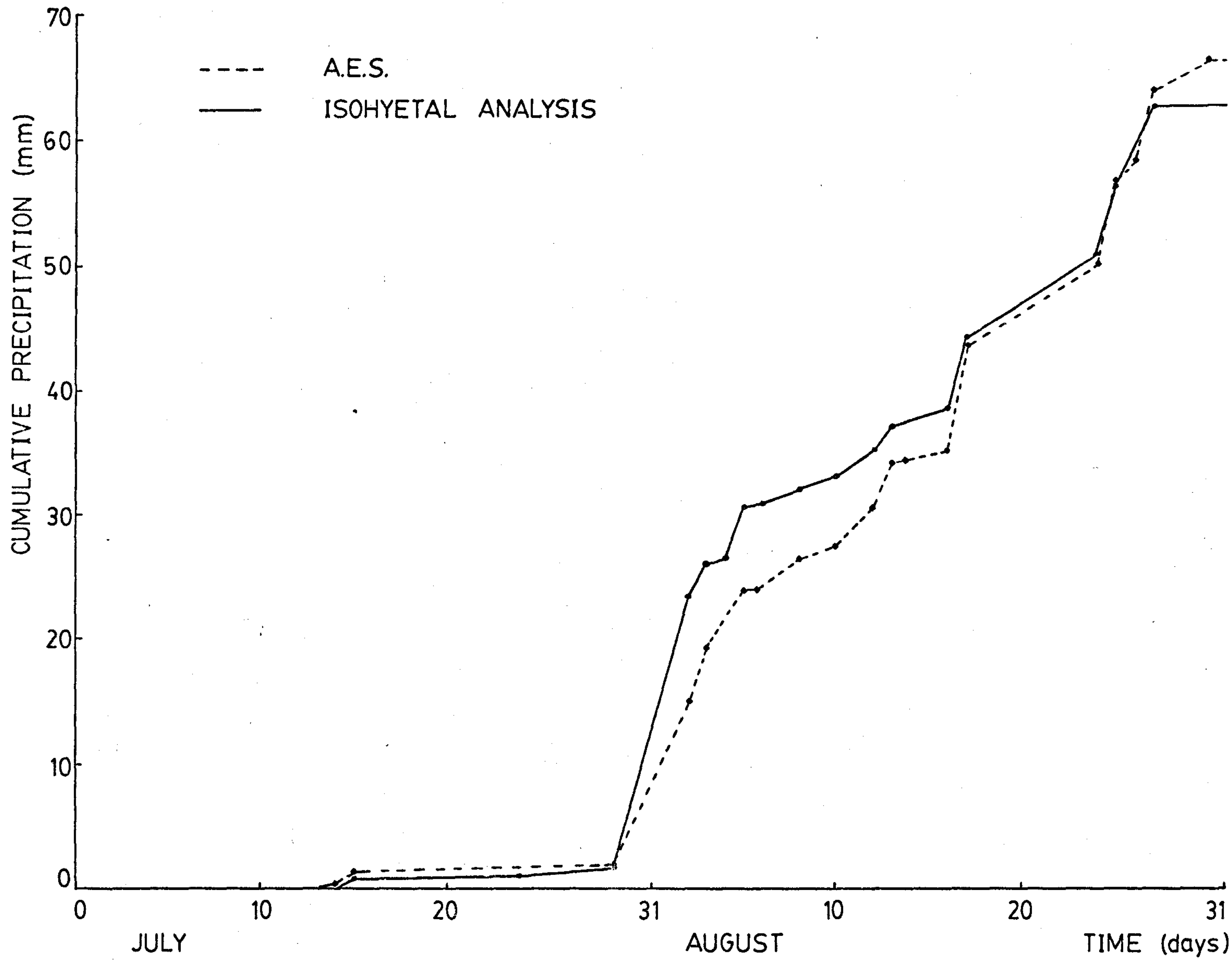


Figure 3.3 Comulative summer precipitation during July and August as recorded at Resolute airstrip by A.E.S. and by sixteen manual gauges in the study basin.

TABLE 3.2

Summer Precipitation By Storm

(all values are in mm.)

Date	Arithmetic Mean	Thiessen Polygon Method	Isohyetal Analysis
July 15	0.8	0.8	0.8
24	0.3	0.3	0.3
29	0.5	0.6	0.6
Aug. 2-3	23.9	23.9	24.3
5	4.9	4.9	4.6
6 (A.E.S. snow)	0.3	0.3	0.3
8	1.7	1.6	1.3
10	1.1	1.0	1.0
12-13	4.2	4.2	4.0
16-17	7.1	7.4	7.2
24-25	12.9	12.7	12.7
27 (A.E.S. snow)	5.6	5.6	5.6
Seasonal Totals	63.2	63.3	62.6
Mean Total from Three Different Methods	63.0		
A.E.S. TOTAL	66.4		

Lake basin. Following this storm, A.E.S. consistently recorded slightly higher precipitation than the gauges at Three Mile Lake until mid-August at which time the curves again show good agreement. The final observation on the A.E.S. curve accounts for 60.7 percent of the total difference between the two curves. This represents a 2.5 mm. snowfall on August 30th which went unrecorded in the lake basin. The general conclusion drawn from Figure 3.3 is that A.E.S. rainfall data is representative of basin rainfall for the study period. However, this should be considered coincidental in view of the spatial variability of precipitation events observed during the study period.

During the ten storms recorded, the predominant wind direction ranges between southeast and south-southeast (Figure 3.4). A general pattern of rainfall distribution shows high values to the west and the south of the lake, decreasing both northwestward and towards the highest parts of the basin. During the storm of August 8th (southwesterly wind) precipitation decreased upslope rather than increasing, whereas a storm four days later (northerly wind) deposited rainfall evenly throughout the basin. The storm of August 16-17 shows increasing precipitation from east to west, possibly a result of winds blowing downslope and not being able to release as much moisture on the west-facing slope. The above analysis shows that topographic effects are very difficult to determine

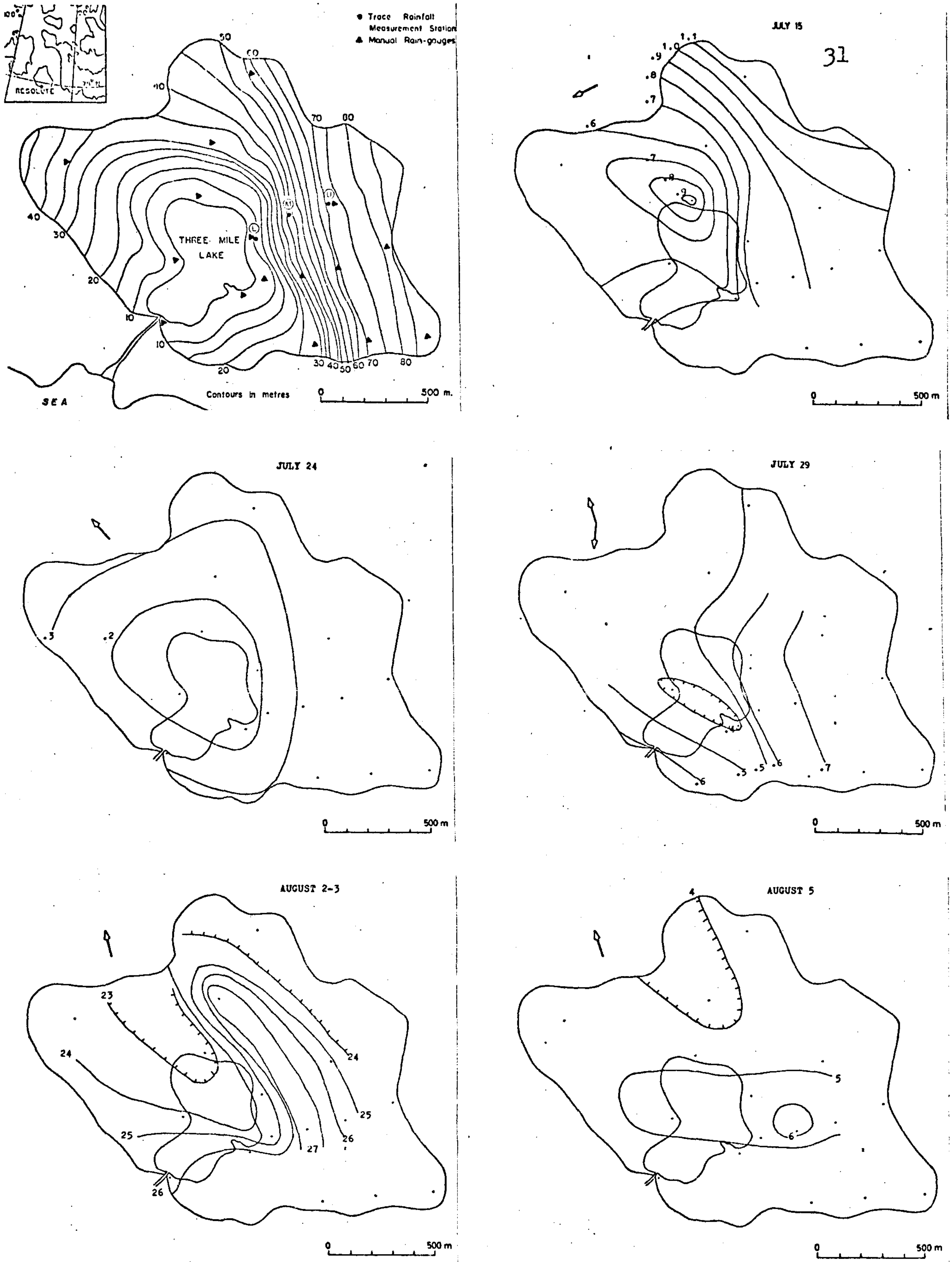


Figure 3.4 Isohyetal analysis of ten storms recorded during the study period. Rain-gauge locations and elevations are shown at the upper left.

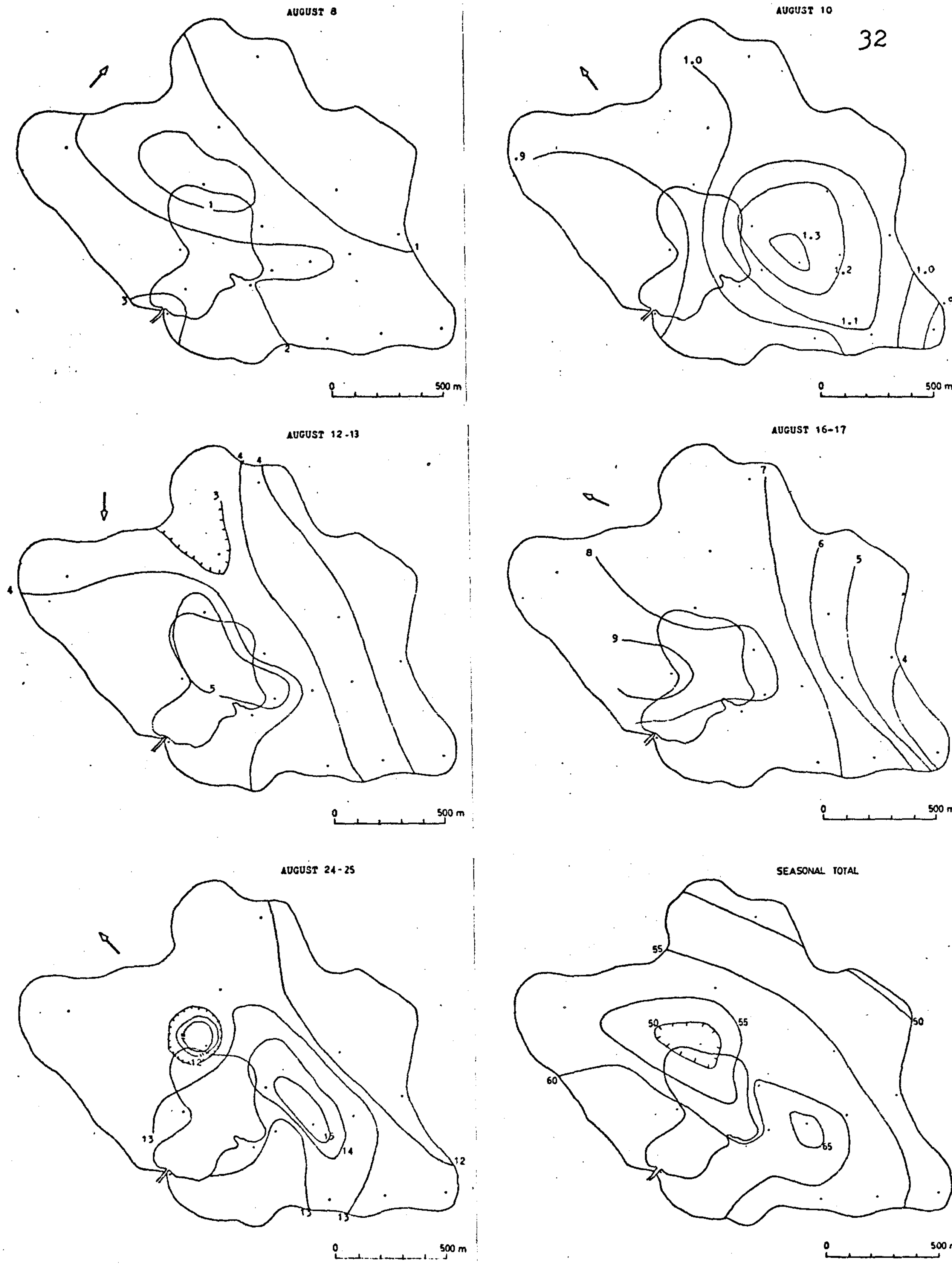


Figure 3.4 cont'd

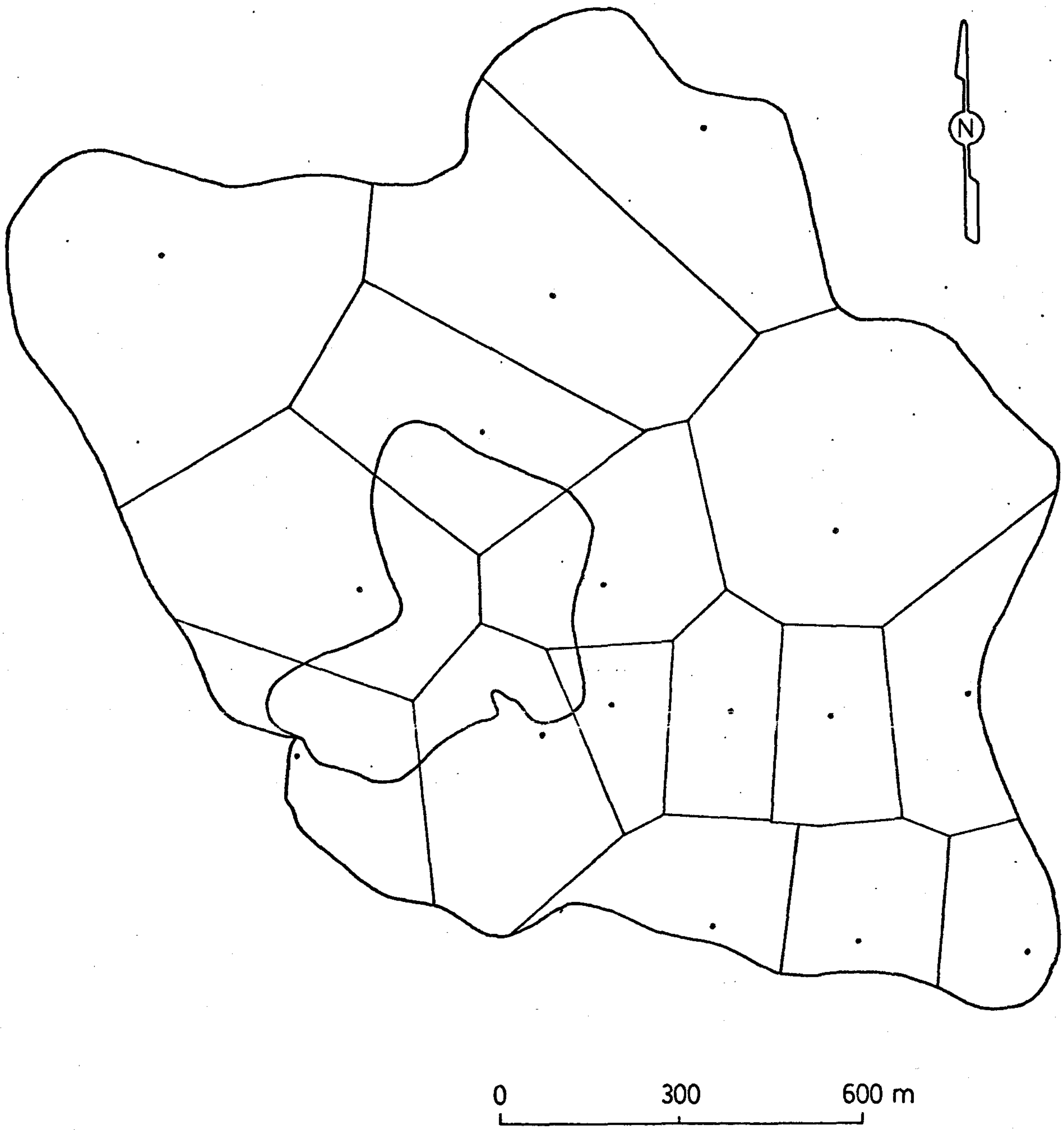


Figure 3.5 Thiessen polygons used to compute mean basin rainfall as reported in Table 3.2.

and that the spatial distribution of summer precipitation over the basin follows no easily determinable pattern.

