HYDROLOGISEN TOIMISTON TIEDONANTOJA
MEDDELANDEN FRÅN HYDROLOGISKA BYRÅN XXV

MEASUREMENT OF EVAPOTRANSPIRATION AND COMPUTATION OF WATER BUDGET IN TREELESS PEATLANDS IN THE NATURAL STATE

JUHANI VIRTA

HELSINKI 1966

List of symbols

A	Parameter in equation (18)
a	Constant term in regression equation
a_1, a_2, a_3	Constants of function $R(W)$
b	Regression coefficient
c	Cloudiness
c_i	Term due to random factors in equation (1)
D	Difference of water level elevation. Subscripts of this symbol refer to the ordinal of water gauges
d	Relative error in the measured evapotranspiration. Differential operator
E	Evapotranspiration
E_a	Auxiliary quantity in equation (2)
E_{pan}	Evaporation from Class A pan
E_z	Evapotranspiration from a point with the natural water depth z (cm)
E^M	Evaporation computed by the Makkink-formula. Subscript (o) indicates free water surface and subscript (p) potential evapotranspiration
E^P	Evaporation computed by the Penman-formula. Subscripts as in symbol E^M
e_a	Water vapour pressure
e_s	Saturated water vapour pressure
e_1, e_2	Constants of function $\bar{f}_{z_0}(W)$
e_1', e_2'	Constants of function $\bar{f}_{z_0}(z)$
f	Function for determining the dependence of evapotranspiration on the water depth
$f_{z_0}(z)$	Ratio of the evapotranspiration from a point with the water depth at z to the evapotranspiration from a point with the water depth at z_0
$\bar{f}_{z_0}$	Areal average of function f_{z_0}. The argument of this function may be either the water depth or the water stage
g	Distribution of the height of bog-surface in relation to the water level
H	Radiation balance
h	Height of bog-surface in relation to the 0-level of the water gauge
I	Water surface inclination
i	Index of summation. General index
j	General index
K	Integrated hydraulic conductivity
k	Hydraulic conductivity
l	Co-ordinate in the horizontal direction normal to the direction of water flow
m_x	Evapotranspiration ratio
n	Number of observations. A parameter in equation (18)
n/N	Relative time of sunshine
P	Precipitation
p	Parameter depending on the error of areal evapotranspiration
Q	Amount of flow in unit time
q	Parameter depending on the error of areal evapotranspiration

R	Runoff. Multiple correlation coefficient in Table 8
R_m	Measured total radiation
R_{l_e}	Effective long-wave radiation
r	Correlation coefficient
S	Total water content
s	Storage coefficient
s_e	Standard deviation of residuals. Standard error on page 51
s_y	Standard deviation of the independent variable
s_1, s_2	Constants of function $s(W)$
T	Absolute temperature
t	Time. Temperature in Table 1
u	Auxiliary variable
v	Wind velocity. Auxiliary variable
v_i	Weighting factor
W	Water stage. Subscript (b) indicates water stage at the beginning of a period, (e) indicates water stage at the end of a period and subscript (0) indicates gauge height corresponding to the bog-surface
X	General meteorological quantity
x	Horizontal axis
y	Vertical axis
z	Water depth
γ	Psychrometer constant
Δ	Slope of the saturated-vapour-pressure curve. Residual. The symbol indicates also the change of a quantity
Δe	Moisture deficit
σ	Boltzmann constant
—	Bar over a symbol denotes time average. Bar over the symbol f denotes areal average

COMMENTATIONES
PHYSICO-MATHEMATICAE
Societas Scientiarum Fennica

Vol 32 Nr 11. 1966

Measurement of Evapotranspiration and Computation of Water Budget in Treeless Peatlands in the Natural State

JUHANI VIRTA

HELSINKI — HELSINGFORS
1966

PREFACE

The main purpose of the measurements in this investigation was to provide information about evapotranspiration from a certain fen situated in Lapland at Korvanen in an area proposed for becoming an artificial reservoir. The measurements there were carried out under the auspices of the League for the Conservation of Nature in Finland during June, July and August in 1959 and 1960. For the sake of obtaining comparison similar measuring stations were established in East Bothnia at Möksy in 1960 and 1961 and in Southern Finland at Loppi in 1962.

The measurements were carried out under the leadership of professor HEIKKI SIMOJOKI, to whom I am very grateful for suggestions concerning the measurements and the treatment of the observation data. He also made available the Finnish Hydrological Office force to assist in the setting up of the measuring stations. To professor RAUNO RUUHIJÄRVI I am deeply indepted for recommendations concerning botanical questions. Special thanks go to Mr. G. W. BLOEMEN for helping me gain insight into the water budget calculation method used in the Netherlands.

The observations for this investigation were carried out by Messrs. MATTI HAAPASAARI, KARI KOSKINEN, MATTI LÄHDEOJA, JUHANI MÄKIPÄÄ and MAUNO OUTINEN. The calibration of some of the meteorological instruments was made by Mr. URHO NEVALAINEN from the Finnish Meteorological Office, and Mr. DIMITRI RÉMY provided a translation of the Russian literature referred to in this work. To all of these I express my deepest gratitude. I would also like to thank Mrs. LORNA J. SUNDSTRÖM for revising the English in my manuscript.

I also gratefully acknowledge receipt of financial assistance from the following sources: the National Research Council for Sciences for support during the measurements and preliminary treatment of the results; and the Delegation of Sohlberg of the Societas Scientiarum Fennica for support while furthering the treatment of the observation material.

Leppävaara, May, 1966.

JUHANI VIRTA

CONTENTS

1. INTRODUCTION

Determination of evapotranspiration and evaporation

Several methods are available for determining evapotranspiration and evaporation. The measurement of the water budget of a lysimeter, a tank filled with water or soil, is the oldest and perhaps also today the most widely used method. In Finland this method has been applied by WÄRE (1947), RENQVIST (1951), PORKKA (1956), JUUSELA (1962) and PÄIVÄNEN (1964) for the determination of evapotranspiration and by BLOMQVIST (1917) and FRANSSILA (1940) for the determination of evaporation from lakes. The same method has been applied by KAITERA (1939) for the determination of evaporation from snow cover.

Evapotranspiration and evaporation can be computed, if other factors in the energy budget have been measured or estimated. This has been done in Finland by FRANSSILA (1940) for the determination of evaporation from a lake.

Evapotranspiration and evaporation can be determined also on the basis of the theory of the mass transfer of water vapour. A simplification of this approach can be used. Evaporation from a water surface is approximately proportional to the product of wind velocity and the difference between water vapour pressure at the water surface and in the air above the surface. This method has been applied by SIMOJOKI (1948) for the determination of evaporation from the Northern Baltic Sea.

Recently the method by which evapotranspiration can be calculated from the water budget of the atmosphere has been improved. This method has been applied by VÄISÄNEN (1962) for the whole of Finland, and by PALMÉN (1963) for the Northern Baltic Sea.

Transpiration from trees can be determined by measuring the change in weight of a twig or a whole tree during a short time interval immediately after cutting; this method has been used in Finland by G. SIRÉN (1955).

Evapotranspiration can also be determined with the aid of the water budget of a catchment area, which is the simplest method of obtaining mean annual evapotranspiration; in Finland this method has been used by A. SIRÉN (1948) and NIINIVAARA (1953). In some cases this water budget method may be applied for shorter periods of, for instance, one or two months (A. SIRÉN, 1936; NIINIVAARA, 1953, 1955; HUIKARI, 1959b) or for even shorter periods (HEIKURAINEN, 1963).

Evapotranspiration has also been computed from the tritium and chlorine contents in water.