3.2 Inputs to the Lake

The input components of the lake water balance equation (1.3) are P_L, precipitation falling directly onto the lake; Q_O, overland flow and channelled flow; and R_S, subsurface flow. Snowmelt from the rest of the basin contributing to lake level changes will be included in the overland flow,channelled flow, and subsurface flow terms. Meltwater from snow actually stored on the lake ice must either percolate through the ice until it refreezes or go into storage on top of the lake ice until such time as the moat around the lake is formed, the ice floats free from the bottom of the lake and meltwater may run off the ice and contribute to lake level. From the day melt began until the day the moat was established the snow stored on the lake ice accounted for 6.2 mm. of basin snowmelt. By this time snow stored on the lake was almost completely dissipated and further increases in lake level were due to snowmelt runoff from the land; precipitation being negligible until after the channel-break.

The contribution of lake-ice ablation to lake level may be ignored because when the ice is free to float, it displaces a volume of water whose weight equals the weight of the ice. Consequently, the lake water level has implicitly included a contribution from the lake ice equivalent to the water equivalent of this ice. Figure 3.6 shows various stages of lake-ice ablation during the summer of 1978. These results are summarized in Figure 3.7. The general shape of the curve is in agreement with ice ablation rates observed by other authors (eg. Minns, 1977; Schindler et al., 1974).

3.2.1 Direct Precipitation

Summer precipitation falling directly onto the lake surface may be easily calculated from the isohyetal analysis previously performed for the entire basin. The results are displayed in Table 3.3. The total for the lake accounts for 10.1 percent of the seasonal total for the entire basin and this represents a significant contribution.

3.2.2 Overland Flow

Overland flow occurs when either or both of two conditions is met; the first instance is when the precipitation rate exceeds the infiltration rate and the second occurs when the soil is saturated before the start of a rainfall event (Linsley et al., 1975). In both cases, the degree of saturation and the depth of the active layer are important factors (Dingman, 1973). Early in the season when the active layer is shallow, total moisture holding capacity is reduced and the soil will saturate more rapidly than under conditions of a deeper active layer (assuming constant antecedent soil moisture).

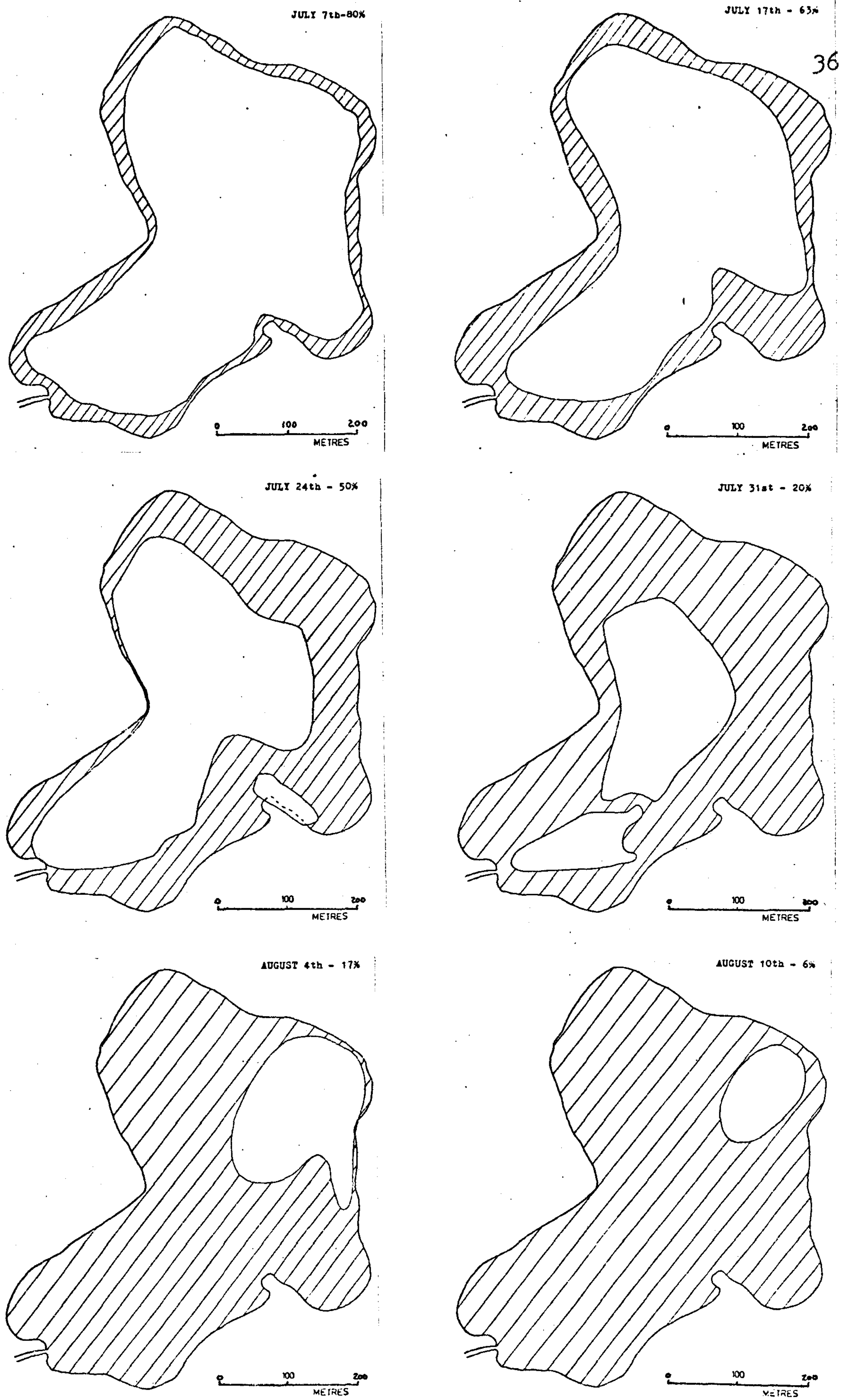


Figure 3.6 Various stages of lake ice ablation during the summer of 1978. By August 11th, all ice had disappeared from the lake surface.

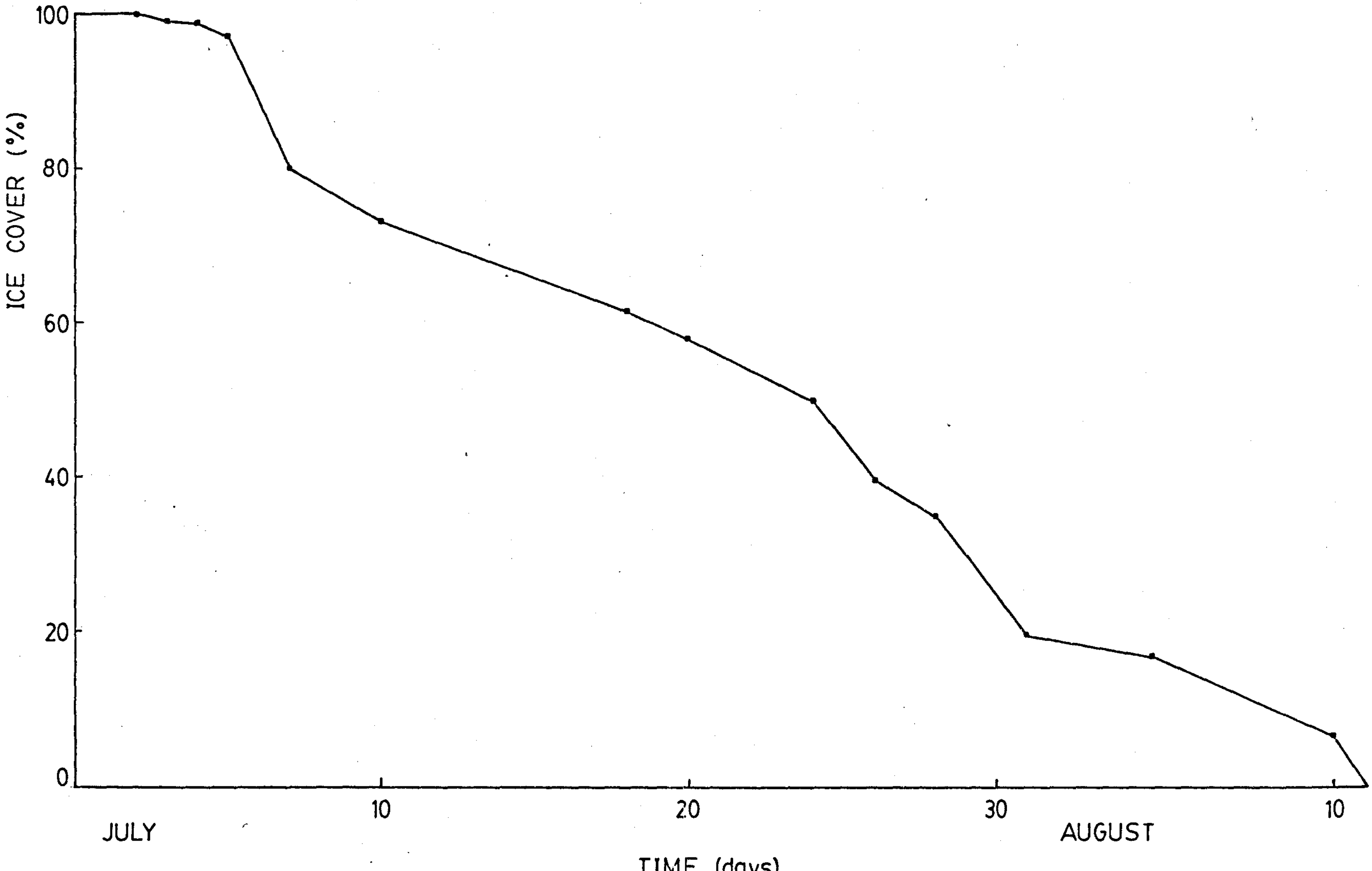


Figure 3.7 Dissipation of lake ice during the 1978 summer season.

TABLE 3.3

Direct Summer Precipitation Onto the Lake Surface

Date	July 15	July 24	July 29	Aug. 2-3	Aug. 5	Aug. 6
Amount Direct Precipitation	0.08	0.02	0.08	2.3	0.5	0.02

Aug. 8	Aug. 10	Aug. 12-13	Aug. 16-17	Aug. 24-25	Aug. 27
0.2	0.08	0.5	0.8	1.3	0.4

total 6.28 mm

percent of seasonal total for entire basin 10.1

In the basin under study, a large volume of overland flow occurs during the snowmelt period. At this time, the active layer is shallow and the outflow channel below the lake is still snow-choked. Under these conditions, the amount of snowmelt contributing to overland flow may be estimated by observing the rate of lake level rise. More specifically:

$$P_G + M - E_G - \Delta S_G = Q_0 + R_S = \Delta S_L + E_L - P_L \qquad (3.1)$$

where P_G is rainfall onto land surfaces; M is melt; E_G is evaporation from land surfaces; ΔS_G is change in land storage; Q_0 is overland flow discharge; R_S is groundwater flow into the lake; ΔS_L is change in lake storage; E_L is evaporation from the lake; and P_L is direct precipitation onto the lake surface. From the first day of snowmelt until the commencement of outflow numerical values for various components of equation (3.1) are as follows: $P_G = 0.7$ mm.; $M = 191.1$ mm.; $E_G = 22.3$ mm.; $E_L = 1.09$ mm.; and $P_L = 0.08$ mm. The ΔS_L term is obtained by multiplying the observed lake level rise by an adjustment factor to convert rise in lake level (mm.) to a water depth spread over the entire land area of the basin. This factor is 0.097 which is a ratio of the lake area to the basin area minus the lake area. By converting the known water level rise of 1265.8 mm. to ΔS_L, a value of 122.4 mm. was obtained. Then,by subtracting the E_L term to get net ΔS_L, ΔS_G can be calculated by:

$$\Delta S_G = P_G + M - E_G - \Delta S_L \qquad (3.2)$$

Using the reported values for the right hand side of equation (3.2) the change in groundwater storage during the snowmelt period was 48.2 mm.

The preceding computations encompass a period from the beginning of snowmelt until the outflow channel opened and discharge from the lake began. Overland flow from this point on was measured using three v-notch weirs (Figures 3.8 to 3.10) draining a total area of 34706 m^2.

Hydrographs for the three overland flow basins are shown in Figure 3.11. From these hydrographs, daily flow was computed and the values are given in Table 3.4. For the period between July 20th and August 31st total overland flow was computed to be 1.3 mm. and this was less than 2 percent of the total overland flow and groundwater flow calculated for the snowmelt period.

3.2.3 Subsurface Flow

The groundwater component of the water balance equation (for the lake) is the most difficult to measure and so for the purposes of this study will be calculated as a residual of the lake water balance equation (1.3). Equation (3.1) will show that:

$$R_S = \Delta S_L + E_L - Q_0 - P_L + Q \qquad (3.3)$$

where R_S is subsurface or groundwater flow; ΔS_L is change in

Figure 3.8 Overland flow measurement weirs. Top: number 1, near the met. site looking up the west-facing slope; Middle: number 2. The south-facing slope rises to the left. Bottom: number 4, looking up the west-facing slope.

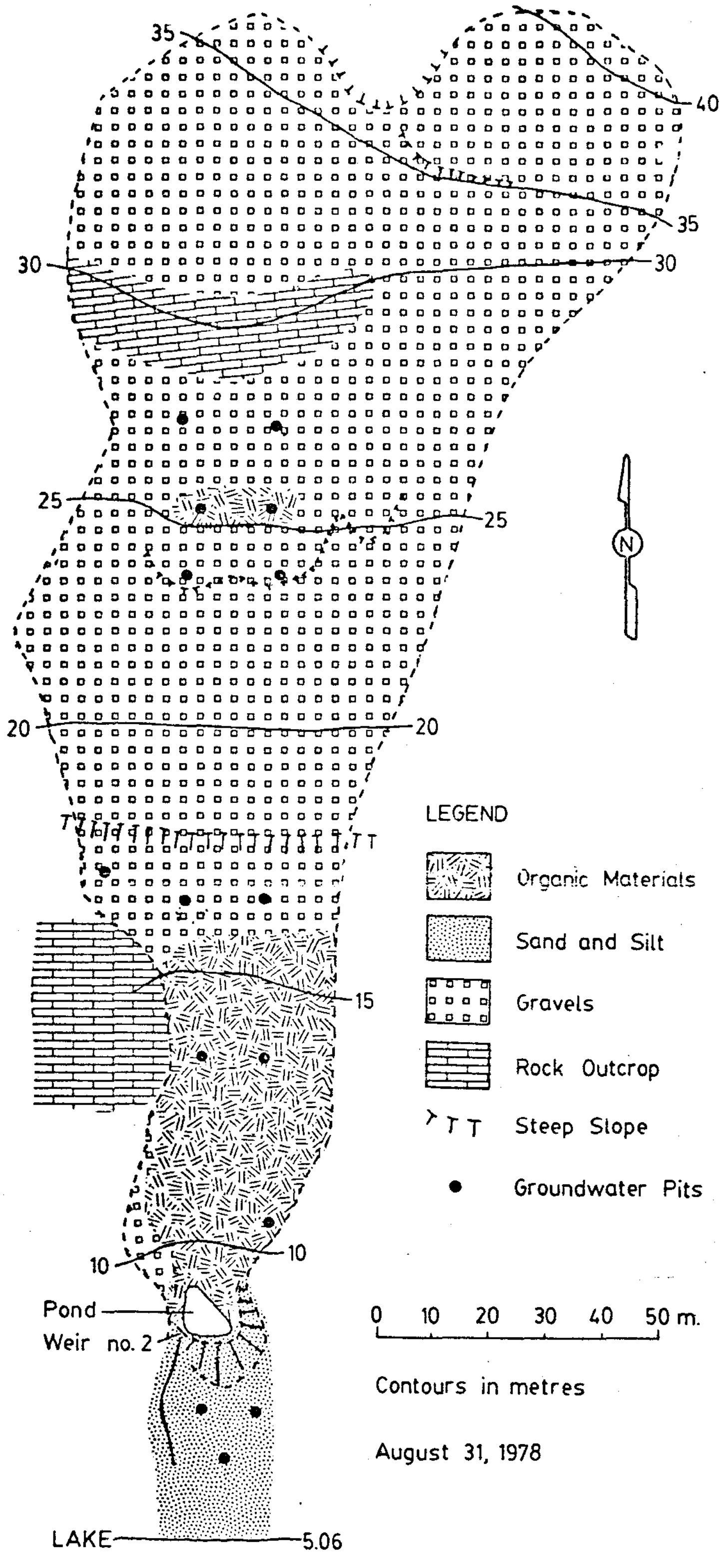


Figure 3.9 Surface features of the south-facing slope in the study basin. Location of the slope within the basin is shown in Figure 2.3.