Evapotranspiration determinations in peatlands

For the determination of evapotranspiration from peatlands use has been made of lysimeters, energy budget, water budget, and chlorine budget methods. A brief outline of measurements will now be given.

In Finland HOMÉN (1893, p. 127) carried out measurements in a fen. Several WILD's evaporimeters were employed, one of which contained only water the rest being filled with different kinds of peat.

BLOMQVIST (1917, p. 63) measured evapotranspiration from lysimeters filled with water and peat-moss. In measurements which were carried out during a period of eight days in August 1913, the total amount of evapotranspiration from the fen was 53 mm. During the same period the evaporation from a tank in an adjacent lake amounted to 40 mm.

Besides other studies in Maasoja in the experimental field of the Board of Agriculture evapotranspiration from lysimeters filled with peat has been studied (WÄRE, 1947). In particular the influence of the ground-water depth on evapotranspiration was studied.

PESSI (1957), who determined evapotranspiration from cultivated peatland by weighing small peat samples, observed that evapotranspiration was greater from peat than from mineral soil.

HUIKARI (1959b) determined evapotranspiration from drained peatlands. Precipitation and runoff were measured and evapotranspiration was calculated for periods with approximately equal water content at the beginning and end of the period. The depth of the groundwater level was considered as an indicator of the water content.

In Denmark (PRYTZ, 1932) evapotranspiration in a raised bog has been measured with lysimeters, and the true evapotranspiration was calculated from the fluctuations of the bog water level.

In Poland (BAC, 1936) the dependence of evapotranspiration on the vegetation growth in peatland has been investigated, in addition to a study of the influence of water depth on evapotranspiration.

In Latvia (NOMALS, 1938) evapotranspiration in a raised bog has been measured with RYKATSCHEV lysimeters. It was noticed that evapotranspiration from the peat was slightly greater than evaporation from free water surfaces.

In the Soviet Union the evapotranspiration from peatlands has been investigated for several years now. For instance, in investigations in Novgorod, it was noticed (DUBACH, 1936) that evapotranspiration is dependent on plant cover and on the depth of the ground-water level. The lysimeter used was of the RYKATSCHEV type. Other measurements carried out in the Soviet Union have been presented by IVANOV (1957) and ROMANOV (1957, 1960).

In Germany (BADEN and EGGELSMANN, 1958; EGGELSMANN, 1963) evapotranspiration measurements have been performed in two raised bogs, the one cultivated and the other in the natural state. Evapotranspiration was determined directly by lysimeters and also by the water budget method.

In Sweden (ODÉN, 1964) evapotranspiration from a raised bog has been computed by the chemical method.

2. MEASUREMENTS

Evapotranspiration measuring stations

The measurements for the present investigation were carried out in three bogs. Fig. 1 shows the situation of the stations.

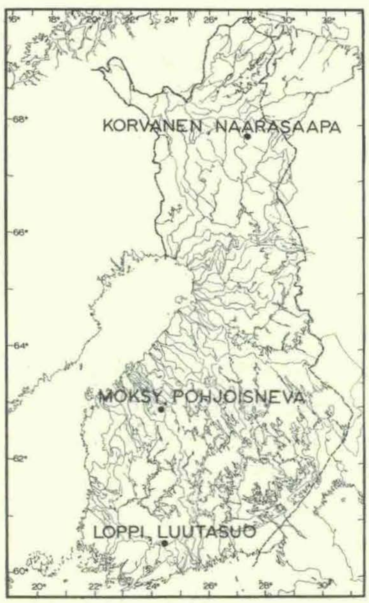

Fig. 1. Evapotranspiration measuring stations.

The evapotranspiration measuring station in Lapland was situated at Korvanen in a string fen called Naarasaapa ($\varphi = 67°57'$N, $\lambda = 27°48'$E). Measurements were carried out during June, July and August in 1959 and 1960. This fen is a large, treeless fen typical of Lapland. The layer of peat at the measuring station is 2 m thick. In Fig. 2a the measuring station with its neighbourhood is shown.

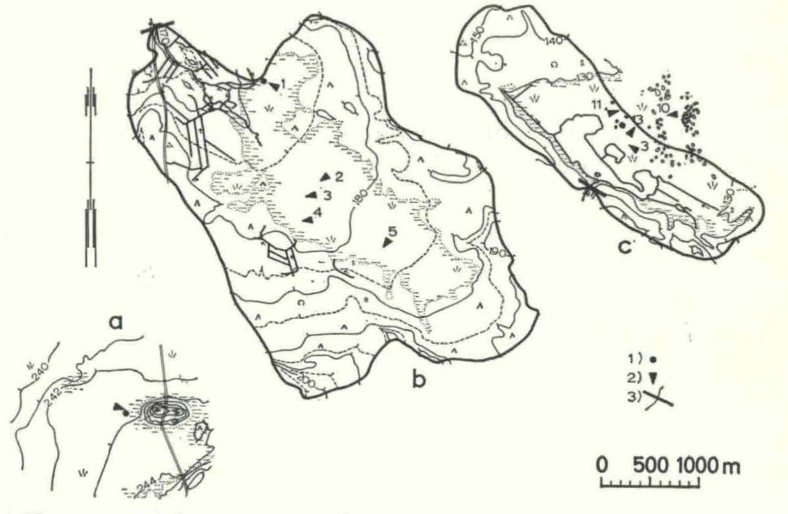

Fig. 2. a) Korvanen, b) Möksy, c) Loppi.
1) Evapotranspiration measuring station, 2) water gauge, 3) runoff measuring site.

The evapotranspiration measuring station in East Bothnia was situated at Möksy in an open fen called Pohjoisneva ($\varphi = 63°5'$N, $\lambda = 24°17'$E). Measurements were carried out during June, July and August in 1960 and 1961. This fen can be divided into two different parts, hollow (»rimpi») fen in the middle and *sphagnum papillosum* fen surrounding this. The evapotranspiration measurements represent the latter type. The layer of peat at this measuring station is 1.5 m thick, and there is a streamlet, Poikkijoki river (Fig. 2b) flowing from the fen.

The evapotranspiration measuring station in Southern Finland was situated at Loppi in a raised bog called Luutasuo ($\varphi = 60°41'$N, $\lambda = 24°19'$E). Measurements were carried out during June, July and August 1962. The layer of peat at the measuring station is 7 m thick, and again a streamlet flows from the bog (Fig. 2c).

Evapotranspiration

It is often considered desirable to arrange the measurements in such a way that the determination of the water budget of a soil sample is based on the weight of the sample. For the determination of evapotranspiration a method based on keeping the water level at a constant height in the soil tank has also been used. For the automatic water level regulator a float system (MATHER, 1950) or a regulator based on »MARIOTTE's bottle» (JUUSELA, 1962) has been used. With a method based on weighing the sample reliable results may be obtained also for short periods. For weighing large soil samples, however, a more complicated set of measuring instruments is required.

For the purpose of using a large sample a measuring system based on constant water level was used, as it was difficult to construct a balance for weighing such a sample. The lysimeter set consisted of a sample container (A in Fig. 3), an overflow tank (C) and a threshold (B). The sample container, which was made of brass plate 2 mm thick, had a diameter of 120 cm and a depth of 50 cm. This container was filled with water and pieces of peat and was submerged into the bog on wooden poles. A uniform plate of peat, 20 to 30 cm thick, cut out from the bog, formed the surface of the sample. From the threshold the sample container was connected with the overflow tank by a plastic tube. In Figures 4 to 7 the sample containers and other parts of the lysimeter during different observation periods are illustrated.

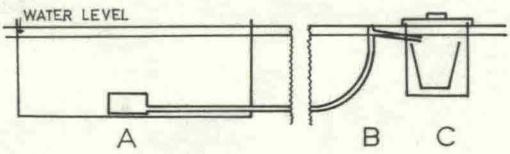

Fig. 3. The lysimeter used in evapotranspiration measurements.
A — the sample container, B — threshold and C — the overflow tank.

For obtaining the evapotranspiration for 24 h periods measurements were carried out in the following manner. In the evening a known amount of water was poured into the sample container and in the morning the amount of water in the overflow tank was measured. The difference between these two amounts, added to the eventual precipitation, was the evapotranspiration of the preceding day and night.

At Korvanen and Möksy in 1959 and 1960 the evapotranspiration was measured from samples with wet surface. The water level in the containers was maintained at a depth of 2 cm. As the evapotranspiration is dependent on the depth of the water level, and since the water level in the bog in the natural state varies and is usually at depths greater than 2 cm, the measured values are obviously too high an estimation of the actual evapotranspiration from the bog. On account of this, measurements of evapotranspiration were repeated later using sample containers with the water level deeper than 2 cm. Thus at Möksy, summer 1961, the water level in the containers was maintained at depths of 2 cm and 15 cm and at Loppi (summer 1962) at depths of 2 cm, 4 cm, 11 cm and 16 cm.

For comparison evaporation was measured from a tank filled with water. This tank, with a diameter of 120 cm and height 25 cm, was set up on the bog surface.

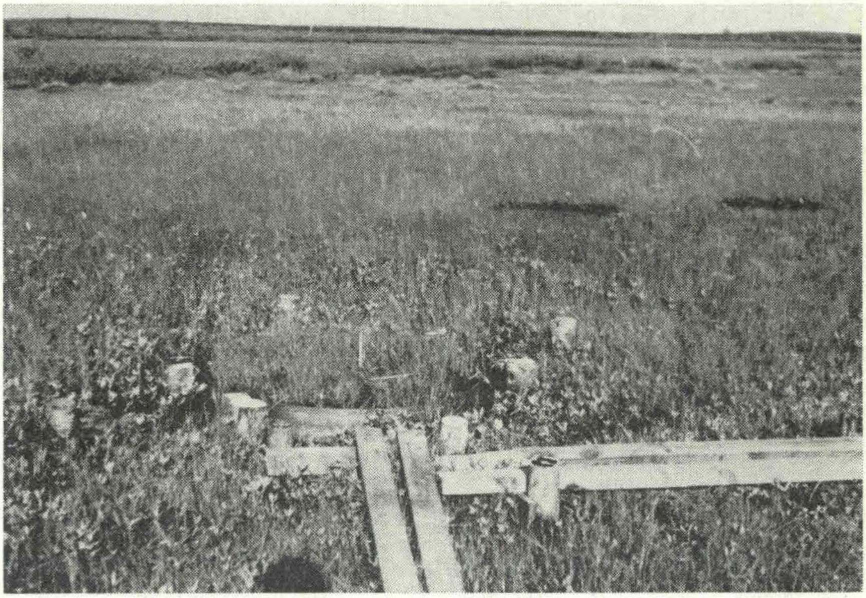

Fig. 4. Sample container at Korvanen in 1960. Water level at a depth of 2 cm.

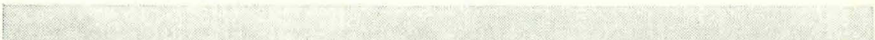

Fig. 5. Sample containers at Möksy in 1961. Water level at a depth of 15 cm.

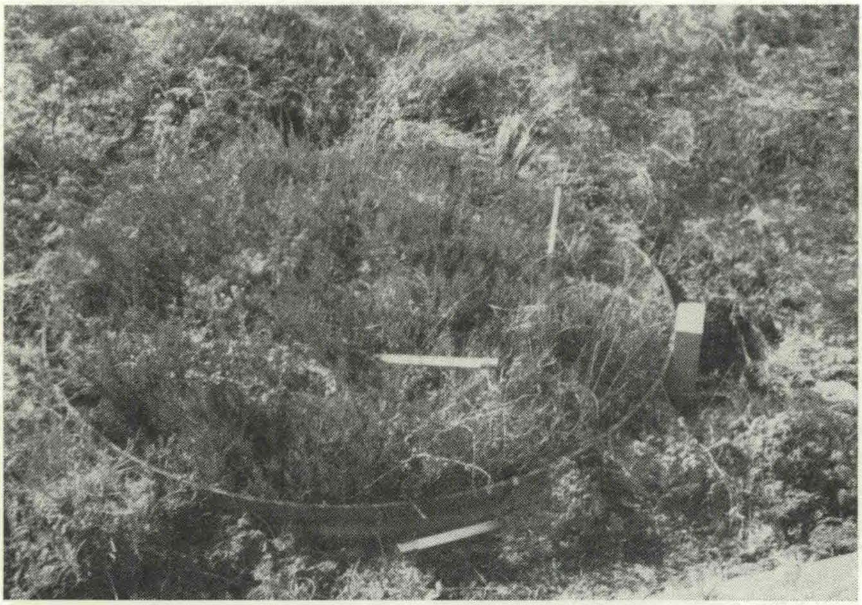

Fig. 6. Sample container at Loppi in 1962. Water level at a depth of 16 cm.

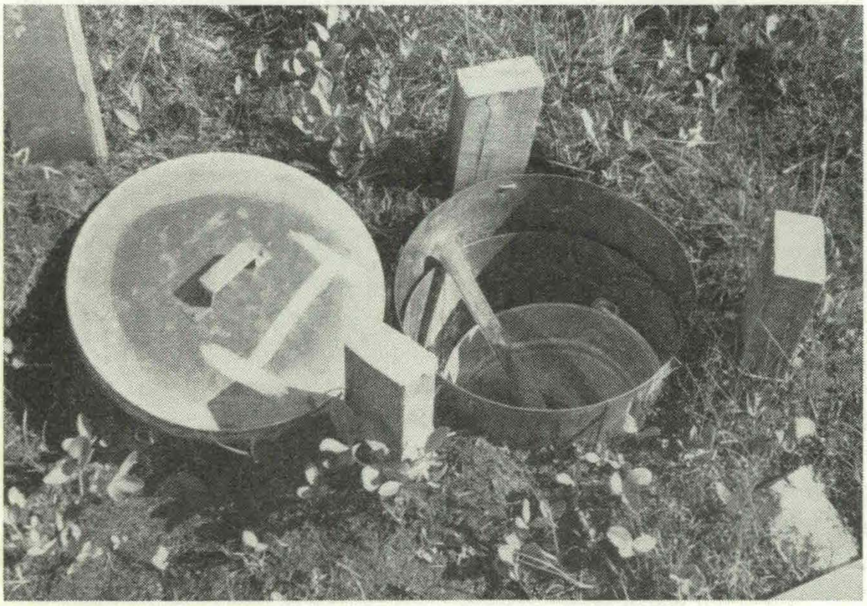

Fig. 7. Overflow tank at Korvanen in 1959.

Precipitation

Precipitation measurements were made using a rain gauge with a collecting area of 500 cm². This was installed on the bog, as shown in Fig. 8, with the rim at a height of 25 cm above the bog surface. Measurements were carried out in the mornings and evenings. For the determination of the time distribution of precipitation, a rain recorder of the Fuess model with a collecting area of 200 cm² and with the rim at a height of 1 m, was used.