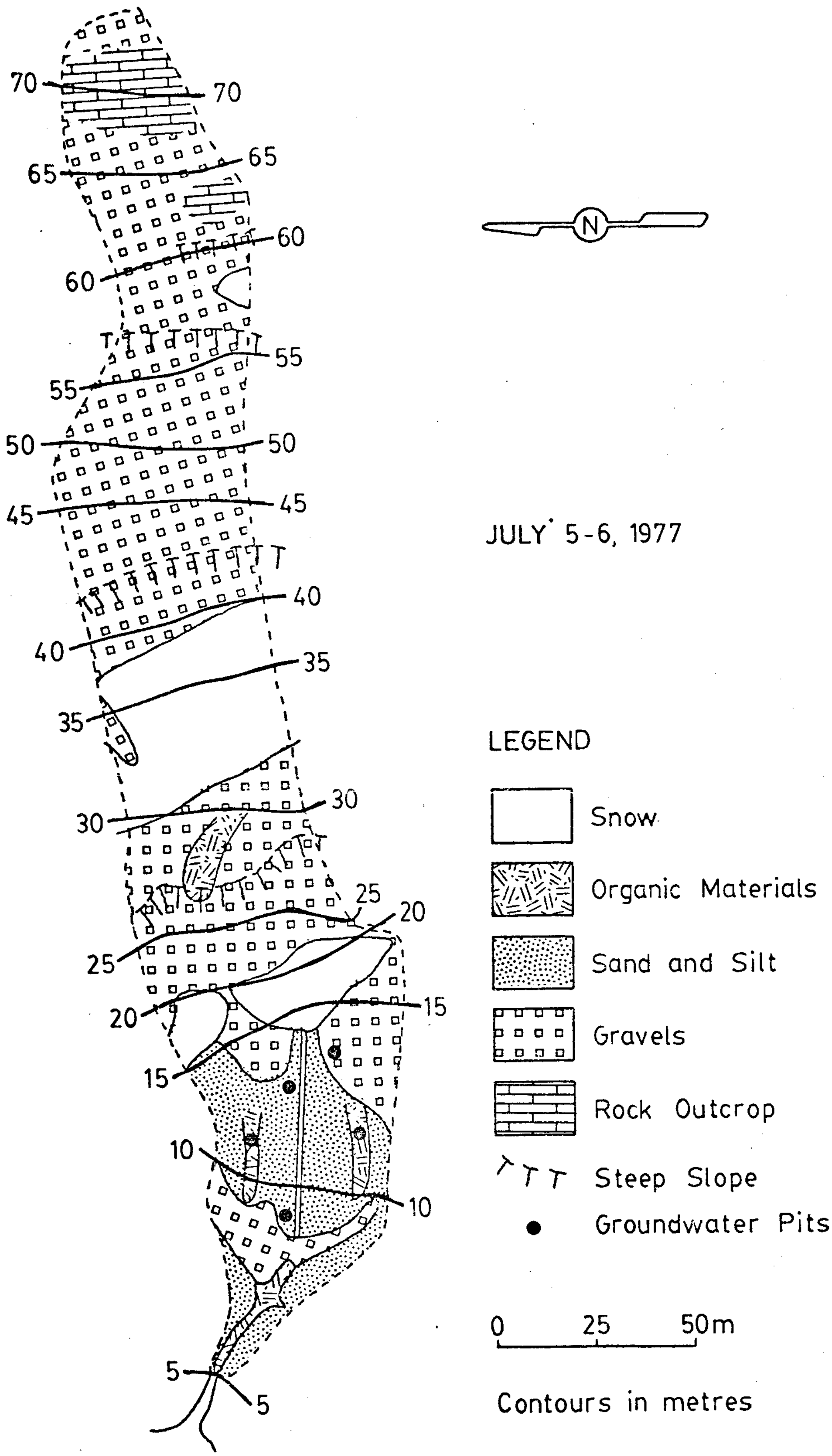


Figure 3.10 Surface features of the west-facing slope in the study basin. Location of the slope within the basin is shown in Figure 2.3.

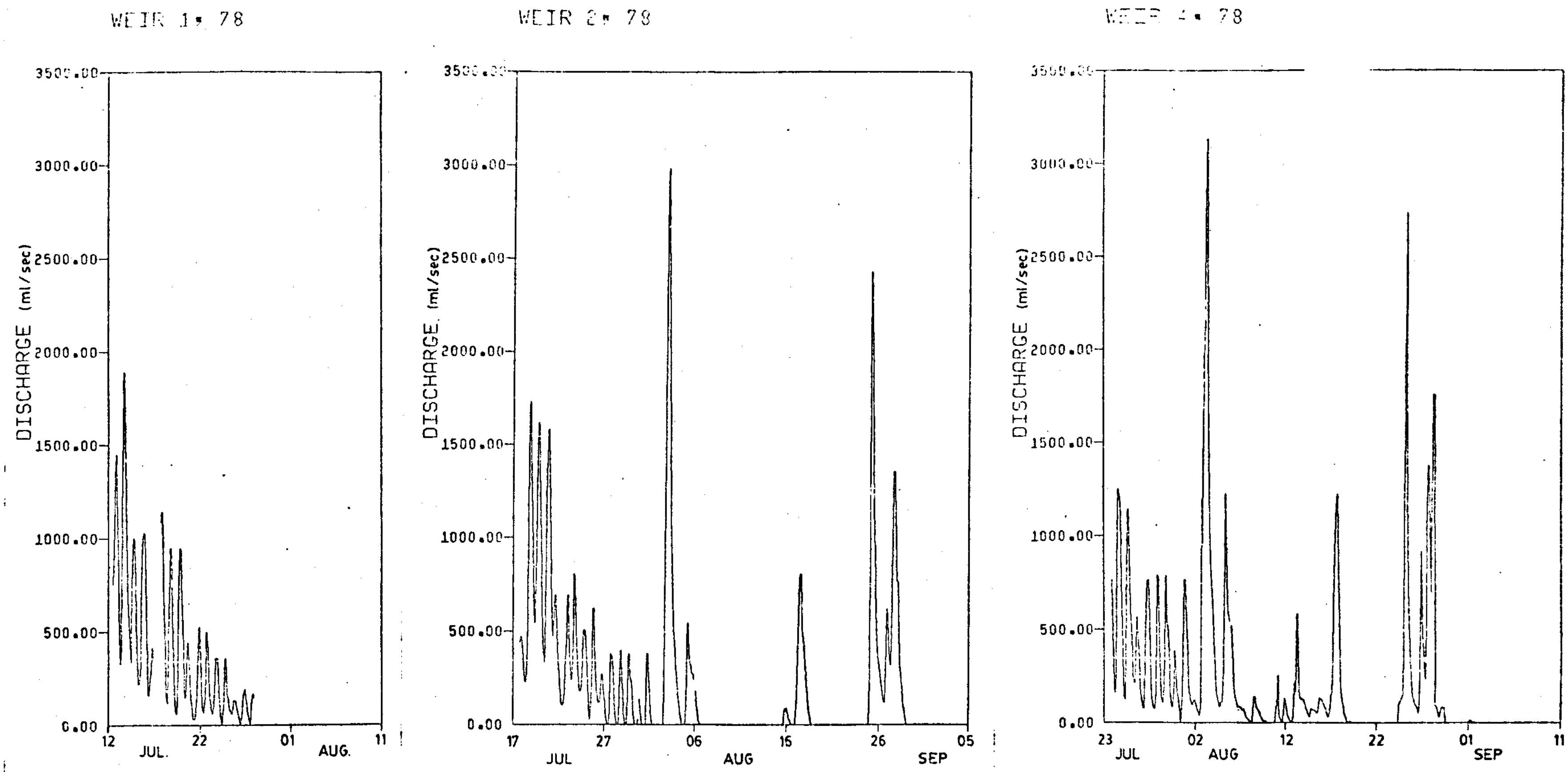


Figure 3.11 Discharge hydrographs for weirs 1, 2, and 4 in the study basin. Photos of each weir appear in Figure 3.8 and the locations within the basin in Figure 2.3.

Table 3.4

Daily Overland Flow Measured at Three Small Basins

Draining into Three Mile Lake

Date	Weir 1	Weir 2	Weir 4	Average
13 07 78	9.18	-	-	9.18
14	14.61	-	-	14.61
15	9.49	-	-	9.49
16	9.58	-	-	9.58
17	3.44	-	-	3.44
18	5.43	5.12	-	5.28
19	7.42	6.80	-	7.11
20	6.93	5.74	-	6.34
21	3.70	3.61	-	3.66
22	3.07	1.83	-	2.45
23	3.41	3.00	-	3.21
24	2.94	2.12	4.28	3.11
25	2.12	1.78	3.47	2.46
26	1.43	1.14	2.17	1.58
27	1.34	1.02	2.30	1.55
28	1.14	0.16	1.90	1.30
29	-	1.00	2.11	1.56
30	-	0.33	1.11	0.72
31	-	0.80	2.05	1.43
01 08 78	-	0.12	0.89	0.51
02	-	4.07	5.60	4.84
03	-	7.40	6.26	6.83
04	-	0.36	0.88	0.62
05	-	1.96	3.69	2.83
06	-	0.28	0.75	0.52
07	-	0	0.26	0.13

Table 3.4 (cont)

Date	Weir 1	Weir 2	Weir 4	Average
08 08 78	-	0	0.38	0.19
09	-	0	0.11	0.06
10	-	0	0.09	0.05
11	-	0	0.45	0.23
12	-	0	0.33	0.17
13	-	-	1.31	1.31
14	-	0	0.37	0.19
15	-	0.18	0.52	0.35
16	-	0.13	0.48	0.31
17	-	3.30	4.28	3.79
18	-	0.49	0.29	0.39
19	-	0	0	0
20	-	0	0	0
21	-	0	0	0
22	-	0	0	0
23	-	0	0	0
24	-	0.13	1.44	0.79
25	-	7.43	4.83	6.13
26	-	1.86	1.58	1.72
27	-	5.15	4.68	4.92
28	-	1.82	3.20	2.51
29	-	0	0.19	0.10
30	-	0	0	0
31	-	0	0	0
			TOTAL	127.55

All measurements in mm per unit area of overland flow basin lake storage; E_L is evaporation from the lake surface; Q_O is overland flow; P_L is direct precipitation onto the lake surface; and Q is outflow from the basin. For the period between the beginning of snowmelt and the commencement of lake outflow, overland flow was measured directly and the groundwater flow component cannot be separated from the overland flow component. During this period, equation (3.2) yields ΔS_G of 48.2 mm. which when applied to equation (3.1), results in a lumped Q_O and R_S term equal to 121.3 mm., after which time overland flow was measured independently. Equation (3.3) yields a groundwater flow value of 15.8 mm.

In detail, groundwater and frost table profiles from the south-facing slope for selected days are shown in Figure 3.12. Measurement points for "line 1" and "line 2" are shown in Figure 3.9. Daily fluctuations in groundwater and frost table are shown for both south- and west-facing slopes in Appendix Three. Pit locations are available from Figures 3.9 and 3.10 and the slope locations within the basin in Figure 2.3.

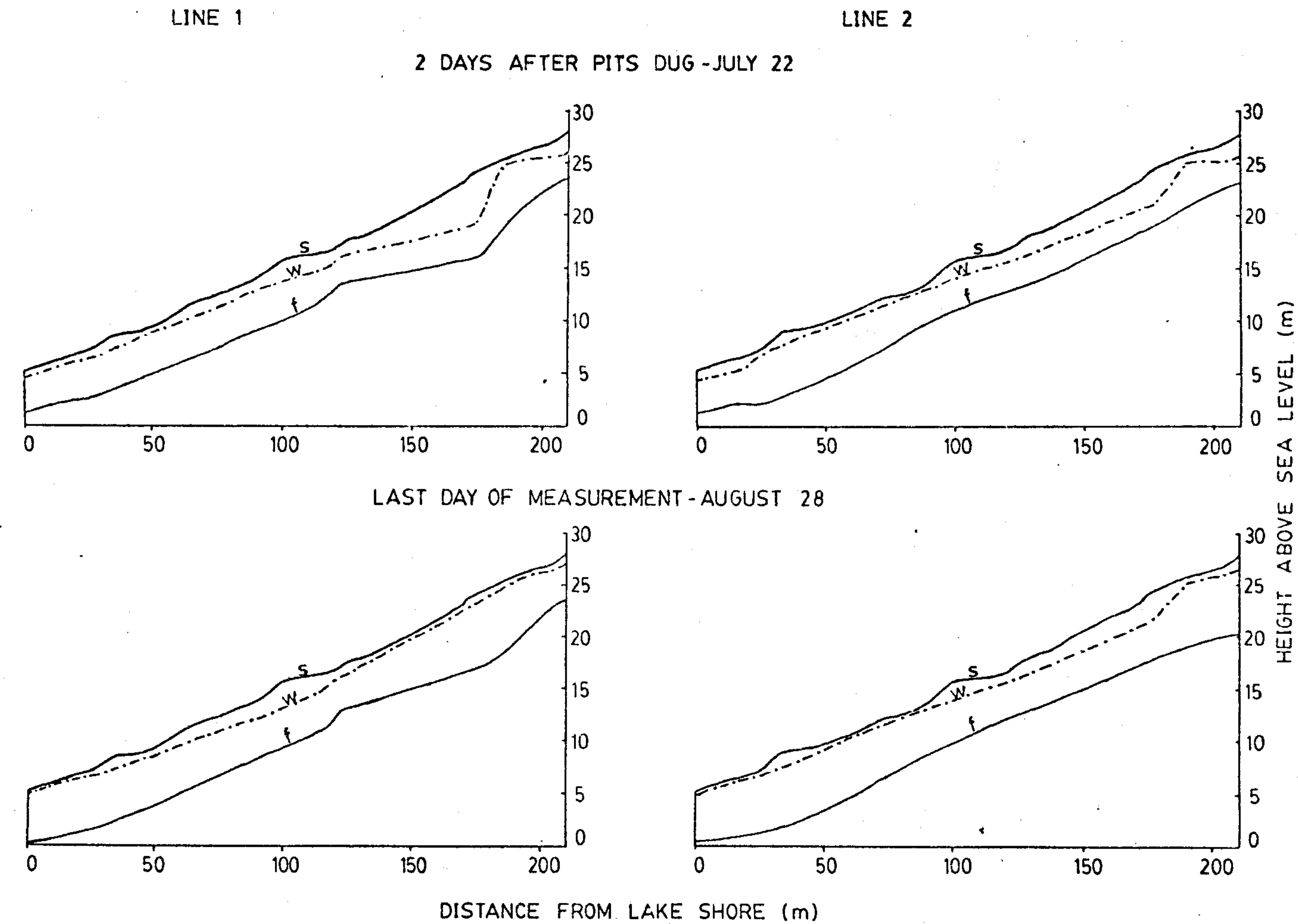


Figure 3.12 Surface (S), groundwater table (W), and frost table (f) profiles for selected days during the study period. "Line 1" consists of all groundwater pits west of an imaginary north-south through the pond in Figure 3.9; "Line 2" consists of all groundwater pits to the east. The southernmost pit is included in both profiles. Vertical exaggeration is 4 x for the surface profile and 10 x for the water and frost profiles.