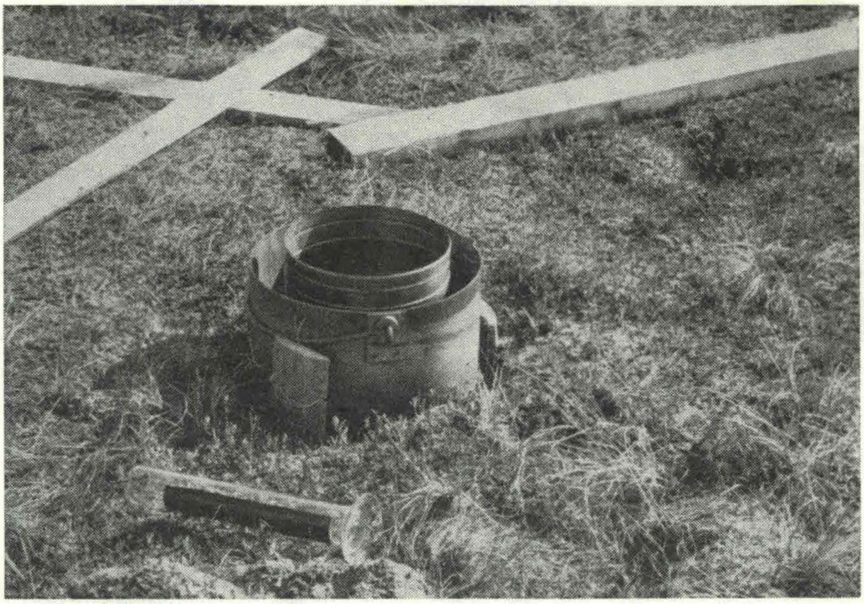

Fig. 8. Loppi 1962. Rain gauge.

Water stage

At each measuring station there was always a water gauge situated close to the sample containers. At Korvanen a graduated measuring rod fixed into a wooden pole served as the gauge. At Möksy and at Loppi the water stage was determined with the aid of a fixed mark in a wooden pole and an independent graduated rod. Measurements were carried out in the evenings and precautions were taken to obtain the readings with an accuracy of 1 mm.

It is difficult to determine the actual water stage of the bog as the water level rises whenever an observer is walking close to the gauge. At Korvanen and at Möksy it was endeavoured to avoid this error factor by measuring the water stage from a bridge supported on wooden poles. However, at Loppi, where the observer had to walk in the bog close to the gauge, steps were taken to ensure that the speed of taking the readings was the same each time, so that the errors included in every measurement would be of the same order. In addition, the gauges were installed in the watery parts between hummocks.

The values of the water stage, presented later, were compared with an arbitrarily chosen 0-level. For practical reasons this was chosen so that the lowest measured water level corresponded approximately to a water stage of 0 cm.

Meteorological observations

For the approximate computation of evapotranspiration several meteorological observations were made.

Total incoming radiation was in general measured with a ROBITZSCH bimetal actinograph. This instrument was installed close to the lysimeters 1.5 m above bog surface. To obtain the values of incoming radiation the calibration curve given by the manufacturer was used.

At Möksy, summer 1960, the incoming radiation was measured with a MOLL—GORCZYNSKY solarimeter. As it was not possible to calibrate the instrument on the spot, the coefficient of calibration was determined by comparing the results of measurements with the hours of sunshine. The dependence of the total incoming radiation on sunshine hours was determined with the aid of measurements by the ROBITZSCH actinograph during summer 1961. The calculation of the mean calibration coefficient for June, July and August was based on the assumption that the dependence of total radiation on sunshine hours is linear during the same period each year (ÅNGSTRÖM, 1928). According to the investigation by MATTSSON and RAPELI (1960) the coefficient of this type of solarimeter may vary by about 5% during a summer. Considering the accuracy of the calculations in this investigation the use of the same coefficient throughout the whole summer may be regarded as sufficient.

Air temperature was recorded by using a bimetal thermograph installed on the bog in a meteorological hut at a height of 2 m. Air humidity was recorded by a hair hygrometer in the same hut. Measurements were checked every morning and evening either by an ASSMANN psychrometer or by a wet and dry bulb thermometer in the hut. In the case of Korvanen these measurements were taken from micrometeorological measurements carried out by FRANSSILA (1960, 1962).

Wind velocity was measured at 2 m height by a Fuess handanemometer, which was furnished with a rotating disk, the position of which was observed every morning and evening. For the computation of wind velocity from the observed data the calibration curve given by the manufacturer was used. In addition the anemometers were calibrated in the Finnish Meteorological Office.

3. EVAPOTRANSPIRATION

3.1 Results of evapotranspiration

Air temperature, cloudiness and precipitation during the months of measurements

Table 1 presents the deviation of mean monthly temperature and cloudiness from the corresponding means of the period 1931—1960 taken at five meteorological stations of the Finnish Meteorological Office. The results of the stations at Ivalo and Sodankylä correspond to the conditions at Korvanen, results from Vaasa and Jyväskylä to the conditions at Möksy, and values measured at Jokioinen to the conditions at Loppi, respectively.

It is observed that the measuring periods at Korvanen were warmer than the corresponding mean, and that cloudiness was close to the average value. At Möksy the air temperature was close to and cloudiness above the average during the measuring periods. At Loppi the air temperature was lower and the cloudiness higher than the average.

Table 1. Deviation of mean air temperature Δt and mean cloudiness Δc (per cent of hemisphere) during months of observation from the corresponding mean values at some meteorological stations.

Station	Year	Δt, °C			Δc, %		
		June	July	August	June	July	August
Ivalo	1959	2.0	−0.2	0.7	−1	−3	1
	1960	1.4	4.1	0.6	3	−10	1
Sodankylä	1959	1.1	−0.4	1.0	−3	−4	−1
	1960	1.2	3.0	0.6	0	−2	4
Vaasa	1960	2.2	0.8	−0.2	0	14	10
	1961	3.0	−0.7	−1.4	0	12	10
Jyväskylä	1960	2.0	1.0	0.2	0	9	8
	1961	2.8	−0.7	−1.1	−5	5	12
Jokioinen	1962	−2.2	−2.8	−2.5	6	18	18

According to the data presented above, it is possible that at Korvanen the meteorological conditions for evapotranspiration were better than normal and in the southern stations normal or lower than normal.

The following arrangement presents the ratio (per cent) of the sum of the precipitation for June, July and August and the corresponding mean for the same three months throughout the period 1931—1960. The mean values were calculated from data presented in the monthly weather report of the Finnish Meteorological Office. It is observed that at Korvanen the precipitation was less than the corresponding mean precipitation and in other measuring stations it exceeded the mean value.

	Korvanen		Möksy		Loppi
	1959	1960	1960	1961	1962
Measured three month precipitation as percentage of corresponding mean precipitation	80	70	130	150	130

Vegetation in sample containers

The vegetation cover of the sample containers is presented in Table 2. Except at Loppi, the figures correspond to the mean conditions of two similar lysimeters.

At Möksy in 1960 the vegetation cover in sample containers was fairly similar to that in the sample containers when the water depth had been kept at 2 cm in 1961.

On errors in evapotranspiration measurements

Several factors account for erroneus values of evapotranspiration obtained by the lysimetric method. One source of error affecting almost every lysimeter is the fact that the sample may differ from the natural state.

Although the sample may have been placed into the container extremely carefully, the micrometeorological conditions at the measuring point may not coincide with the average conditions of a large area. Inaccurate values may be obtained if the results of such measurements are generalized to cover large areas.

Since it is very difficult to eliminate errors caused by the above mentioned factors, some method should be tried for adjusting the measured values (KING, TANNER and SUOMI, 1956). Thus partly for this reason a calculation to estimate the terms of the water budget with the aid of water stage measurements was carried out.

In spite of the fact that the wooden poles supporting the sample containers were driven into the mineral soil, the height of the threshold of the container may have altered. On account of this, the distance from the rim of the sample container to the bog water surface was measured at four points every evening. In addition a water gauge was installed in the bog. By means of these measurements any movements of the sample container could be established. After it had been determined that a change of 1 mm in the sample container height corresponded to 0.5 mm in evapotranspira-

Table 2. Vegetation percentage in sample containers.

Station	Korvanen		Möksy		Loppi			
Year	1959	1960	1961		1962			
Water depth, cm	2	2	2	15	2	4	11	16
Andromeda polifolia	+			5	+	5	5	1
Betula nana				+				3
Calluna vulgaris				+			7	30
Empetrum nigrum						+		+
Vaccinium microcarpum						+	+	+
V. oxycoccos	+			2	1	2	1	1
Carex chordorrhiza			+	+				
C. lasiocarpa		+		4				
C. limosa	30	30	10	+				
C. magellanica	1							
C. pauciflora				1				
C. rostrata				1				
Eriophorum angustifolium		4	3					
E. gracile	1							
E. vaginatum				1	5	20	3	3
Rhynchospora alba			10					
Trichophorum caespitosum						+	20	
Drosera anglica	1		+		+			
D. rotundifolia	+			+				
Menyanthes trifoliata	20	10	1					
Rubus chamaemorus				+				2
Scheuchzeria palustris		+	1		2			
Sphagnum balticum					30	30	70	
S. compactum			60					
S. cuspidatum					70	30		
S. fuscum						15	10	80
S. Lindbergii	+							
S. papillosum				70				
S. parvifolium				30				3
S. rubellum						20	10	
S. tenellum							2	
Calliergon stramineum	20	1						
Cinclidium stygium	+							
Drepanocladus exannulatus		90						
D. fluitas	10							
D. procerus	50							
Polytrichum strictum				+				
Cladonia arbuscula						1		5
C. rangiferina						2		10
C. squamosa						1	3	

tion, the error in the result of evapotranspiration could be calculated. Usually the error due to movement of the sample container was small. The greatest error was measured at Möksy in summer 1960, the value of this error being 2 mm over a period of five days.

In Luutasuo the threshold was fixed to the rim of the container as the layer of peat was too thick to allow wooden poles to be used for supporting the sample containers. In this case an error could have been caused by tilting of the container during the summer. By measuring the distance from the rim to the level of the bog an attempt was made to determine this tilt. The accuracy of measurement, however, was not very good, and the results have not been adjusted in this respect.

In the case of precipitation during daytime, the amount of precipitation was added to the difference between the water poured into the sample container and the water measured in the overflow tank for obtaining evapotranspiration. The same method could not be applied to the precipitation occurring during the night as that part of the precipitation which flowed into the overflow tank and that which was retained in the container could not be distinguished. The amount of water flowing into the overflow tank owing to precipitation at night-time was dependent on the instant of precipitation and the amount of evapotranspiration during the night. The resulting error could be estimated with the help of a rain recorder and nightly evaporation measured with a water pan.

In general the lysimeters functioned quite satisfactorily, except that once at Korvanen in 1960 the plastic tube of one lysimeter became clogged during one five-day period. At Möksy in 1960 and 1961 the settling of the water level in two lysimeters was too slow, the trouble being caused by too thin a tube in the threshold. In this case rainy periods in particular caused errors in the daily values of evapotranspiration. Moreover, the sample containers were submerged too deeply into the bog so that they were occasionally under the bog water surface for short periods.

In addition to the errors mentioned above, there may be individual sources of errors connecting to each sample container. There may be systematic errors introduced by, for instance, erroneous determination of the water depth in the container, and random errors introduced by, for instance, errors in determination of the amount of water poured into or flowing out of the sample container. It might be possible to examine these kinds of error by means of several similar sample containers. However, it is not possible to make this kind of examination here as only two similar sample containers were used in the measurements at Korvanen and Möksy. Such an examination provides the possibility to calculate only whether there are systematic differences between results obtained with two similar sample containers.

In the following analysis daily values will be used, but not all measured values could be utilized. The results of those diurnal periods during which the amount of precipitation exceeded 4 mm were neglected, as the eventual difference in slowness of the different lysimeters may otherwise have an effect upon the results. Thus in each summer some 40 to 60 measured values could be utilized.

As the difference between the results of two similar sample containers is obviously rather small, it may be assumed that in the event that the difference is systematic it depends linearly on the evapotranspiration. In the equation

$$E^1_{zi} - E^2_{zi} = bE^m_{zi} + a + c_i \tag{1}$$

the symbols E^1_{zi} and E^2_{zi} have been used to indicate the evapotranspiration from sample containers 1 and 2 with the water level at a depth of z (cm). The symbol E^m_{zi} denotes the mean evapotranspiration from those sample containers with the water level at a depth of z. Further a and b are constants to be determined and c_i is a term due to random factors. The index i indicates the day in question. In the computations the quantity E^m_{zi} has been estimated from the quantity $(E^1_{zi} + E^2_{zi})/2$.

From the material the constants a and b may be determined by the method of least squares, for instance. The significance of the results may be tested with the aid of F- and t-tests. Should the results be significant, it may be concluded that there are systematic differences between the results of similar sample containers.

Results of computations are presented in Table 3. The symbol (*) indicates that the corresponding value is significant at the 0.05 level. It is observed that although precautions had been taken to install the containers in exactly similar ways, systematic differences may have occurred. As a consequence of the rather small correlation coefficient, the influence of the systematic factors upon the evapotranspiration difference is small. The differences are mainly due to random factors.

Table 3. Constants a and b in equation (1), correlation coefficient r between quantities $E^1_{zi} - E^2_{zi}$ and E^m_{zi}, standard deviation s_y of the quantity $E^1_{zi} - E^2_{zi}$ and number n of observations. Depth of water level is indicated by z.

Station	Year	z cm	b	a mm/day	r	s_y mm/day	n
Korvanen	1959	2	0.14*	−0.49*	0.33*	0.46	53
	1960	2	0.04	0.05	0.17	0.34	60
Möksy	1960	2	0.14	−0.61	0.20	1.16	54
	1961	2	0.26	−0.69	0.30*	0.94	43
	1961	15	0.16*	−0.23	0.33*	0.33	47

The standard deviation of the difference between the results of two similar containers is rather large. This fact cannot be accounted for by only errors in weighing the amounts of water, since the errors of weighing are less than 50 g, which corresponds to an evapotranspiration of 0.04 mm.

At Möksy in summer 1960 and 1961 the higher standard deviation s_y in the results from the wetter containers is due mainly to the fact that these lysimeters were too slow to allow measurement of the diurnal evapotranspiration.

Results of evapotranspiration

Table 4 presents the monthly totals of evapotranspiration measured during different observation periods. Except for the values found at Loppi, the figures present the mean results of two similar sample containers.

Table 4. Measured monthly evapotranspiration during different periods of measurements. The depth of water level is indicated by z.

Station	Year	z cm	Evapotranspiration, mm			
			June	July	August	Sum
Korvanen	1959	2	138*	112	74	324
	1960	2	105	136	63	304
Möksy	1960	2	126	104	86	316
	1961	2	92*	69	59*	220*
	1961	15	66	64	43	173
Loppi	1962	2	104	84	56	244
	1962	4	77	69	46	192
	1962	11	51	51	39	141
	1962	16	57	60	40	157

* *More than 10 per cent of the value is based on comparison.*

In Lapland the measured evapotranspiration has apparently been as great as or perhaps greater than in southern parts of the country, at Möksy and Loppi. This may be because the meteorological conditions for evapotranspiration in Lapland may have been more favourable, whereas in the stations in East Bothnia and Southern Finland they were less favourable than the average.