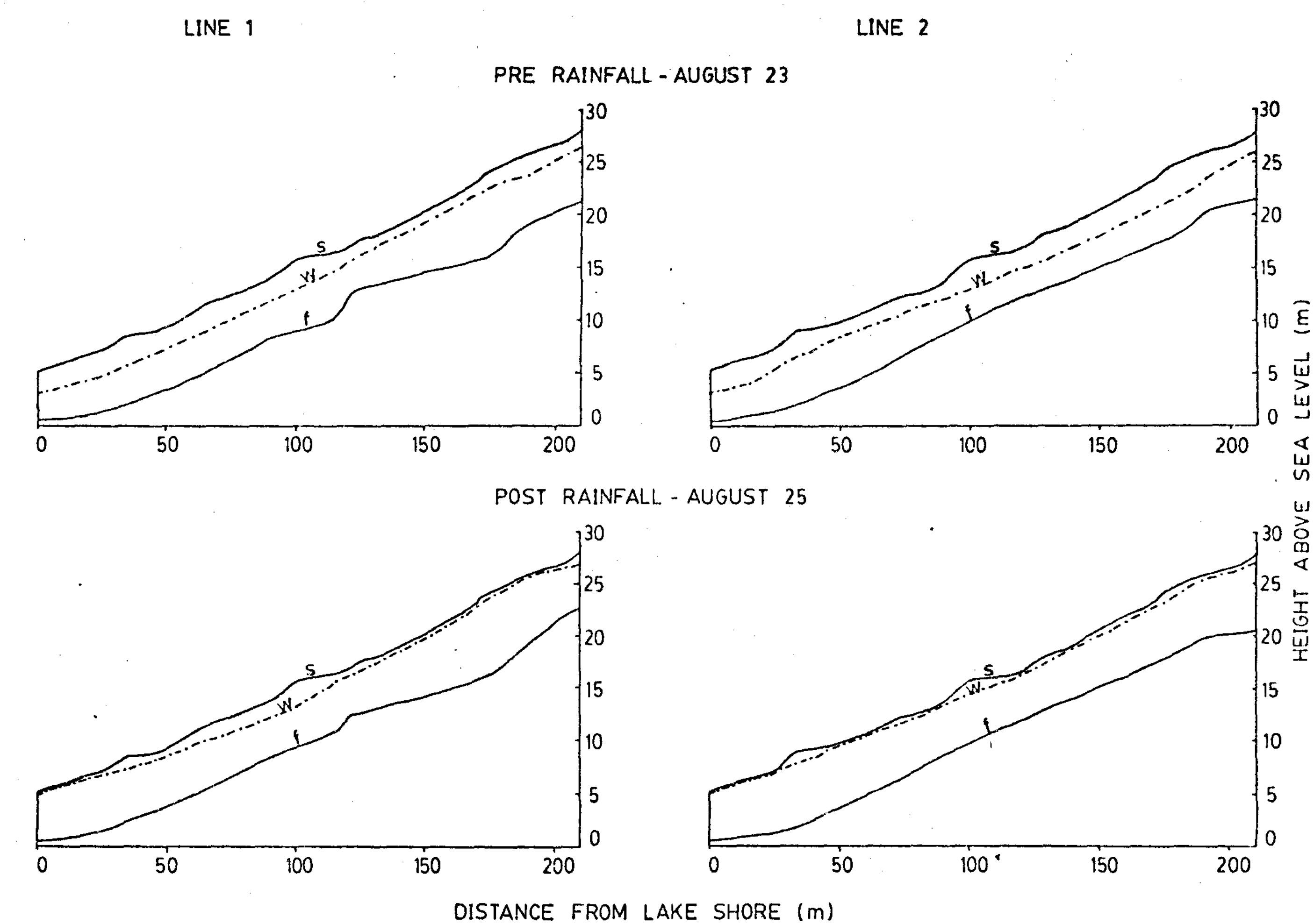


Figure 3.12 cont'd

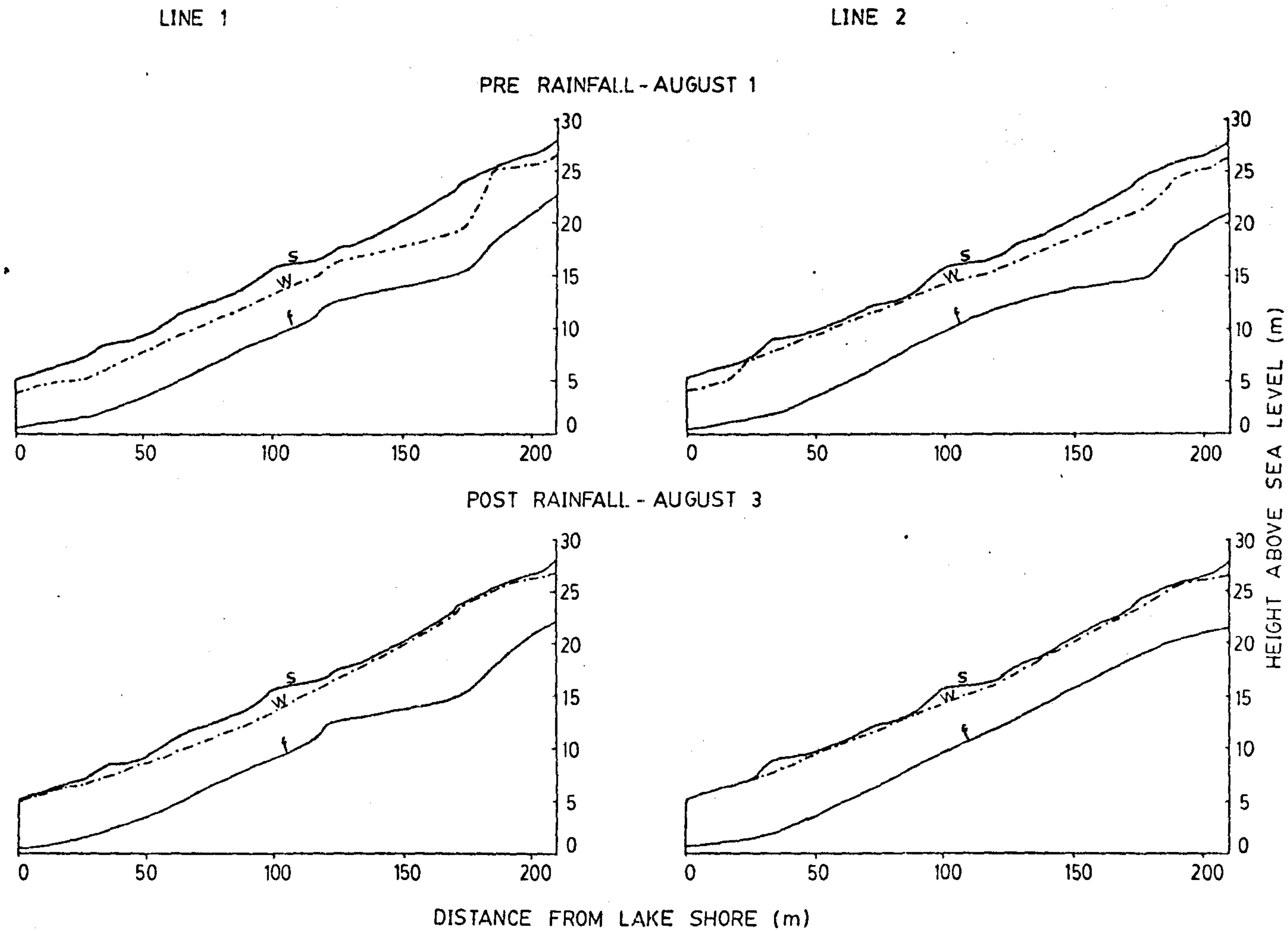


Figure 3.12 cont'd

CHAPTER 4

WATER OUTPUTS AND CHANGES IN STORAGE

Water outputs from the basin and from the lake must be examined separately, although the output components of each are similar. Outflow from the basin and outflow from the lake are identical and so will be analyzed in one section. Seasonal basin storage is divided into change in storage in the lake and change in storage of groundwater.

4.1 Water Outputs

4.1.1 Outflow

Both basin and lake outflow occurs at the outlet of Three Mile Lake . From July 19th when outflow began, until August 31st, total streamflow from the basin yielded 135.9 mm.

During the winter months, snowdrifts accumulate in the outflow channel of Three Mile Lake to the extent that when spring-melt begins the channel is completely snow-choked and outflow does not begin immediately. Snowmelt began on June 30th and by noon of July 19th the lake level had risen 1.27 m. By mid-afternoon slush moved along the surface of the snow in the channel but no channel was formed until early evening. Figure 4.1 shows the outflow channel several days before the break and in the evening of July 19th approximately five hours after the channel opened. Due to water depths and

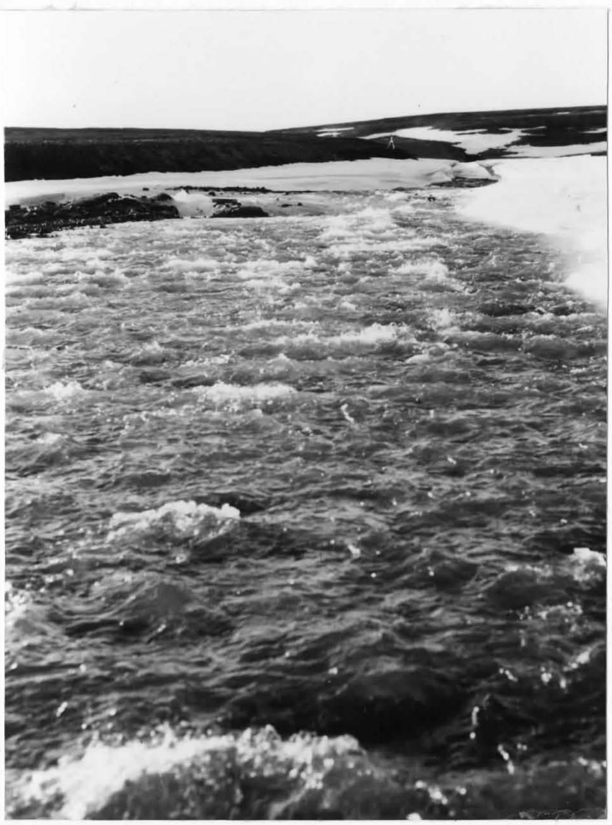


Figure 4.1 Top: The outflow channel several days before breaking (looking north). Bottom: View looking up the outflow channel approximately five hours after the break. Note the tripod (middle background) for scale.

high flow velocities at the start of the outflow period, current metering was not attempted and discharge was computed using the lake area and the lake level drop over time. Current metering began on the next morning and continued until a stage discharge rating curve was established. The hydrograph for the lake (Figure 4.2) illustrates the rapid rise and subsequent drop of the lake level. Minor fluctuations occurring in the rest of the season are due to major rainfall events; specifically the storms of August 2-3 and August 24-25. Smaller rainfall events do not significantly affect the hydrograph and flow remains low for the most part of the season. A lag time of approximately twenty-four hours is evident between a rainfall event and the resultant increase in discharge (Table 5.1). Precipitation falling directly onto the lake for individual storms was generally not significant (Table 3.3), so that immediate increases in discharge did not occur.

4.1.2 Evaporation From the Basin

Evaporation from gravel and bog was measured at a study site east of Three Mile Lake and evaporation over water was measured at the lake. Using equation (2.2) and a suitable α value in equation (2.3), evaporation from the three types of surfaces and for the entire study period is shown in Table 4.1.

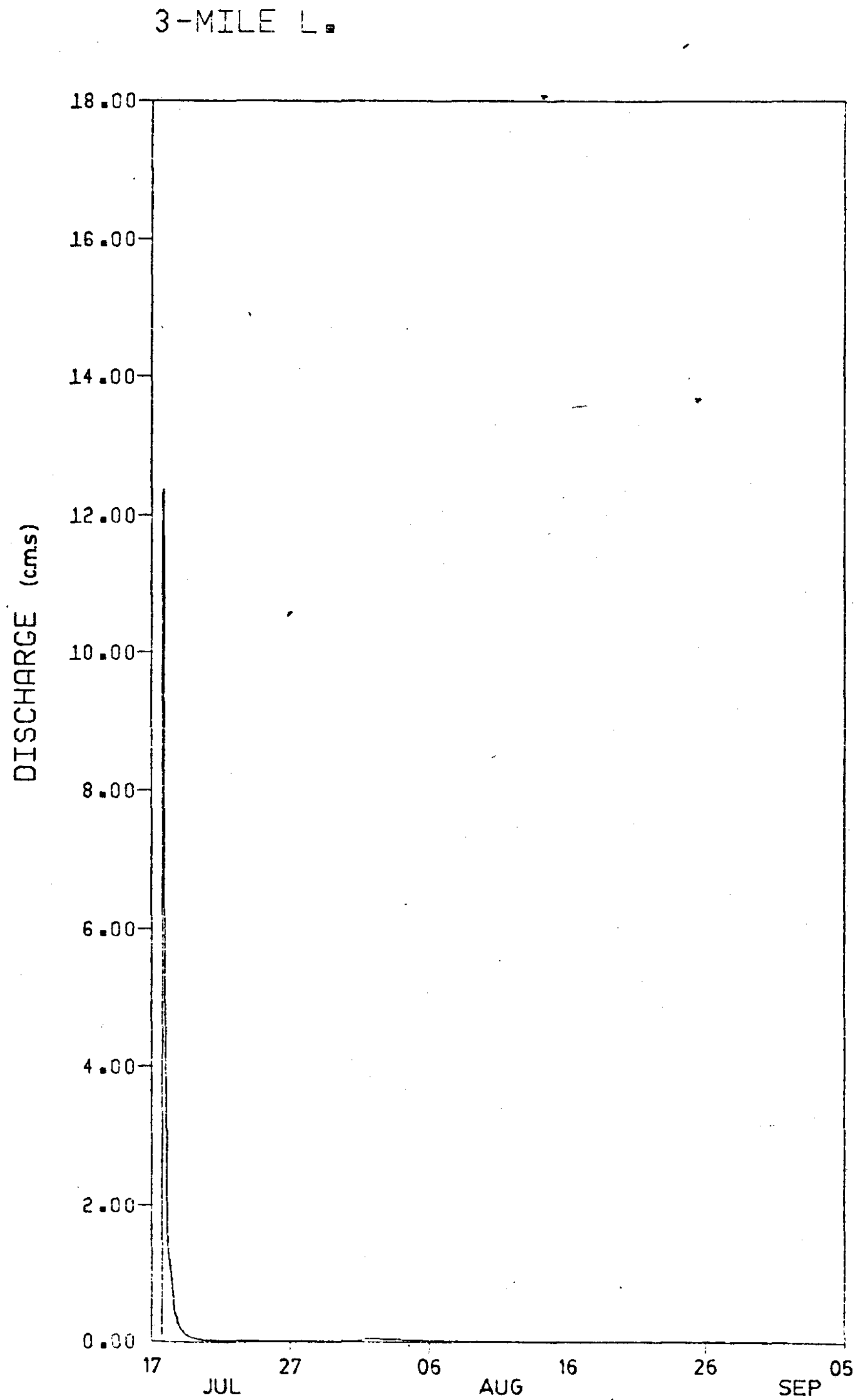


Figure 4.2 The discharge hydrograph for Three Mile Lake. Following the initial outflow, slight fluctuations during the season are attributed to major storms.

Table 4.1 Evaporation from Different Surface Types in the Study Basin

Surface Type	α	E (mm.)
Gravel	0.24-1.24	40.4
Bog	1.26	14.7
Lake	1.26	5.9

It has been shown (Priestley and Taylor, 1972; Stewart and Rouse, 1977) that an α of 1.26 adequately describes evaporation from continually wet surfaces. Snowmelt kept bog surfaces saturated until towards the end of July when some drying may have occurred. August was too wet for the bog area to dry significantly and so $\alpha = 1.26$ is an acceptable value. A daily α value for gravel was computed using soil moisture measurements with high α values occurring during snowmelt and after heavy rainfall events. Evaporation is an important component of the water balance during the snowmelt period when there is abundant moisture. Between June 30th and July 31st (the end of the snowmelt period) 64 percent of the total seasonal evaporation had occurred, the remaining 36 percent evaporating in August.

Taking the entire basin into consideration, evaporation for the study period totalled 61 mm., a value which is almost the same as summer precipitation. This result is similar to that of Brown et al. (1968), but should probably be treated as coincidental. Marsh (1978) in a study of four basins to the east of Three Mile Lake found evaporation to be significantly higher (12.9 percent to 112.9 percent) than rainfall during a summer considerably drier than 1978 (31 mm. summer precipitation vs. 63 mm.).

4.1.3 Evaporation from the Lake

Evaporation from the lake surface may be calculated using equation (2.2) and α = 1.26 in equation (2.3). This calculated value for water surface evaporation is adjusted for any ice cover present so that total daily lake evaporation is obtained as:

$$E_L = E_W \frac{(A_L - A_i)}{A_L} \qquad (4.1)$$

where A_L is lake area; A_i is area with an ice cover; and E_W is daily evaporation from the water surface. Total seasonal evaporation from the lake was 5.9 mm. as opposed to 6.3 mm. direct rainfall input. The magnitude of evaporation from snow and ice surfaces is small (Weller and Holmgren, 1974) and so may be ignored for the purposes of this study. Lake ice did not disappear completely until August 11th (Figure 3.7). The remaining ice cover reduced the area of lake water surface, hence preventing evaporation from the entire lake. However, rainfall onto the ice surface still reached the lake.

Measured evaporation and precipitation at Three Mile Lake are plotted in Figure 4.3 for the period that the

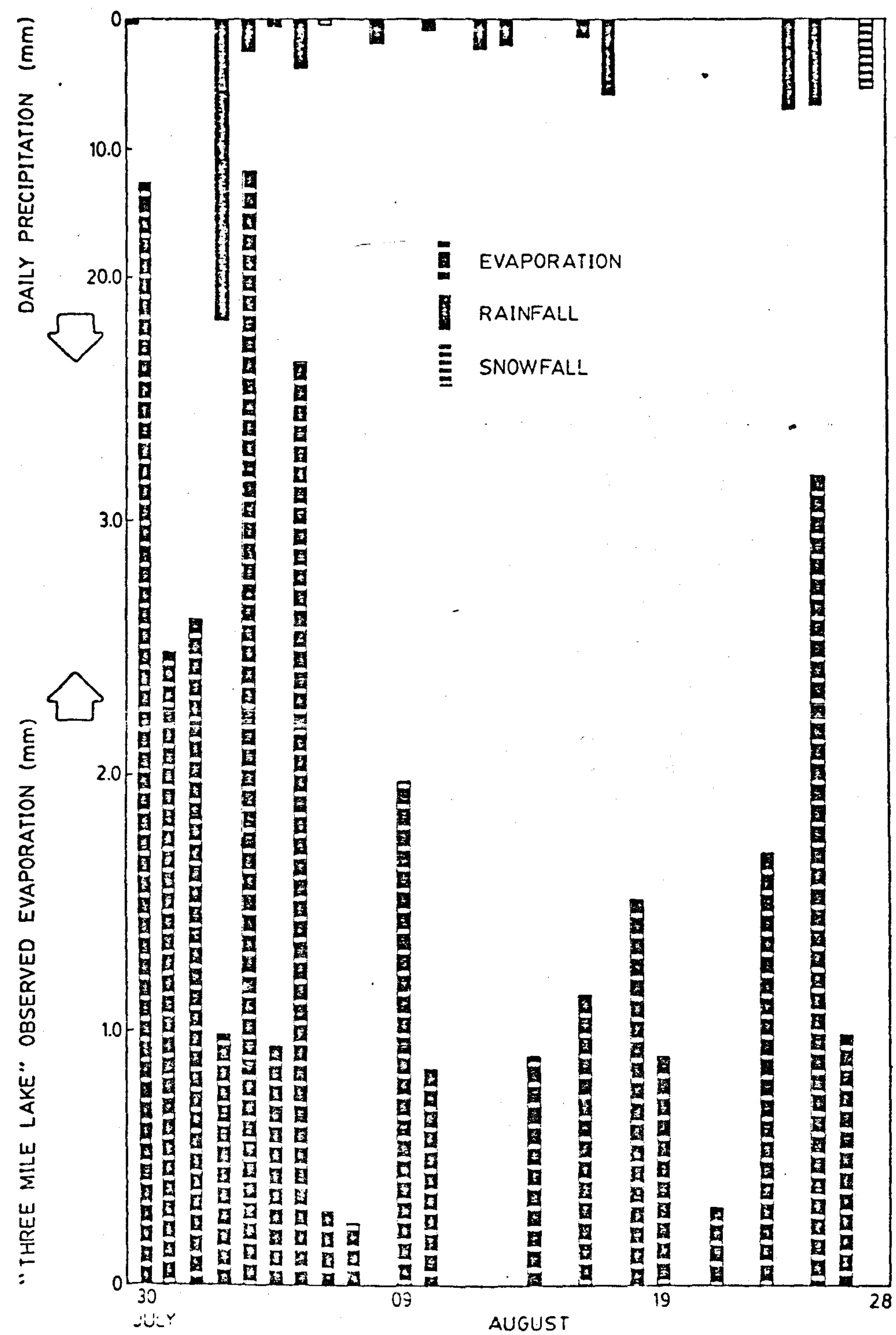


Figure 4.3 Observed evaporation and precipitation at "Three Mile Lake". Gaps in the evaporation plot indicate missing data. Precipitation type is distinquished.

evaporation pan was monitored. Gaps in the evaporation plot indicate missing data. Figure 4.4 plots observed evaporation from Three Mile Lake and A.E.S. pan measurements against calculated evaporation using equation (2.3) and α = 1.26. Divergence from the 1:1 line follows a consistent pattern and is due to different thermal regimes at each pan site. Both sets of pan data, however, showed definite linear trends when compared with the calculated values, suggesting that the relative magnitude of daily evaporation shown by all three types of information are similar.

4.2 Changes in Storage

4.2.1. Changes in Basin Storage (ΔS_B)

Total storage changes are calculated as a residual of the water balance equation (1.1). For the study period between June 30th and August 31st, net change in basin storage was 62.8 mm. (Table 5.1). This figure may be partitioned between lake storage (ΔS_L) and groundwater storage (ΔS_G).