The values in Table 4 indicate that evapotranspiration depends on the depth of water level. This fact will be discussed in chapter 3.3.

3.2 Evapotranspiration and meteorological factors

Approximate methods for computing evapotranspiration

Several types of approximate equations have been developed for computing evapotranspiration and evaporation. THORNTHWAITE (1948) has presented a nomograph for computing the monthly potential evapotranspiration from data on temperature and day length. TURC (1954) has presented an equation for computing mean annual evapotranspiration provided that air temperature and precipitation are known. According to TURC, evapotranspiration may be computed also for periods of one month, in which case the yield of the vegetation has to be taken into account. In BLANEY'S (1954) method the potential evapotranspiration is calculated using the air temperature and the percentage of daytime hours. In KALWEIT'S method (see UHLIG, 1956) the effect of soil moisture on evapotranspiration has to be taken into account. The method for computing potential evapotranspiration presented by PENMAN (1954, 1956) is based on a combination of the energy budget and mass transfer approaches. In order to facilitate the computations MAKKINK (1957) simplified the PENMAN-formula. ROMANOV (1957) showed that evapotranspiration from a bog can be expressed as approximately proportional to the radiation balance. In addition, air moisture deficit has often been used to indicate the evaporative capacity of the air; this quantity has been compared by WÄRE (1947) even to the evapotranspiration from peat.

In the PENMAN-method (PENMAN, 1956) evaporation from an imaginary water surface is first computed. Then the computed evaporation is adjusted by an empirical coefficient to obtain potential evapotranspiration from the grass cover. This coefficient has a value 0.8 during June, July and August.

Evaporation from a water surface may be calculated from the equation

$$E_o^P = \frac{\Delta H + \gamma E_a}{\Delta + \gamma}, \tag{2}$$

where the subscript (o) in the evaporation E_o^P indicates that a free water level is in question and the superscript (P) that the result has been computed with the aid of PENMAN-formula. The slope of the saturated-vapour-pressure curve is denoted by Δ (mm Hg/°C) and γ is the psychrometer constant (0.49 mm Hg/°C). The quantity E_a is calculated from $E_a = (e_s - e_a)(a + bv)$, where the difference $e_s - e_a$ is the moisture deficit (mm Hg), and v indicates wind velocity (m/sec). The values of the constants a and b are 0.18 and 0.19 respectively. By using these values the quantity E_a is obtained in units of mm/day.

H is indicative of the radiation balance (mm/day), which has been computed by the following method. The incoming total radiation R_m was measured. The reflected radiation was estimated by assuming a 9% reflectivity for the water surface (BUDYKO, 1958, p. 40). As the effective longwave radiation could not be measured directly, it was computed from the equation (PENMAN, 1956)

$$R_{le} = \sigma T^4 \left(0.10 + 0.90\,\frac{n}{N}\right)(0.56 - 0.09\sqrt{e_a}), \tag{3}$$

where the BOLTZMANN constant is denoted by σ, air temperature by T (°K), vapour pressure by e_a (mm Hg) and the ratio of the actual time of sunshine to maximum possible time of sunshine by n/N.

An attempt to calculate the potential evapotranspiration E_p^P directly for the bog was also made. In place of the reflectivity of water surface the measured reflectivity 16% (VIRTA, 1960) of the bog surface was used; this value was measured at Korvanen in 1959 and corresponds to the mean reflectivity during June, July and August.

Utilization of the PENMAN-formula is complicated. Moreover, the effective long-wave radiation has to be computed by an approximate equation, thus particularly for short periods the error involved may be great. For instance, in Lake Hefner investigations (ANDERSON, 1952) it became evident that errors up to 10% in the values for one day emerged when using an equation similar to the equation (3). In this case it was assumed that the structure of the air mass is similar to the air mass in which the measurements for obtaining the values of constants have been done. However, there is no assurance of the accuracy with which the equation (3) fits the conditions in Lapland, for instance.

For the reasons mentioned above, MAKKINK (1957) has tried to simplify the PENMAN-formula. In measurements carried out in the Netherlands both the effective long-wave radiation and the quantity E_a were observed to be dependent on the quantity $\Delta R_m/(\Delta + \gamma)$. On this basis the potential evapotranspiration and evaporation may be expressed by

$$E^M = b\,\frac{\Delta R_m}{\Delta + \gamma} + a, \tag{4}$$

where the values of constants a and b are

evapotranspiration	E_p^M,	$a = -\,0.12$ mm/day
		$b = 0.61$
evaporation	E_o^M,	$a = -\,0.50$ mm/day
		$b = 1.01$

When using these constants, the total radiation R_m must be expressed in units of mm/day, and the result of calculation is in the same units.

Comparison of meteorological quantities with the measured evapotranspiration

Daily means and totals were used as the quantities in the equations (2)—(4). Evapotranspiration was calculated separately for each day. From these daily values five-day totals were calculated, and in the following these totals will be compared with the measured five-day evapotranspiration.

This comparison was carried out with the aid of the formula

$$E_z = bX + a\,. \tag{5}$$

E_z indicates evapotranspiration from the sample container with water level at the depth of z (cm). X denotes in turn the following quantities: total radiation R_m, radiation balance H, evaporation E_o^P in the equation (2), the corresponding evapotranspiration E_p^P for the bog, and the quantity $\Delta R_m/(\Delta + \gamma)$. Every quantity is expressed in units of mm/five-day. Results for the constants a and b calculated by using the method of least squares are presented in Table 5.

Every correlation and regression coefficients in Table 5 are significant at the level 0.01. The significance of terms a has been symbolized by (**) at the level 0.01 and by (*) at the level 0.05.

The following dependences between measured evapotranspiration and moisture deficit Δe (mm Hg) were computed:

Korvanen
1959 — 1960 $E_2 = 3.47\,\Delta e + 3.3$ (mm/five-day)
$n = 33,\ r = 0.88,\ s_e = 3.0$ mm/five-day

Loppi
1962 $E_2 = 4.79\,\Delta e + 2.7$ (mm/five-day)
$n = 20,\ r = 0.90,\ s_e = 2.1$ mm/five-day

From the values in Table 5 it can be seen that the best correlation may in general be obtained by using the MAKKINK variable. Evaporation values computed by the PENMAN-formula give the second best correlation. The poorest correlation is obtained by radiation balance. This last may be due to the fact that the long-wave radiation has been estimated from an approximate formula.

In addition it is seen that the term a is greater in the case where the evapotranspiration has been measured from sample containers with deep water level. This is due to the possibility that the dependence on the meteorological quantities is not linear for results from containers with deep water level. This aspect will be examined in the following chapter.

The values of the constant a are not zero even when the evapotranspiration has been measured from containers with shallow water levels. This may be due to errors in measurements of total radiation. The values of daily radiation measured at Loppi in 1962 have been compared with the corresponding values supplied by the observatory of the Finnish Meteorological Office at Jokioinen, which is situated about 50 km westward from Loppi. From the comparison it became evident that with a weak daily

Table 5. The regression line E_z versus X (equation (5)). Both quantities are expressed in mm per five-day. z is water depth, r the correlation coefficient, s_e the standard deviation of residuals and n number of observations.

Station	Year	z cm		X in equation (5)				
				R_m	H	E_o^P	E_p^P	$\frac{\Delta R_m}{\Delta+\gamma}$
Korvanen	1959–1960	2	b	0.60	0.80	1.00	1.08	0.92
			a, $\frac{\text{mm}}{\text{5-day}}$	−2.5	−1.6	−1.1	−1.0	−1.2
			r	0.92	0.81	0.95	0.94	0.96
			s_e, $\frac{\text{mm}}{\text{5-day}}$	2.5	3.8	2.1	2.1	1.8
			$n = 33$					
Möksy	1960–1961	2	b	0.50	0.67	0.77	0.81	0.71
			a, $\frac{\text{mm}}{\text{5-day}}$	−1.8	0.8	0.7	1.1	0.0
			r	0.84	0.69	0.77	0.76	0.82
			s_e, $\frac{\text{mm}}{\text{5-day}}$	2.6	3.5	3.0	3.1	2.8
			$n = 29$					
	1961	15	b	0.27	0.35	0.42	0.44	0.38
			a, $\frac{\text{mm}}{\text{5-day}}$	1.2	2.7	2.6	2.7	2.2
			r	0.74	0.64	0.75	0.76	0.75
			s_e, $\frac{\text{mm}}{\text{5-day}}$	2.0	2.3	1.9	1.9	2.0
			$n = 20$					
Loppi	1962	2	b	0.51	0.87	1.08	1.17	1.00
			a, $\frac{\text{mm}}{\text{5-day}}$	−2.2	−2.7	−2.9	−2.7	−4.7**
			r	0.89	0.88	0.94	0.94	0.96
			s_e, $\frac{\text{mm}}{\text{5-day}}$	2.2	2.3	1.6	1.6	1.3
			$n = 20$					
		16	b	0.19	0.35	0.46	0.50	0.42
			a, $\frac{\text{mm}}{\text{5-day}}$	2.9*	2.2*	1.7	1.8	1.0
			r	0.74	0.83	0.90	0.89	0.91
			s_e, $\frac{\text{mm}}{\text{5-day}}$	1.4	1.2	1.0	1.0	0.9
			$n = 20$					

radiation greater values of radiation were measured at Loppi than at Jokioinen. With a strong daily radiation smaller values were obtained at Loppi than at Jokioinen.

In March 1964 the hourly values measured by the actinograph were compared in the Finnish Meteorological Office in Helsinki with the corresponding values measured by the EPPLEY pyrheliometer. The comparison indicates the possibility that the calibration coefficient of the ROBITZSCH meter is not quite a constant, but is dependent on the radiation. This fact may account for the negative values of the constant a in Table 5.

It is evident that the results of the dependences of the evapotranspiration on the meteorological quantities are sensitive to errors in the measured radiation. However, over a long period the effect of these errors is obviously reduced. The size of the error in the radiation sum may be illustrated by the fact that in 1962, a 4% greater sum of radiation was measured at Loppi than in the Jokioinen observatory.

In the following a comparison between the measured evapotranspiration and the different meteorological quantities is given based on the expression

$$m_x = \frac{\Sigma E_2}{\Sigma X}, \tag{6}$$

where the summation has been taken over the period from June to August. X denotes in turn radiation balance H, evaporation E_o^P for water surface (equation (2)), corresponding evapotranspiration E_p^P for bog surface, evaporation E_o^M (equation (4)), and corresponding evapotranspiration E_p^M. Results are presented in Table 6.

Table 6. Ratio m_x in equation (6).

Station	Year	X in equation (6)				
		H	E_o^P	E_p^P	E_o^M	E_p^M
Korvanen	1959	0.88	0.97	1.05	0.98	1.50
	1960	0.88	0.92	1.00	0.96	1.46
Möksy	1960	0.79	0.89	0.97	0.87	1.34
	1961	0.62	0.70	0.79	0.72	1.09
Loppi	1962	0.73	0.89	0.97	0.85	1.28

It is observed that the quantity E_p^P is on the average approximately as great as the evapotranspiration E_2. Except for values measured at Möksy in 1961, the ratio between measured and computed evapotranspiration differs from the value 1.00 by at the most 5 per cent. Furthermore, it is seen that the fluctuations of the ratio m_x are least when X denotes either one or the other of the evapotranspiration values computed from the PENMAN-formula. Thus evapotranspiration E_2 may be considered as potential evapotranspiration.

Storage of heat energy

The methods for computing evapotranspiration presented previously are based entirely on the meteorological conditions. The energy being conducted into the soil has not been taken into account. This energy obviously depends on the time of the year, and a disregard of it may require different equations for calculating the evapotranspiration in different seasons. In this section this energy will be examined in the light of peat temperature measurements.

Temperature measurements of the bog peat were carried out by the method of FRANSSILA (1960, p. 10). Usually the measurements were carried out at five points beginning at a depth of 10 cm. The greatest depth of measurements at Korvanen was 140 cm in 1959 and 300 cm in 1960. At Möksy the measurements were carried out to a depth of 100 cm in 1960 and to a depth of 140 cm in 1961. At Loppi in 1962 the greatest depth of measurements was 290 cm. The thermometers were read once in every five days.

With the aid of the temperature measurements the change in heat energy in the bog can be calculated. For the heat capacity of the peat saturated with water the value 0.97 cal/cm^3/°C (PESSI, 1956, p. 76) was used.

We shall next present the calculated change of heat energy during a period of time from the beginning of observations at the end of May or in the beginning of June to the 18th of August. The reason for choosing the last mentioned date is the fact that the heat content of the bog was at that time near its maximum value. In order to be able to compare heat storage with the evapotranspiration, heat storage is expressed in units of mm/five-day. The following were obtained:

	Korvanen		Möksy		Loppi
	1959	1960	1960	1961	1962
The increase of heat content of the bog mm/five-day	0.8	1.5	1.5	1.2	0.9

The average increase of heat energy is about 10 per cent of the evapotranspiration E_2. Compared to the actual evapotranspiration from the bog, the ratio between heat energy storage and evapotranspiration is even greater.

3.3 Evapotranspiration and the depth of water level

Some experimental results

In the following the term local evapotranspiration will be used to indicate the evapotranspiration from a certain point, and the term areal evapotranspiration to indicate average evapotranspiration from a certain area.

In the Soviet Union the dependence of evapotranspiration on the depth of water level, both by using lysimeters and by energy budget studies, has been examined. It was found (ROMANOV, 1957) that the evapotranspiration from a bog may be calculated approximately from

$$E_s = \bar{f}(z) H \, , \tag{7}$$

where E_s is the areal evapotranspiration, H the radiation balance and $\bar{f}(z)$ a mean function depending on the depth of the water level. In this examination it is necessary to differentiate between mean and local functions f.

Again according to ROMANOV, equation (7) has to be used for periods longer than 10 days. If it is desired to examine evapotranspiration for shorter periods, a constant term should be added to this equation. This constant would represent evapotranspiration in the event that the radiation balance is zero, and is caused by energy being conducted from the air and from the soil to the evaporative surface.

Furthermore, according to ROMANOV (1960), the coefficient $\bar{f}$ depends on the time of the year. Were there an equation representing the evapotranspiration from a bog during June, July and August, the evapotranspiration during May could be computed from this equation by dividing the result by 1.84. Evapotranspiration during autumn can be calculated in a similar way, but now the result must be divided by 3.68. These results refer to a raised bog in Zelenogorski.

In this examination the evapotranspiration from the bog may be compared with the evapotranspiration from a sample container. Assuming that the radiation balance is independent of the water level depth and that an equation similar to formula (7) is suitable also for indicating evapotranspiration from a sample container, the following is obtained

$$\frac{E_z}{E_{z_0}} = \frac{f(z)}{f(z_0)} = f_{z_0}(z) \, . \tag{8}$$

E_{z_0} indicates the evapotranspiration from a sample container with the water depth z_0 (cm) and E_z the local evapotranspiration from a point with the natural water depth z (cm).