4.2.2 Changes in Lake Storage

Changes in lake storage are easily calculated using the following equation:

$$\Delta S_L = \frac{\Delta L}{\Delta t} \cdot (0.097) \qquad (4.2)$$

where $\Delta L/\Delta t$ is the change in lake level with time (in metres); and 0.097 is a proportionality constant representing the lake

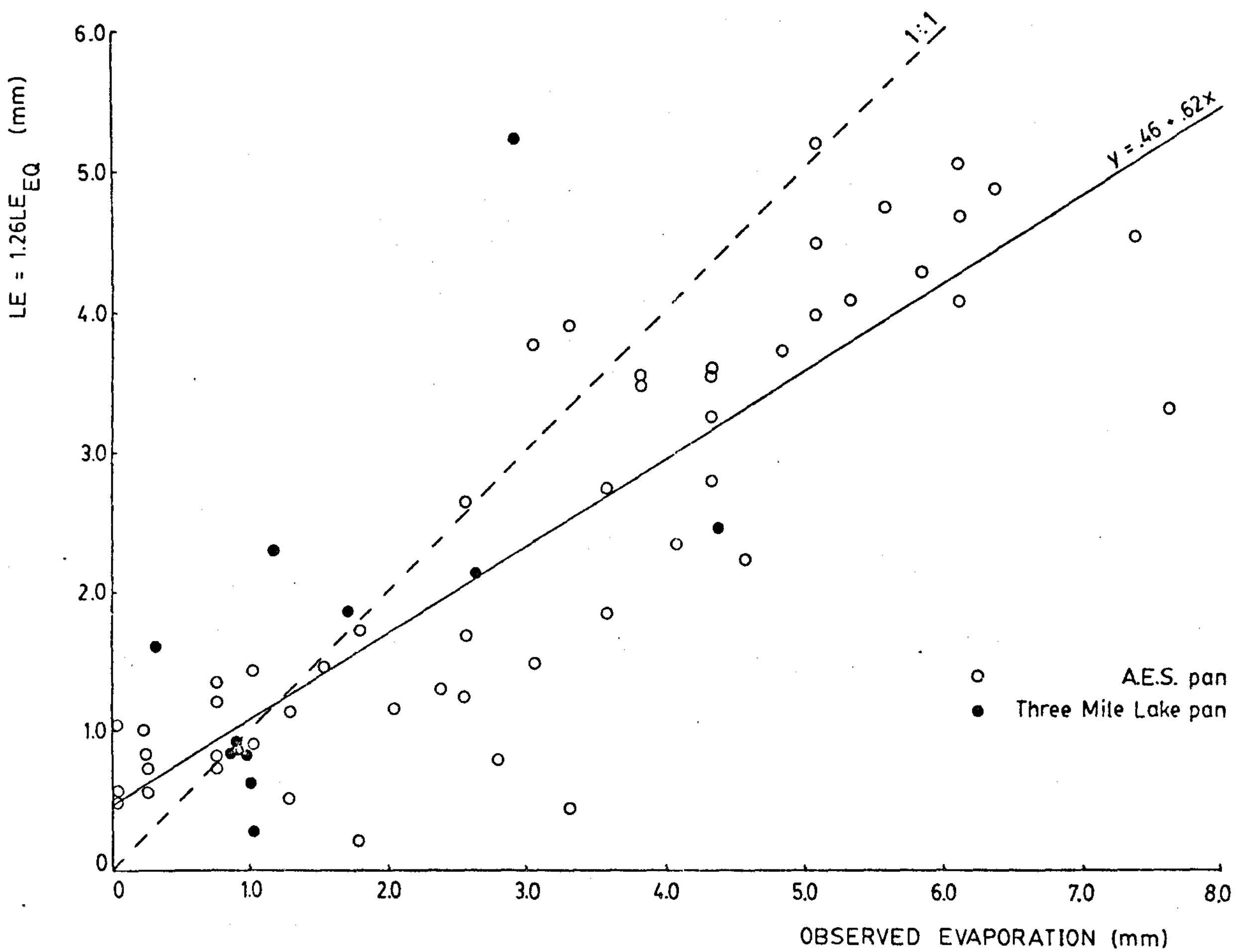


Figure 4.4 The relationship between observed A.E.S. and Three Mile Lake pan evaporation and evaporation calculated from equation 2.3.

area:basin area ratio. For the purposes of this study $\Delta L/\Delta t$ was calculated from 2400 hours on day i to 2400 hours on i + 1 and the proportionality constant converts the result into a depth in metres over the entire basin. Daily values for changes in lake storage are presented in Table 5.1, the seasonal total being 5.0 mm.

4.2.3 Changes in Groundwater Storage

Changes in groundwater storage are calculated as a residual of the previous two storage terms as follows:

$$\Delta S_G = \Delta S_B - \Delta S_L \tag{4.3}$$

where ΔS_G is change in groundwater storage; ΔS_B is change in net basin storage; and ΔS_L is change in lake storage. Daily values of ΔS_G appear in Table 5.1. It should be noted that until the snowpack is entirely melted (July 31st) this term also encompasses changes in meltwater storage within the snowpack. For the period beginning snowmelt (June 30th) until the day before the outflow channel break (July 18th) cumulative changes in storages are shown in Figure 4.5. Daily values are presented in Table 5.1; the seasonal total being 57.8 mm.

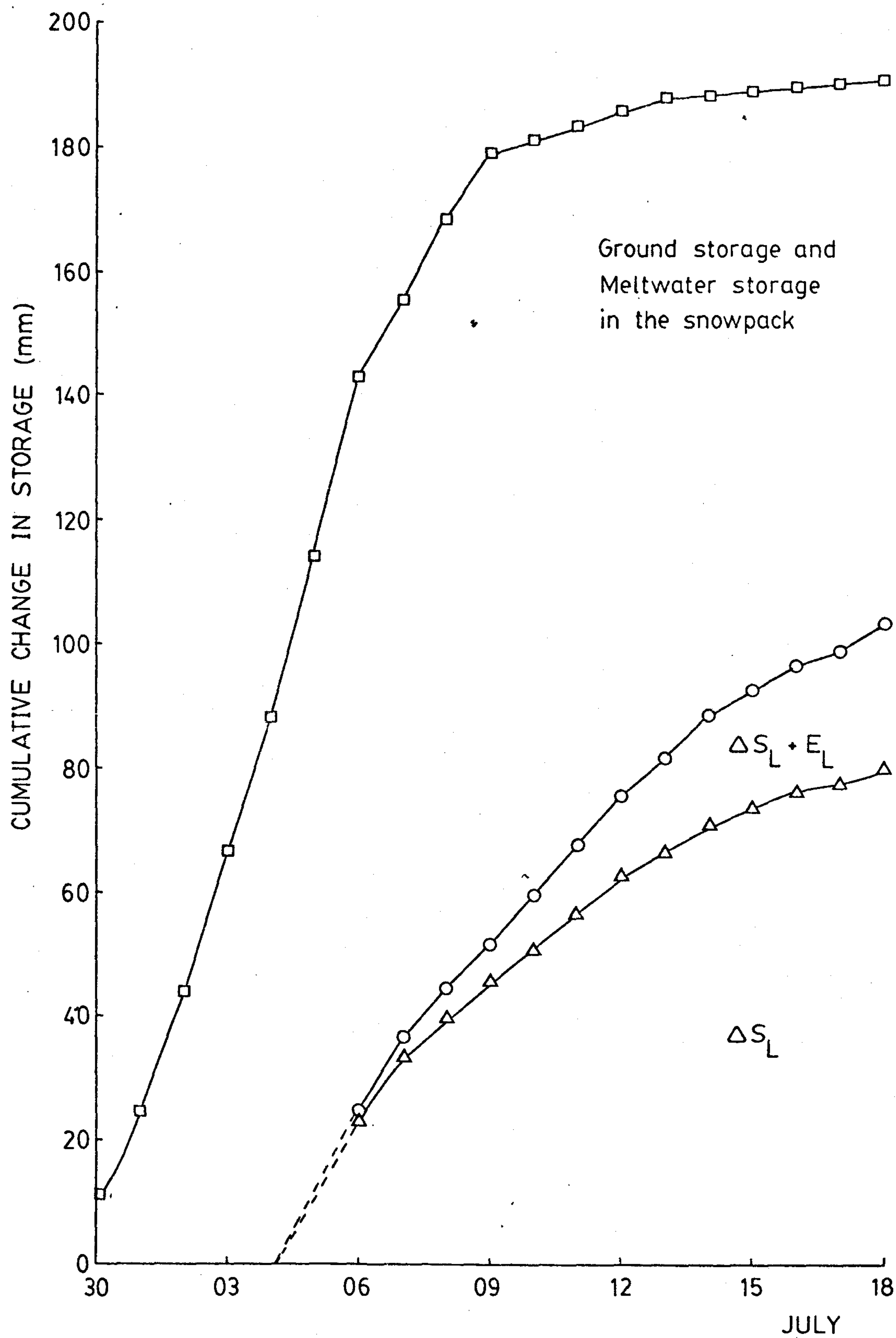


Figure 4.5 Cumulative changes in storage for the lake, the ground, and the snowpack during the snowmelt period. E_L is evaporation from the lake.

CHAPTER 5

WATER BALANCE OF THREE MILE LAKE

Previous chapters have described the magnitudes of various components of the waterbalances for the basin and for Three Mile Lake. This chapter will relate the magnitude of these components to each other and compare the results with other water balance studies.

5.1 Seasonal Water Balance for the Basin

Daily values for each component of the basin water balance equation are presented in Table 5.1. The change in basin storage (ΔS_B) is computed as a residual of the input and output terms. Results of the basin water balance are summarized in Table 5.2.

These results are consistent with those of other water balance studies reported by Marsh (1978). In these studies reported and where precipitation is distinguished into snowfall and rainfall, the mean snow contribution is 78.6 percent of total precipitaion. Values range between 57 percent and 88 percent. For rainfall, the mean contribution is 20.5 percent, the range being 10 percent to 43 percent. This illustrates the dominance of precipitation as snowfall in high Arctic areas. Discharge removes a mean value of 68.3 percent of the total water input but the range is extreme; 17 - 100 percent. Likewise, the range of evaporation

TABLE 5.1

Daily Values For Components of the Water Balance Equation

Date	Snowmelt	Precip.	Evap.	Discharge	ΔS_B	ΔS_L	ΔS_G
June 28/78	0	-	-	-	-		
29	0	-	-	-	-		
30	10.6	-	-	-	10.6		
July 01/78	13.8	-	-	-	13.8	22.9	65.0
02	19.3	-	-	-	19.3		
03	22.7	-	-	-	22.7		
04	21.6	-	.1	-	21.5		
05	26.3	-	.2	-	26.1	13.2	12.9
06	28.8	-	1.4	-	27.4	17.0	10.4
07	12.5	-	1.6	-	10.9	18.9	- 8.0
08	13.0	-	1.6	-	11.4	9.3	2.1
09	10.9	-	1.6	-	9.3	5.7	3.6
10	2.0	-	2.2	-	- .2	5.6	- 5.8
11	1.9	-	2.4	-	- .5	5.7	- 6.2
12	2.8	-	2.0	-	.8	5.7	- 4.9
13	2.1	-	2.0	-	.1	4.0	- 3.9
14	.55	-	2.4	-	- 1.9	4.5	- 6.4
15	.55	.8	1.4	-	- .5	2.8	- 2.85

TABLE 5.1 (cont'd)

Date	Snowmelt	Precip.	Evap.	Discharge	ΔS_B	ΔS_L	ΔS_G
July 16/78	.55	-	1.7	-	- 1.2	2.5	- 3.7
17	.55	-	1.2	-	- .7	2.1	- 2.8
18	.55	-	1.7	-	- 1.2	2.5	- 3.7
19	.55	-	2.1	75.0	-76.6	-75.0	- 1.6
20	.55	-	2.1	28.6	-30.2	-28.6	- 1.6
21	.55	-	.8	5.0	- 5.3	- 6.8	1.5
22	.55	-	1.6	1.8	- 2.9	- 3.0	.1
23	.55	-	1.4	1.1	- 2.0	- 1.3	- .7
24	.55	.3	1.3	.9	- 1.4	- .5	- .9
25	.55	-	1.4	.8	- 1.7	- .8	- .9
26	.55	-	.9	.7	- 1.1	- 1.0	- .1
27	.55	-	1.0	.6	- 1.1	- .6	- .5
28	.55	-	1.2	.5	- 1.2	- .4	- .8
29	.55	.6	.6	.5	- .1	- .4	.3
30	.55	-	1.0	.5	- 1.0	- .3	- .7
31	.55	-	1.2	.5	- 1.2	- .1	- 1.1
Aug. 01/78	-	-	.5	.4	- .9	- .2	- .7
02	-	21.9	.7	.4	20.8	4.4	16.4
03		2.4	.8	2.0	- .4	5.1	- 5.5

TABLE 5.1 (cont'd)

Date	Snowmelt	Precip.	Evap.	Discharge	ΔS_B	ΔS_L	ΔS_G
Aug. 04/78	-	.6	1.0	1.9	- 2.3	- 1.4	- .9
05	-	4.0	1.4	1.7	.9	- .5	1.4
06	-	.3	.6	1.4	- 1.7	- 1.1	- .6
07	-	-	.8	1.1	- 1.9	- .9	- 1.0
08	-	1.3	.9	.9	- .5	- .7	.2
09	-	-	1.3	.7	- 2.0	- 1.3	- .7
10	-	1.0	.8	.5	- .3	- .8	.5
11	-	-	1.3	.5	- 1.8	- .9	- .9
12	-	2.1	.4	.4	1.3	0	1.3
13	-	1.9	.9	.4	.6	- .6	1.2
14	-	-	.6	.4	- 1.0	- .4	- .6
15	-	-	.8	.4	- 1.2	- .2	- 1.0
16	-	1.3	.6	.4	.3	- .1	.4
17	-	5.9	.7	.4	4.8	2.1	2.7
18	-	-	1.3	.4	- 1.7	.3	- 2.0
19	-	-	.8	.4	- 1.2	- .9	- .3
20	-	-	.7	.4	- 1.1	- .7	- .4
21	-	-	.4	.3	- .7	- .5	- .2
22	-	-	.5	.2	- .7	- .4	- .3
23	-	-	.5	.2	- .7	- .3	- .4

TABLE 5.1 (cont'd)

Date	Snowmelt	Precip.	Evap.	Discharge	ΔS_B	ΔS_L	ΔS_G
Aug. 24/78	-	6.4	.4	.2	5.8	.7	5.1
25	-	6.2	.8	.5	4.9	2.8	2.1
26	-	-	.5	.5	- 1.0	.2	- 1.2
27	-	5.6	.2	.6	4.8	1.1	3.7
28	-	-	.4	.6	- 1.0	.9	- 1.9
29	-	-	.2	.5	- .7	- 2.7	2.0
30	-	-	.2	.4	- .6	- 1.0	.4
31	-	-	.1	.3	- .4	- .6	.2
TOTALS	198.2	62.6	61.2	135.9	62.8	5.0	57.8

TABLE 5.2

Water Balance For Three Mile Lake Drainage Basin

INPUTS		OUTPUTS		CHANGE IN STORAGE
Precipitation (mm)		Evaporation (mm)	Discharge (mm)	
Winter Snow	Summer Snow and Rain			
198.2	62.6	61.2	135.9	62.8
(76%)	(24%)	(23.5%)	(52.1%)	(24.1%)
100%		100%		

is great (1 - 64 percent), but the contribution obtained in this study is consistent with the mean of 29.2 percent.

The importance of evaporation in the water balance of high Arctic basins has been reported by other authors (ie. Addison, 1972; Woo, 1976; Marsh and Woo, 1977). The results of this study re-affirm these conclusions. Although evaporation removes only 23.5 percent of the total water input to the basin, the magnitude of evaporation equalled 97.8 percent of total summer precipitation. This observation is consistent with that of Marsh and Woo (1977) who reported a 91 percent loss of summer precipitation to evaporation over six weeks; Brown et al. (1968) who observed that summer precipitation approximately equalled summer evaporation; and Marsh (1978) whose results indicate that evaporation removes more water than summer precipitation.

The change in storage term (ΔS_B) is often assumed to be zero over a season (ie. Marsh, 1978; Marsh and Woo, 1977), however, for the present study this is not so. The positive storage at the end of the season may be attributable to i) an actual increase in basin storage of 62.8 mm., assuming that all the input and output components of the basin water balance equation were measured accurately, or ii) an error in measurement of one or more of the components of the water balance, or iii) a combination of both possibilities.

Snow storage in the basin was adjusted for new snowfall after the survey was completed. It is possible that an over-adjustment was made. However, this is not very likely since A.E.S.-recorded snowfall was used in the adjustment and weather station records are already considered to be underestimates of snowfall (Woo and Marsh, 1978).

The summer precipitation measurements are considered to be accurate. Rain-gauges were emptied following each precipitation event and the results of three methods of computing basin rainfall are all within 1 percent of each other.

Evaporation over the study period is assumed accurate as calculated using equations (2.2) and (2.3). This approach has been successfully applied by other authors (ie. Woo, 1976; Rouse et al., 1977).

Some error in discharge calculations may occur over the initial discharge period. As previously mentioned, current metering did not commence until the day following the breakup. Shortly after the channel opened the stage recorder grounded and had to be moved further into the lake. Consequently discharge was computed using the drop in lake level as measured on a staff gauge. When reading the staff gauge, errors of up to 1.0 cm. could result from wave action. This error could be compounded at the next reading if wave conditions persisted. Even so, it is estimated that the magnitude of this error would be of the order of only 2 - 3 mm. over the entire basin and it should be realized that, due to the source of the error(wave action), the result could be either an increase or a decrease in discharge over the season. It should also be noted that this source of error is only applicable for the one day that current metering could not be accomplished.

The previous discussion suggests that errors in the water balance terms are minor indicating that the change in storage term is reasonably accurate. Qualitatively, this is supported by field observations. The 1977 field season was comparatively dry (Marsh, 1978), suggesting that at the time of freeze-up little water remained in the active layer. The 1978 field season (when this study was carried out) was wet and the active layer could not dry out sufficiently before freeze-up occurred. Hence, at the end of the field season (August 28, 1978), suprapermafrost groundwater storage was quite large (Table 5.3). However, these measured values of groundwater storage still fall short of the computed change in groundwater storage ($\triangle S_G$) of 57.8 mm. for the 1978 field season. The possible sources of error are snow storage determination (from the snow survey) and discharge measurements (complicated by the channel breakup) because evaporation and summer precipitation are likely to have been accurately determined.

TABLE 5.3

Groundwater Storage in the Active Layer Determined At the End of the 1978 Field Season

Slope	Pit	Material	Specific Yield	Saturated Depth (mm)	Water Storage (mm)
West-facing	#1	silt	.06	493	29.6
	2	silt	.06	578	34.7
	3	gravel	.03	552	16.6
	4	gravel	.03	579	17.4
	5	gravel	.03	499	15.0
South-facing	#1	silt	.06	518	31.1
	2	silt	.06	511	30.7
	3	silt	.06	570	34.2
	4	silt	.06	557	33.4
	5	silt	.06	452	27.1
	6	gravel	.03	362	10.9
	7	gravel	.03	375	11.3
	8	gravel	.03	660	19.8
	9	gravel	.03	573	17.2
	10	gravel	.03	458	13.7
	11	gravel	.03	650	19.5
	12	gravel	.03	425	12.8
	13	gravel	.03	315	9.5
	14	gravel	.03	400	12.0
	15	silt	.06	370	11.1

The lake storage term may be slightly underestimated. In 1977, outflow from the lake ceased on August 29th but subsurface flow further lowered the lake level slightly (Woo et al., 1977). In 1978 outflow had not ceased by the end of the first week in September. This indicates a positive lake storage term over the season which is, in fact, the case. However, the lowest lake level read on the last day of August 1977 was 13.1 cm. above an arbitrary datum. At the end of the flow season in 1978, the lowest lake level was at 21.3 cm. above the same datum. This gives a lake level difference of 8.2 cm. which converts to 7.8 mm. when spread over the basin. Thus, the value of ΔS_L = 5.0 mm. (Table 5.1) is an underestimate.

5.2 Lake Water Balance

Seasonal lake water balance is calculated from equation (1.3); numerical values for the various components are: P_L = 6.28 mm., $Q_O + R_S$ = 138.4 mm. for the entire season, E_L = 5.9 mm., Q = 135.9 mm., and ΔS_L = 5.0 mm. The result is a seasonal lake water balance of -2.12 mm. In view of the small magnitude of the residual term, the results are considered to be satisfactory.

To assess the magnitudes of the lake water balance components it is necessary to distinguish two time periods: the pre-channel breakup period and the period that followed. Table 5.4 shows the magnitudes of each component for each time period.

TABLE 5.4

Components of the Water Balance for Three Mile Lake

Component	P_L	Q_0	R_S	E_L	Q	S_L
Pre-channel breakup	0.08	121.3		1.09	0	122.4
Post-channel breakup	6.2	1.3	15.8	4.8	135.9	-117.4

All values are shown in mm. units.

5.2.1 The Pre-Channel Breakup Period

During this two and one-half week period (July 1 to July 19) snowmelt was the dominant process, giving rise to overland flow and subsurface flow on the slopes. Of the total water input during this period, precipitation supplied less than 0.1 percent. Outflow from the lake has not yet begun and so any water input to the lake added to an increase in the ΔS_L term. Removal of water was entirely due to evaporation which was insignificant (less than 1.0 percent of total water input) because of an extensive ice cover over the lake.

5.2.2 The Post-Channel Breakup Period

Channel breakup produced a sudden release of water from the lake, and this began on July 19th. By the end of the second day after the breakup 76 percent of the total outflow had already left the lake. By the end of July, this figure increased to 85.7 percent. During the same time interval, however, rainfall only accounted for 14.5 percent of total precipitation.

In the post breakup period, subsurface flow provided 67.8 percent of the total water input to the lake; summer precipitation 26.6 percent; and overland flow 5.6 percent. This indicates the importance of groundwater flow in providing input to the lake. Marsh and Woo (1977) found that, for a pond on Ellesmere Island during a two-week period in July, the groundwater contribution was 89 percent of the water input. Examination of Appendix Three reveals considerable fluctuation in groundwater levels at Three Mile Lake over the summer period. It is also apparent that, with the exception of heavier rainstorms, precipitation during this period was stored in the active layer. The storm of August 2-3 raised groundwater levels above the ground to produce overland flow at most observation pits. Following this storm, levels declined despite occasional periods of rainfall. Water levels did not rise again until another major rainstorm arrived on August 12-13.

During the post channel breakup period, evaporation from the lake accounted for 77.4 percent of the precipitation onto the lake, again illustrating the importance of this process in the water balance of High Arctic basins.

5.2.3 Lake Water Balance Summary

For the entire summer period, overland flow and subsurface flow were the most important terms, together providing more water than could be accounted for by lake outflow. During the pre-channel breakup period 87.6 percent of the total was contributed by snowmelt, thus demonstrating the importance of this process. Evaporation from the lake surface removes 94 percent of direct summer precipitation; a figure consistent not only with the results of this study, but also similar to the results of other studies carried out in a High Arctic environment.

CHAPTER 6

SUMMARY OF FINDINGS

Despite the abundance of lakes in the Canadian Arctic, few comprehensive attempts have been made to evaluate all the major components of the water balance of basins containing lakes. It was, therefore, the objective of this dissertation to study all components of the water balance of a basin containing a small lake in the High Arctic and to evaluate the relative magnitudes of these components.

For a snow-dammed lake, accurate determination of the snow stored in the basin prior to melt is important because, until such time as the outflow channel opens and discharge begins, a substantial proportion of snowmelt runoff produces large and rapid lake level rises. Outflow is significant for only a few days following the channel breakup, by which time outflow decreases to low but steady levels, increasing noticeably only following large rainfall events. Evaporation is an important process, removing almost as much water as the summer precipitation input. In this study, evaporation is more important during the snowmelt period when there was abundant surface water and very little rainfall. The condition of the active layer during freeze-up is important because this exerts an influence on the amount of water which may later be stored during snowmelt. It is this factor which largely determines the surplus or deficit of the basin water balance.

Previous studies have suggested that the change in storage term approaches zero on an annual basis. Results of this study indicate that this is not always the case and suggest that in future studies greater attention should be paid to the calculation of this term in annual water balances.

REFERENCES

Addison, P.A., 1972. Studies on Evapotranspiration and Energy Budgets on the Truelove Lowland, Devon Island, N.W.T. In Devon Is. IBP Project, High Arctic Ecosystem Project Report 1970-71, Bliss, L.C., ed., 73-88.

Ambler, D.C., 1974. Runoff from a Small Arctic Watershed. In Permafrost Hydrology, Proc. Workshop Sem. Can. Natl. Comm. Intl. Hydrol. Decade, 45-49.

Anderson, J.C., 1974. Permafrost - Hydrology Studies at Boot Creek and Peter Lake Watersheds, N.W.T. In Permafrost Hydrology, Proc. Workshop Sem. Can. Natl. Comm. Intl. Hydrol. Decade, 39-44.

Bostock, H.S., 1970. Physiographic Subdivisions of Canada. In Geology and Economic Minerals of Canada, Douglass, R.J.W., ed., Geol. Surv. Can. Econ. Report 1, Information Canada, Ottawa, 10-30.

Brown, J., Dingman, S.L., and Lewellen, R.I., 1968. Hydrology of a Drainage Basin on the Alaskan Coastal Plan. U.S. Army Corps of Engin., CRREL Research Report 240.

Church, M.A., 1974. Hydrology and Permafrost with References to Northern North America. In Permafrost Hydrology, Proc. Workshop Sem. Can. Natl. Comm. Intl. Hydrol. Decade, 7-20.

Cook, F.A., 1967. Fluvial Processes in the High Arctic. Geogr. Bull.,9, 262-268.

Cruikshank, J., 1971. Soil and Terrain Units Around Resolute, Cornwallis Is. Arctic, 24, 195-209.

Davies, J.A. and Allen, C.D., 1973. Equilibrium, Potential and Actual Evaporation from Cropped Surfaces in Southern Ontario. J. Appl. Met., 12, 649-657.

Dept. of Env., 1971. Temperature and Precipitation 1941-1970, The North - Y.T. and N.W.T., Downsview, 24.

Dingman, S.L., 1973. Effects of Permafrost on Streamflow Characteristics in the Discontinuous Permafrost Zone of Central Alaska. In Permafrost: The North American Contribution, Second Intl. Conf. Natl. Academy of Sciences, Washington, 447-453.

Gray, D.M., ed., 1970. Handbook on the Principles of Hydrology. Publ. The Secretariat, Can. Natl. Comm. Intl. Hydrol. Decade. Copyright 1970 by N.R.C. of Canada.

Hare, F.K., and Hay, J.E., 1971. Anomalies in the Large-Scale Annual Water Balance over Northern North America. Can. Geogr., 15, 79-94.

Hartman, C.W., and Carlson, R.F., 1972. Water Balance of a Small Lake in a Permafrost Region. University of Alaska, Inst. Water Resources, Rpt No. 1WR-42.

Kane, D.L. and Carlson, R.F., 1973. Hydrology of the Central Arctic River Basins of Alaska. University Alaska College, Inst. Water Resources, Rpt. No. 1WR-41.

Linsley, R.K., Kohler, M.A. and Paulhus, J.L.H., 1975. Hydrology for Engineers. McGraw Hill, New York, 482.

Mackay, D.K. and Løken, O.H., 1974. Arctic Hydrology. In Arctic and Alpine Environments, Ives, J.D. and Barry, R.G., eds., 111-132.

Marsh, P., 1978. Water Balance of a Small High Arctic Basin. Msc. Thesis, McMaster Univ., 108.

Marsh, P. and Woo, M.K., 1977. The Water Balance of a Small Pond in the High Arctic. Arctic, 30, 109-117.

Marsh, P. and Woo, M.K., 1979. Annual Water Balance of Small High Arctic Basins. Paper to be presented at Canadian Hydrology Symposium: 79 Cold Climate Hydrology, Vancouver, B.C.

Minns, C.K., 1977. Limnology of Some Lakes on Truelove Lowland. In Truelove Lowland, Devon Island, Canada: A High Arctic Ecosystem, The University of Alberta Press, 569-585.

Priestley, C.H.B. and Taylor, R.J., 1972. On the Assessment of Surface Heat Flux and Evaporation Using Large-Scale Parameters. Mon. Wea. Rev., 100, 81-92.

Schindler, D.W., Welch, H.W., Kalff, J., Brunskill, G.J., and Kritsch, N., 1974. Physical and Chemical Limnology of Char Lake, Cornwallis Island, (75° N. Lat.), J. Fisheries Research Board of Canada, 31, 585-607.

Stewart, R.B. and Rouse, W.R., 1977. Substantiation of the Priestley and Taylor Parameter α = 1.26 for Potential Evaporation in High Latitudes. J. Appl. Met., 16, 649-650.

Weller, G. and Holmgren, B., 1974. The Microclimates of the Arctic Tundra. J. Appl. Met., 13, 854-862.

Woo, M.K., 1976. Evaporation and Water Level in the Active Layer. Arctic Alpine Res., 8, 213-217.

Woo, M.K., 1976. Hydrology of a Small Canadian High Actic Basin During the Snowmelt Period. Catena, 3, 155-168.

Woo, M.K., Heron, R., Marsh, P., and Sauriol, J., 1977. Hydrology of Nival-Regime Basins in the Vicinity of Resolute, Cornwallis Is., N.W.T. A Report on Hydrologic Investigations Undertaken in May to August 1977, 41.

Woo, M.K., and Marsh, P., 1978. Analysis of Error in the Determination of Snow Storage for Small High Arctic Basins. J. Appl. Met., 17, 1537-1541.

Woo, M.K., and Steer, P., 1979. Measurement of Trace Rainfall at a High Arctic Site. Arctic, (in press).

APPENDIX ONE

LIST OF SYMBOLS

	Upper Case Roman	Units
A	area	Km^2
E	evaporation	mm
G	ground heat flux	cal cm^{-2} min^{-1}
I	inflow volume	m^3
K	shortwave radiation	cal cm^{-2} min^{-1}
L	latent heat of vaporization	cal g^{-1} $^oC^{-1}$
M	snowmelt	mm
O	outflow volume	m^3
P	precipitation	mm
Q	discharge rate	m^3 sec^{-1}
Q_O	overland flow rate	m^3 sec^{-1}
Q*	net radiation flux	cal cm^{-2} min^{-1}
R_S	subsurface flow (depth over basin)	mm
S	storage volume	
s	slope of the saturation vapour pressure vs. temperature curve	mbar $^oC^{-1}$
Sy	specific yield	
V	volume	m^3

	Lower Case Roman	Units
a	empirical coefficient	
b	empirical coefficient	
t	time	hrs.

	Greek	
α	proportionality term	
γ	psychometric constant	mbar $^{o}C^{-1}$
Δ	change in magnitude	

	Subscripts
B	basin
EQ	equilibrium
G	ground
i	ice
L	lake
R	rainfall
S	snowfall
s	sample
T	trace
W	water

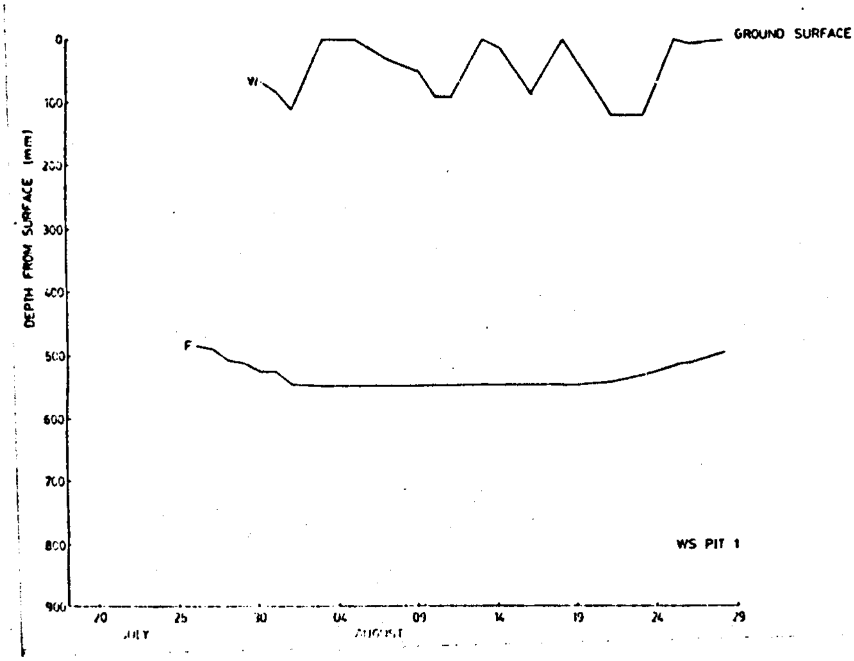
APPENDIX TWO

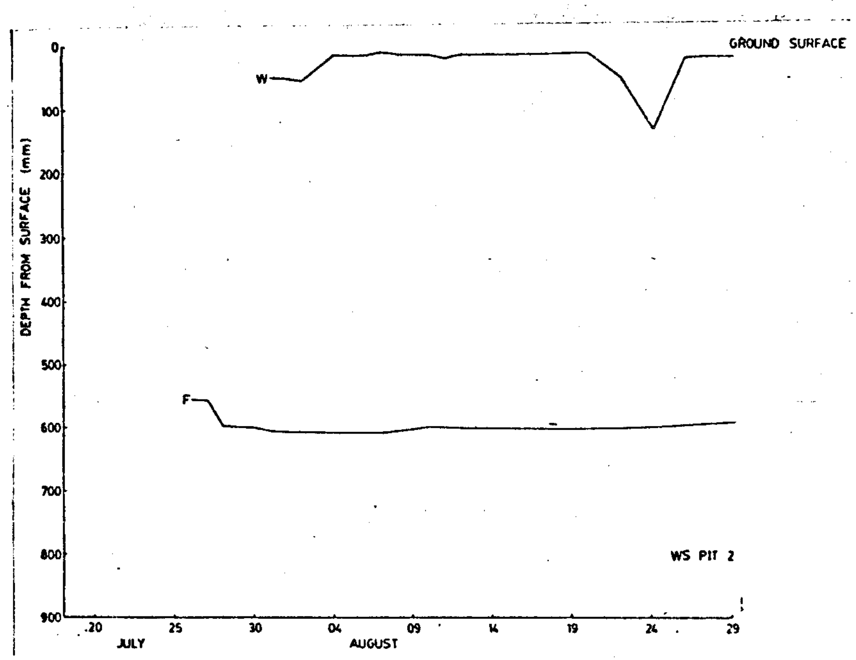
Curve Equations for Figure 2.8

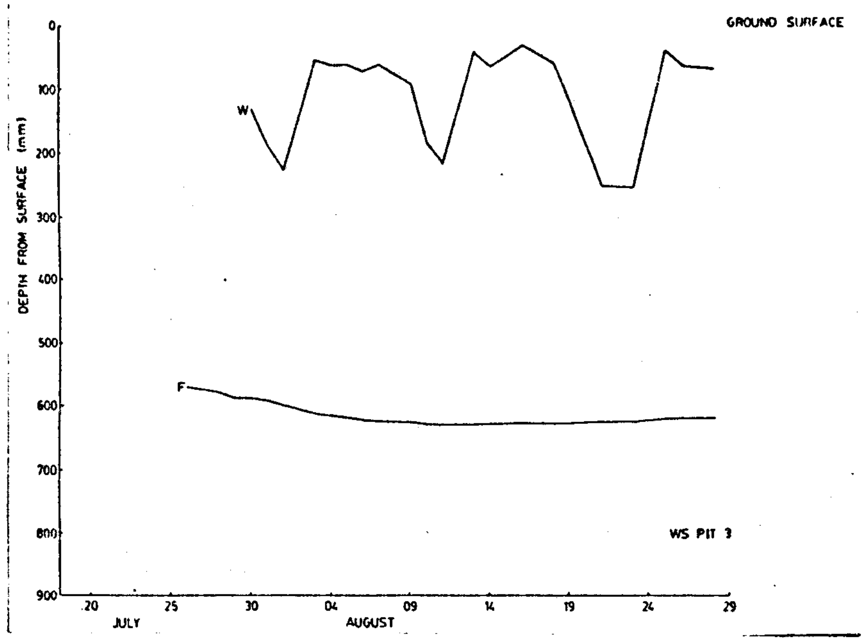
Weir	Upper Section	Lower Section
#1	$y = 9565.61 - 278.99x$	$y = 3199.83 - 89.74x$
2	$y = 5936.07 - 369.2x$	$y = 2732.27 - 153.73x$
3	$y = 3100.66 - 195.41x$	$y = 763.63 - 44.71x$
4	$y = 9109.86 - 221.42x$	$y = 3099.41 - 72.55x$
5	$y = 575.46 - 47.97x$	
Lake	$\log y = -83.337 + 111.698 \log x$	

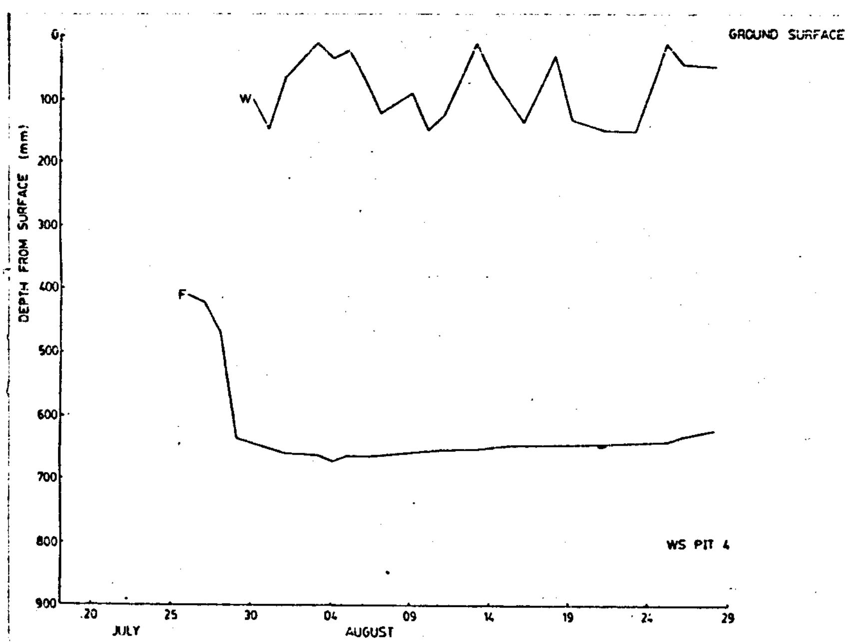
APPENDIX THREE

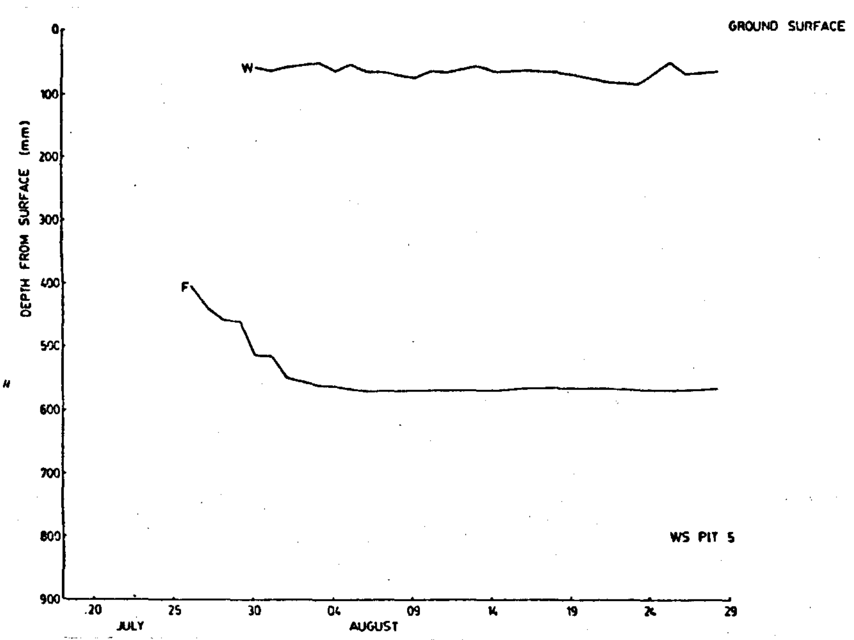
Seasonal Fluctuations in Groundwater Table (W) and Frost Table (F) for observations pits on the west-facing slope (WS) and the south-facing slope (SS)

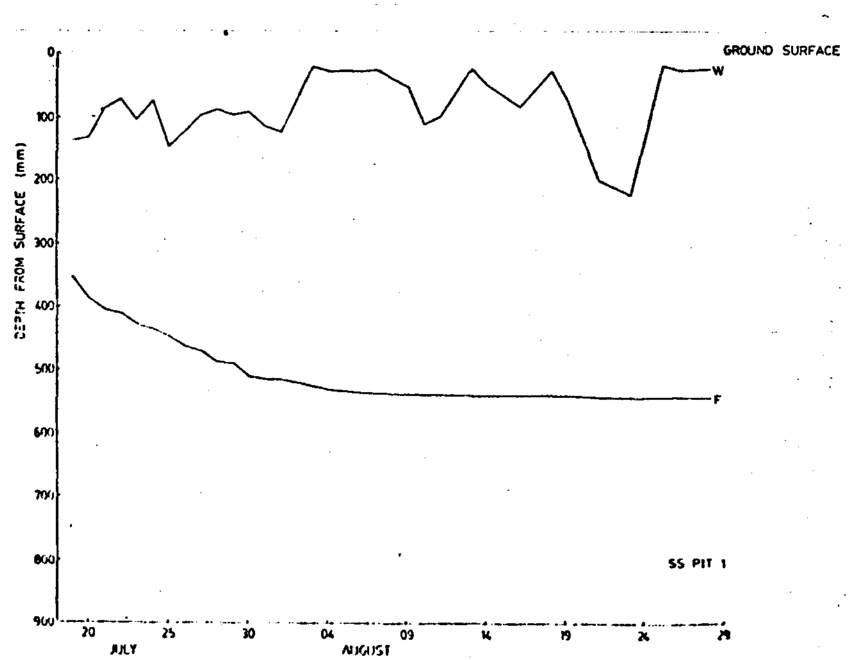

GROUND SURFACE
W
F
DEPTH FROM SURFACE (mm)
0
100
200
300
400
500
600
700
800
900
WS PIT 1
20
25
30
04
09
14
19
24
29
JULY
AUGUST

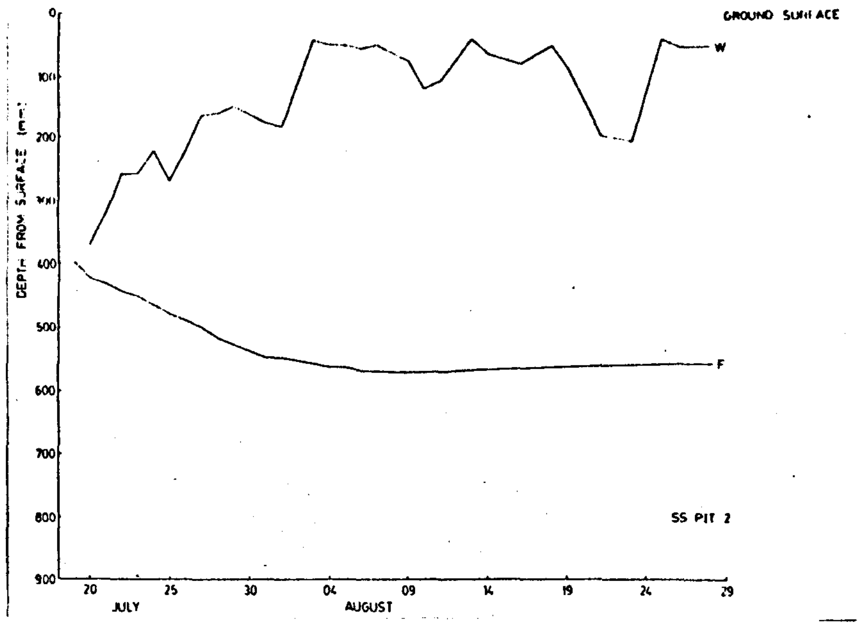

GROUND SURFACE
W
F
DEPTH FROM SURFACE (mm)
0
100
200
300
400
500
600
700
800
900
WS PIT 2
20
25
30
04
09
14
19
24
29
JULY
AUGUST

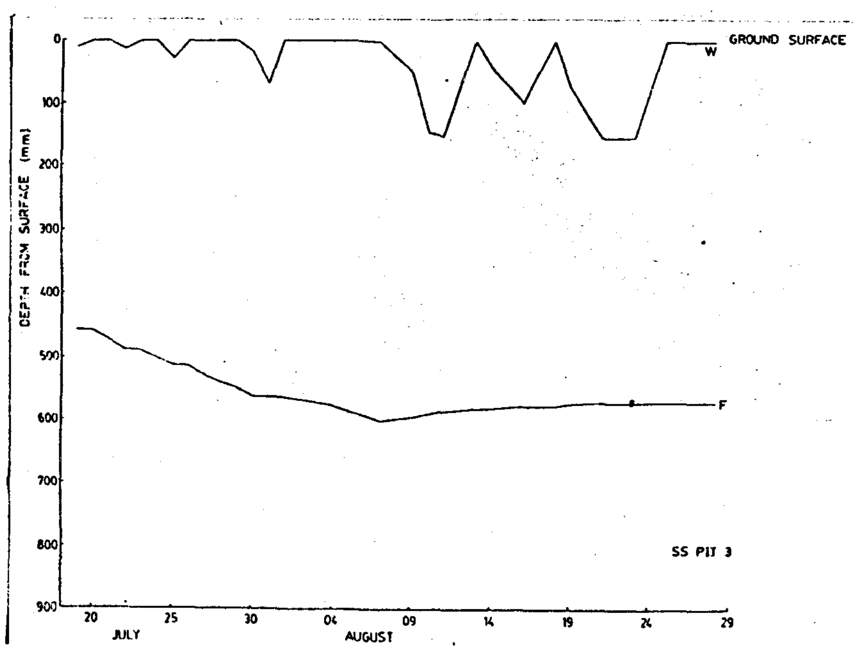

GROUND SURFACE
W
F
WS PIT 3
DEPTH FROM SURFACE (mm)
0
100
200
300
400
500
600
700
800
900
20
25
30
04
09
14
19
24
29
JULY
AUGUST

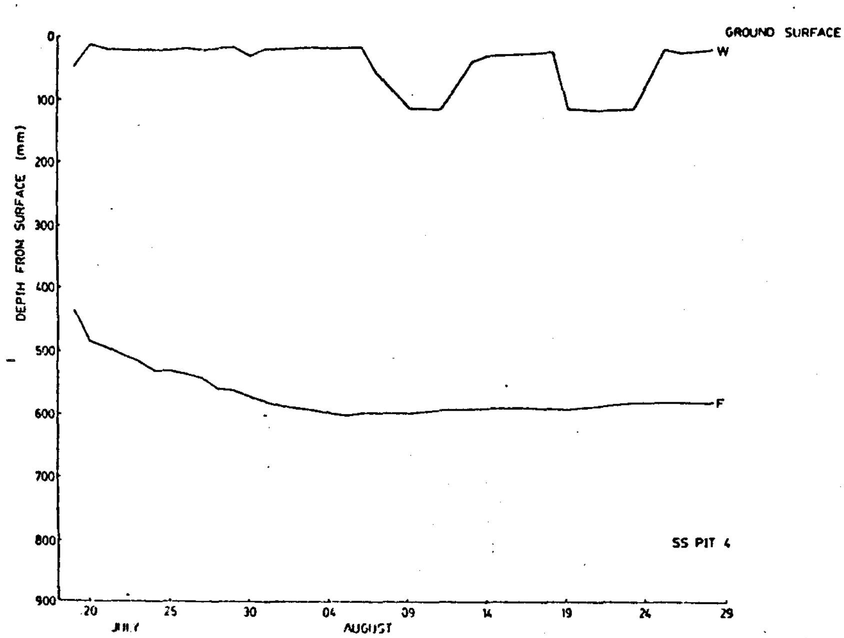

GROUND SURFACE
W
F
WS PIT 4
DEPTH FROM SURFACE (mm)
0
100
200
300
400
500
600
700
800
900
20
25
30
04
09
14
19
24
29
JULY
AUGUST

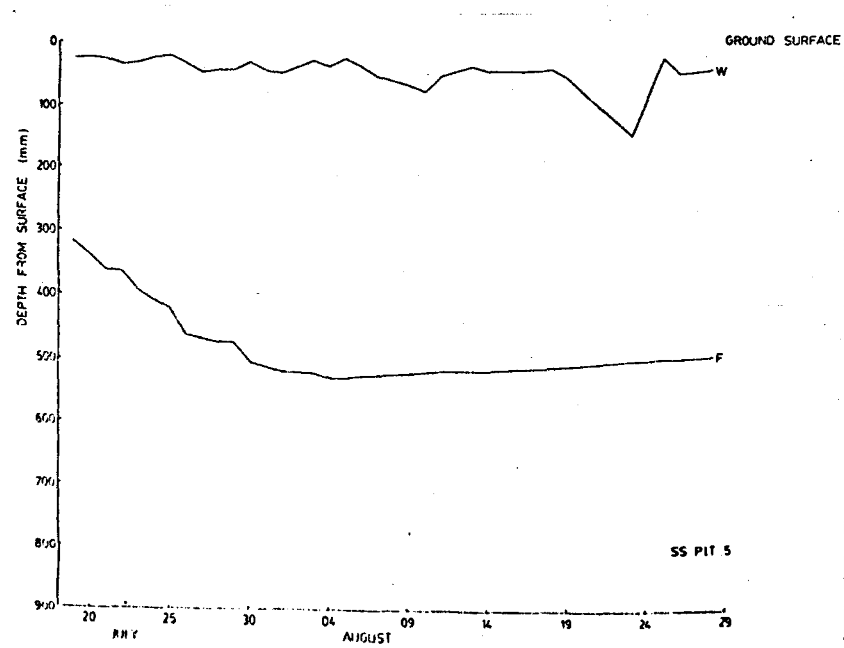

GROUND SURFACE
W
F
DEPTH FROM SURFACE (mm)
0
100
200
300
400
500
600
700
800
900
WS PIT 5
20
25
30
04
09
14
19
24
29
JULY
AUGUST

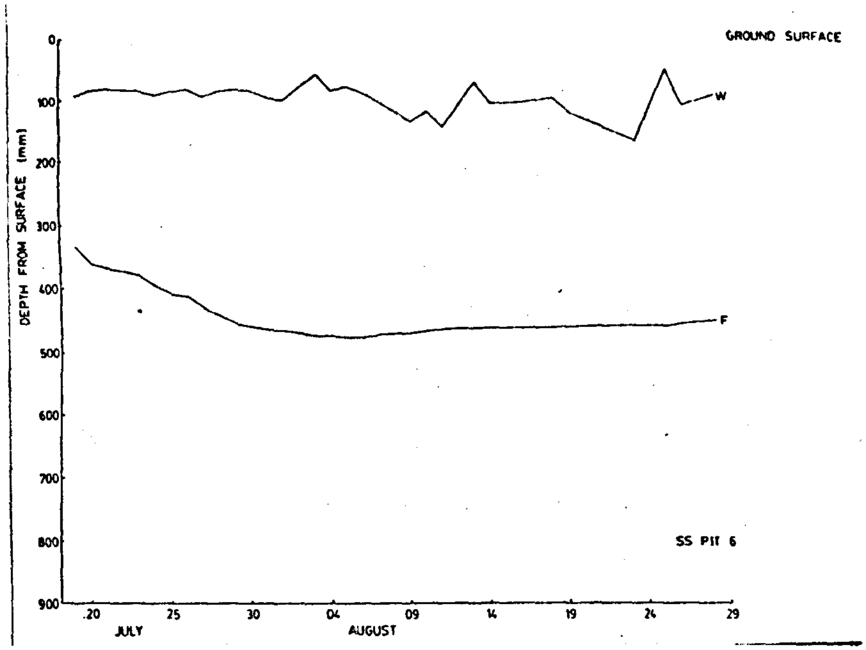

GROUND SURFACE
W
F
DEPTH FROM SURFACE (mm)
0
100
200
300
400
500
600
700
800
900
SS PIT 1
20
25
30
04
09
14
19
24
29
JULY
AUGUST

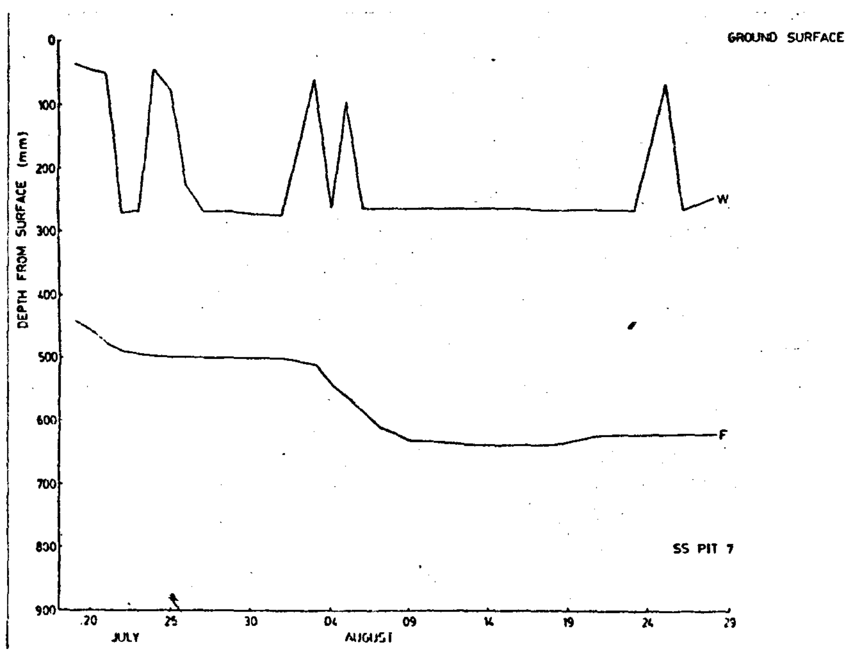

GROUND SURFACE
W
F
DEPTH FROM SURFACE (mm)
0
100
200
300
400
500
600
700
800
900
SS PIT 2
20
25
30
04
09
14
19
24
29
JULY
AUGUST

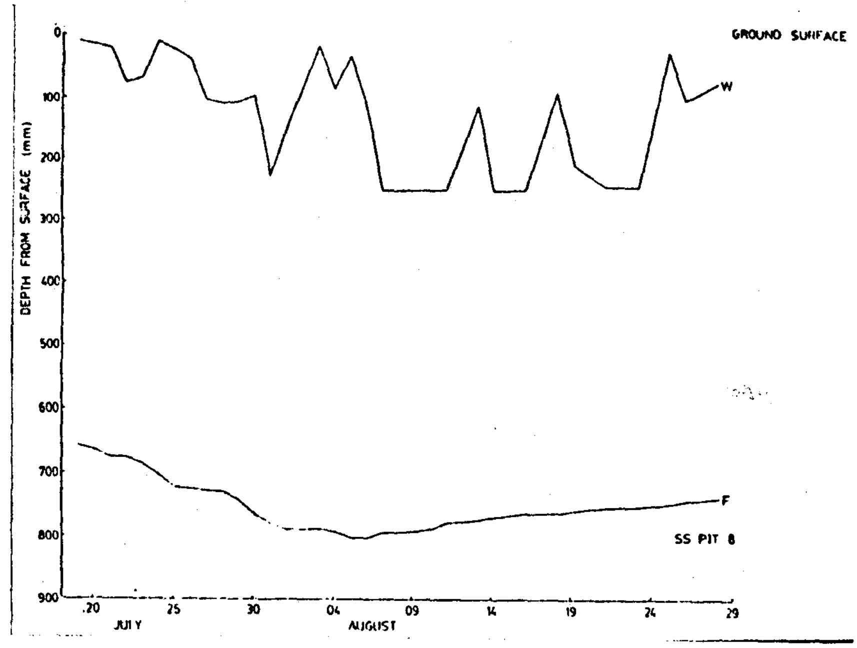

GROUND SURFACE
W
F
DEPTH FROM SURFACE (mm)
0
100
200
300
400
500
600
700
800
900
SS PIT 3
20
25
30
04
09
14
19
24
29
JULY
AUGUST

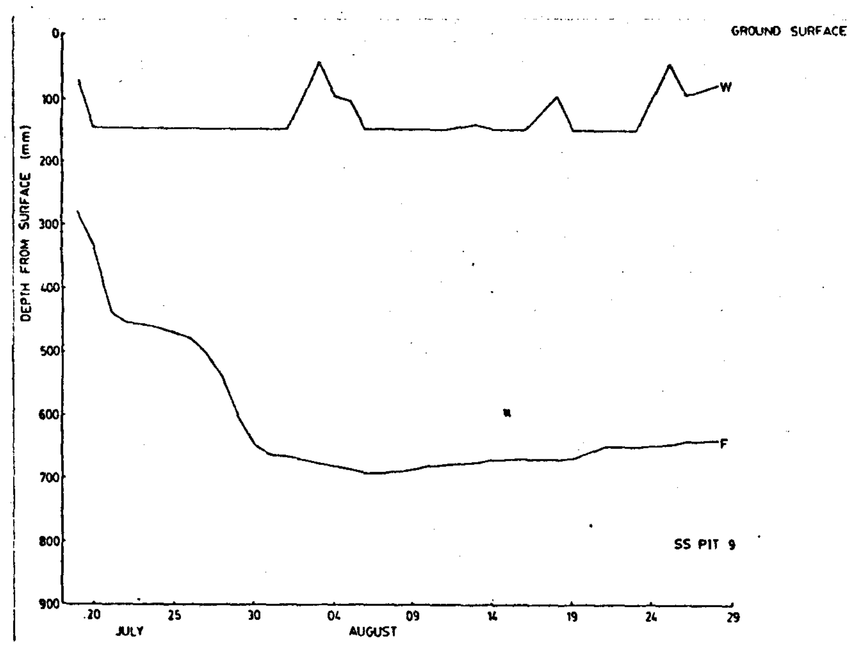

GROUND SURFACE
W
F
DEPTH FROM SURFACE (mm)
0
100
200
300
400
500
600
700
800
900
SS PIT 4
20
25
30
04
09
14
19
24
29
JULY
AUGUST

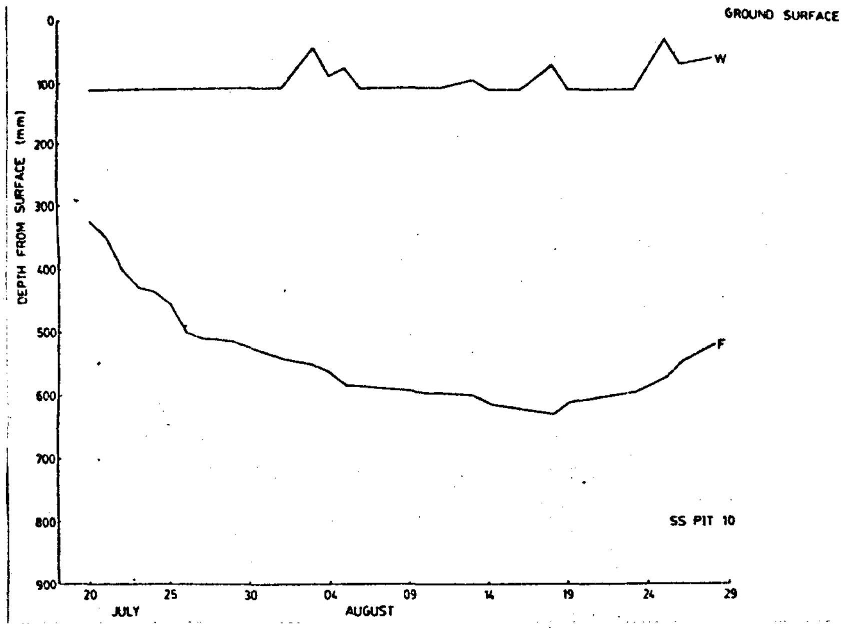

GROUND SURFACE
W
F
DEPTH FROM SURFACE (mm)
0
100
200
300
400
500
600
700
800
900
SS PIT 5
20
25
30
04
09
14
19
24
29
JULY
AUGUST

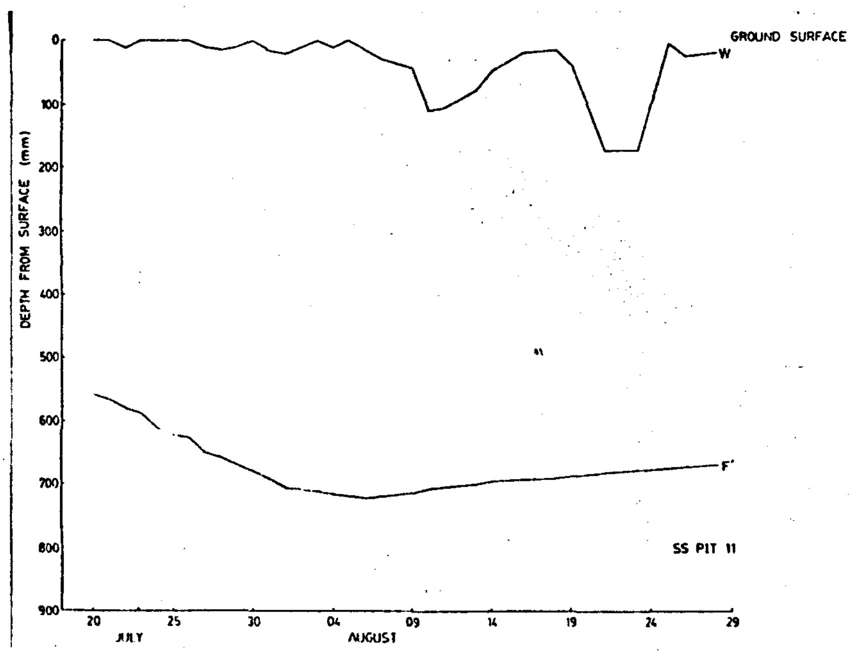

GROUND SURFACE
DEPTH FROM SURFACE (mm)
0
100
200
300
400
500
600
700
800
900
W
F
SS PIT 6
.20
25
30
04
09
14
19
24
29
JULY
AUGUST

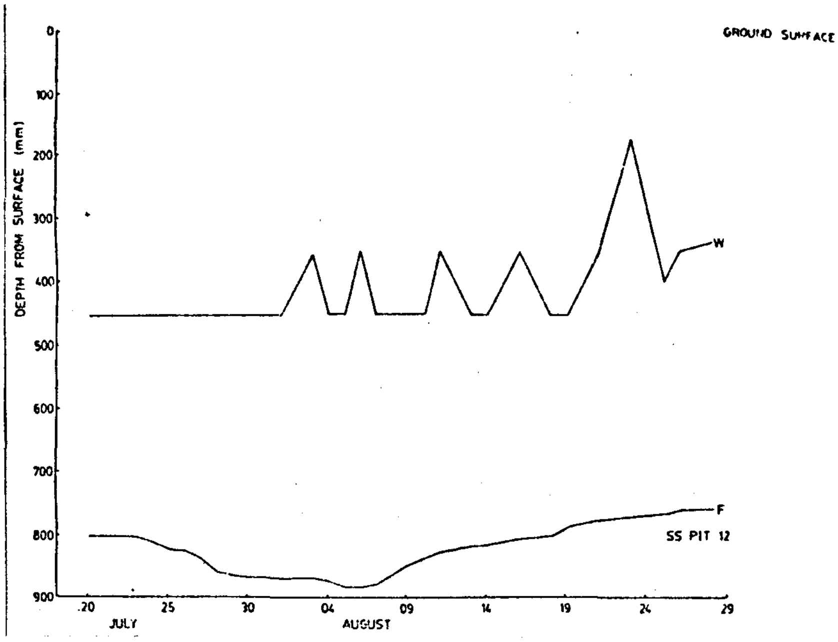

GROUND SURFACE
DEPTH FROM SURFACE (mm)
0
100
200
300
400
500
600
700
800
900
W
F
SS PIT 7
.20
25
30
04
09
14
19
24
29
JULY
AUGUST

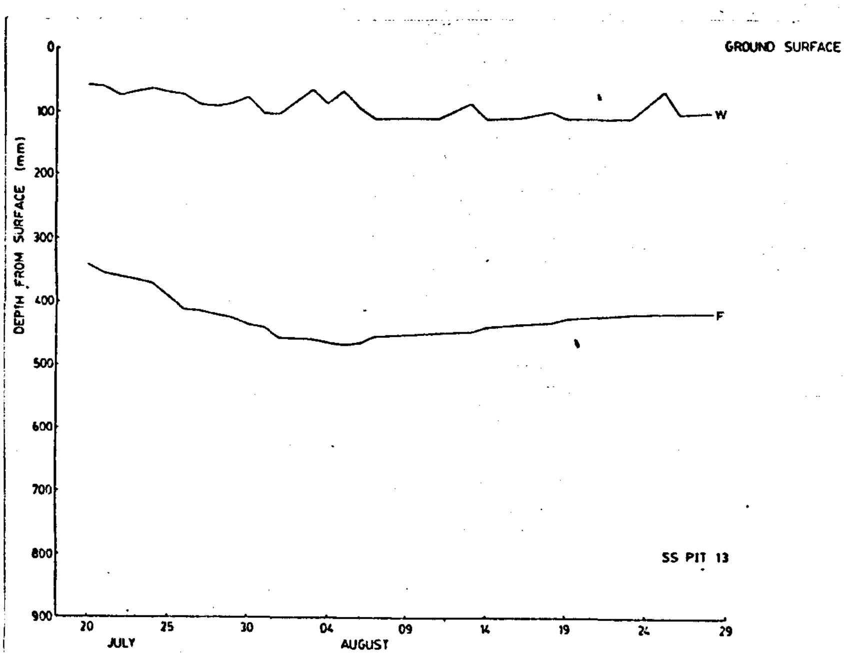

GROUND SURFACE
W
F
SS PIT 8
DEPTH FROM SURFACE (mm)
0
100
200
300
400
500
600
700
800
900
20
25
30
04
09
14
19
24
29
JULY
AUGUST

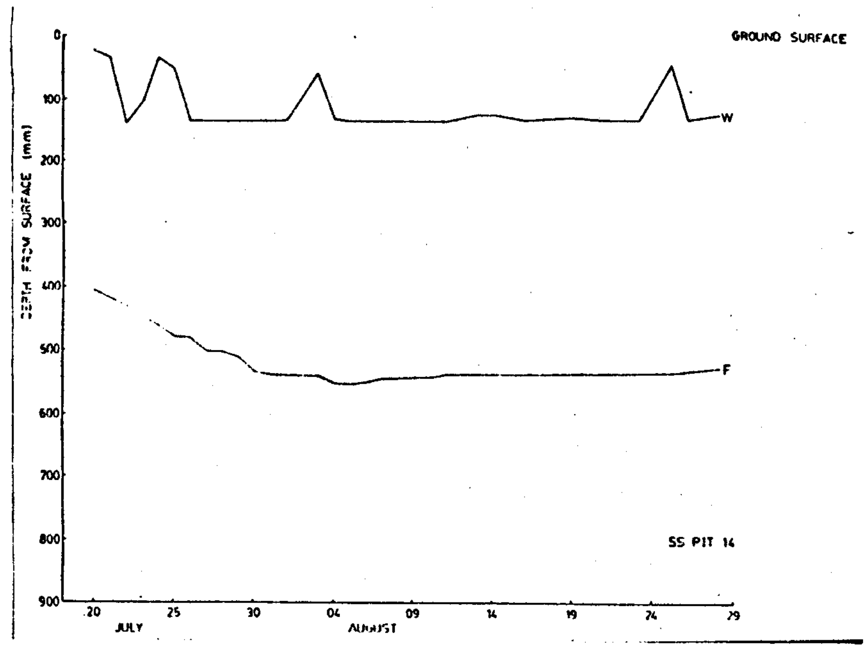

GROUND SURFACE
W
F
SS PIT 9
DEPTH FROM SURFACE (mm)
0
100
200
300
400
500
600
700
800
900
20
25
30
04
09
14
19
24
29
JULY
AUGUST

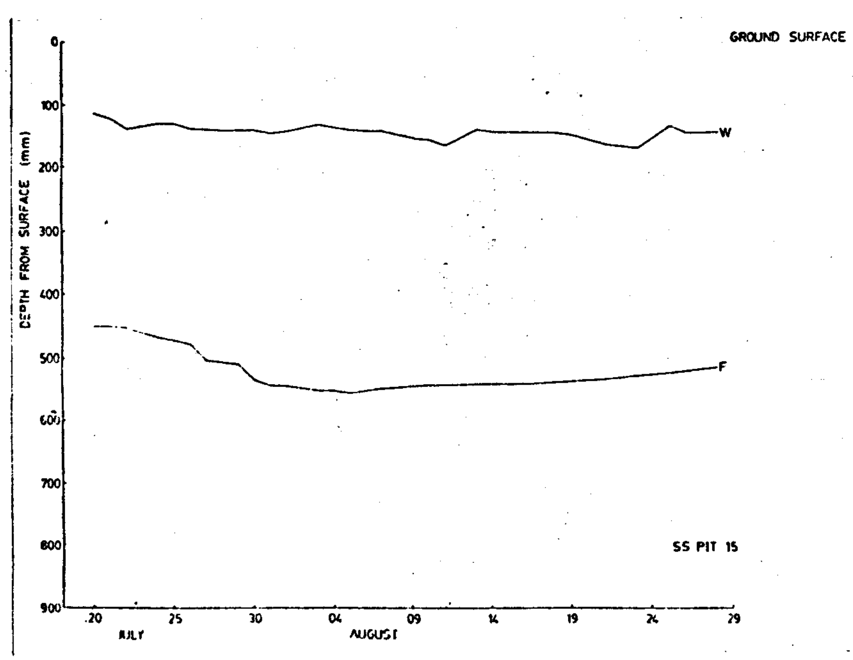

GROUND SURFACE
W
F
SS PIT 10
DEPTH FROM SURFACE (mm)
0
100
200
300
400
500
600
700
800
900
20
25
30
04
09
14
19
24
29
JULY
AUGUST


GROUND SURFACE
W
F
SS PIT 11
DEPTH FROM SURFACE (mm)
0
100
200
300
400
500
600
700
800
900
20
25
30
04
09
14
19
24
29
JULY
AUGUST


GROUND SURFACE
DEPTH FROM SURFACE (mm)
0
100
200
300
400
500
600
700
800
900
W
F
SS PIT 12
20
25
30
04
09
14
19
24
29
JULY
AUGUST


GROUND SURFACE
DEPTH FROM SURFACE (mm)
0
100
200
300
400
500
600
700
800
900
W
F
SS PIT 13
20
25
30
04
09
14
19
24
29
JULY
AUGUST


GROUND SURFACE
W
F
SS PIT 14
DEPTH FROM SURFACE (mm)
0
100
200
300
400
500
600
700
800
900
20
25
30
04
09
14
19
24
29
JULY
AUGUST


GROUND SURFACE
W
F
SS PIT 15
DEPTH FROM SURFACE (mm)
0
100
200
300
400
500
600
700
800
900
20
25
30
04
09
14
19
24
29
JULY
AUGUST