In equation (8) the function f_{z_0} may also be considered as a mean function over a certain area. In this case E_z is the mean evapotranspiration from that area, and the water depth z must be related to some fixed surface characteristic of the bog. By way of illustration, the function $\bar{f}_2$ is presented in Fig. 9 for an open fen and for a raised bog, as determined by ROMANOV (1960). His measurements have been carried out in bogs situated in Poland, and near Novgorod in the Soviet Union.

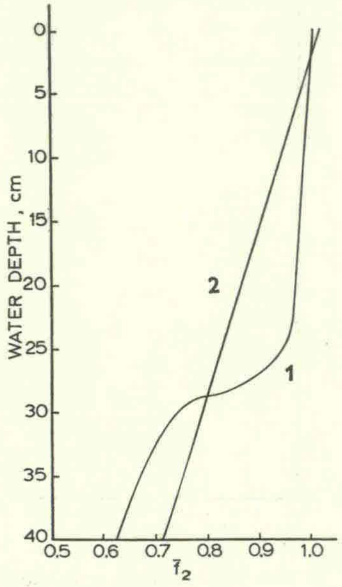

Fig. 9. The function $\bar{f}_2(z)$ in equation (8) according to ROMANOV (1960). 1) Raised bog, 2) open fen.

Determination of local function f_{z_0}

During the first two periods of measurement no systematic measurements were made of the influence of the depth of the water level on evapotranspiration. However, at Korvanen in 1959, the water level in one of the containers was lowered during the period of measurements. It was observed that lowering the water level by 1.3 cm caused about 16% reduction of evapotranspiration (VIRTA, 1960). This result refers to a situation where the water level depth is small.

At Möksy during the summer 1961 evapotranspiration was also measured from two sample containers with water level depth of 15 cm. At Loppi, 1962, evapotranspiration was measured from sample containers with water level at the depths of 2 cm, 4 cm, 11 cm and 16 cm. From these measurements the ratio of the evapotranspiration sum for different sample containers to the sum ΣE_2 has been calculated. The summation covers the entire period of measurements. Results are presented in Fig. 10. This ratio may be considered as an approach to the function f_2.

From Fig. 10 it is seen that, except for those values measured at Möksy, the dots seem to be distributed in such a way that the function $f_2 = \Sigma E_z/\Sigma E_2$ may be represented by the curve in the Figure. For water depths less than 2 cm the path of this curve is uncertain.

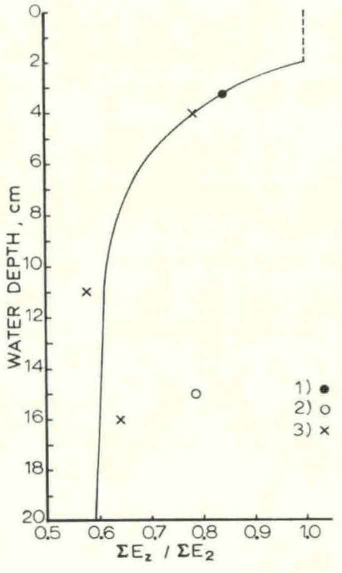

Fig. 10. The ratio $\Sigma E_z/\Sigma E_2$ at different measuring stations. 1) Korvanen 1959, 2) Möksy 1961, 3) Loppi 1962.

The value measured at Möksy in summer 1961 does not appear to agree with the presented curve f_2. This deviation may be due to differences in the peat samples, or perhaps has been caused by some systematic error in the measurements of evapotranspiration. For instance, the depth of the water level in the measurements of the evapotranspiration E_2 may exceed the measured value. Evidence for the existence of a systematic error in this evapotranspiration may be found from values in Table 6. The ratios m_x are smaller than others in this case. In this Table the ratio $\Sigma E_2/\Sigma E_p^P$ has a mean value of approximately 1.00, from which it follows that $\Sigma E_{15}/\Sigma E_2 \approx \Sigma E_{15}/\Sigma E_p^P$. The value of the last mentioned ratio at Möksy, 1961, was 0.62, which agrees well with the curve in Fig. 10.

Measurements by OVERBECK and HAPPACH (1957) with different kinds of *Sphagnum* peat, especially with *S. magellanicum* and *S. rubellum* show similar dependence of evapotranspiration on the water depth as presented in Fig. 10.

Construction of mean function $\bar{f}_{z_0}$

In the absence of direct measurements, it is necessary to construct the mean function $\bar{f}_{z_0}$ by computation, which may be carried out using the local function f_{z_0} in Fig. 10 and the distribution of the height of the bog surface. In this calculation the height should be compared to the water level. However the calculation is approximate. It is assumed here that evapotranspiration is dependent on only meteorological conditions and the depth of the water level. Thus, for example, the eventual influence of different vegetation is not taken into account. However, as the calculated dependences between the evapotranspiration and meteorological factors are practically similar in different bogs, the influence of different peat and different vegetation on the evapotranspiration is presumed to be small.

In the following it is better to use water stage W instead of the water depth z. This water stage was the actual quantity that was measured. In Fig. 11 the relation between the water depth and the water stage is shown. Between the water depth z and the water stage W there exists a relationship $z = W_0 - W$, where W_0 is the water stage corresponding to the bog surface.

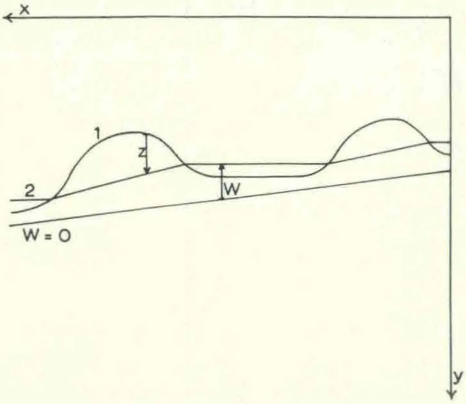

Fig. 11. The relation between the water depth z and the water stage W. The horizontal axis x and the vertical axis y are also presented. 1) Surface of bog, 2) water level.

The mean function $\bar{f}_2(W)$ near a water gauge may be computed from

$$\bar{f}_2(W) = \sum_{i=-\infty}^{\infty} g(W + i\Delta W) f_2(i\Delta W) . \tag{9}$$

The quantity $g(W + i\Delta W)$ denotes the relative area of a zone with the mean height h of its surface referred to the 0-level of the gauge in the interval $W + (i - 0.5)\Delta W \leq h < W + (i + 0.5)\Delta W$. The argument $i\Delta W$ of the local function f_2 indicates the mean water depth for the zone in question. The change in water level is denoted by ΔW, i is an index of summation and W is the water stage and is an integral multiple of the quantity ΔW. Equation (9) indicates a weighted average of the function $f_2(z)$.

Let us now calculate the mean function $\bar{f}_2$ for Luutasuo at Loppi. In October 1964, in Luutasuo, a surface levelling was carried out. This was done so that three squares with dimensions 40·40 m² were marked on the bog, the levelling readings for the sides of each square being taken at intervals of 2 m. The heights of the points of levelling were compared to the height of the water level in the centre of each square. Since in 1962 it was observed that the fluctuations of water level in different parts of the bog were similar, the distribution g of the bog-surface height may be computed as a mean from the results of different squares and this mean can be compared to the water stage measured at any gauge. In Fig. 12 this distribution compared to the water stage of the gauge 13 is presented.

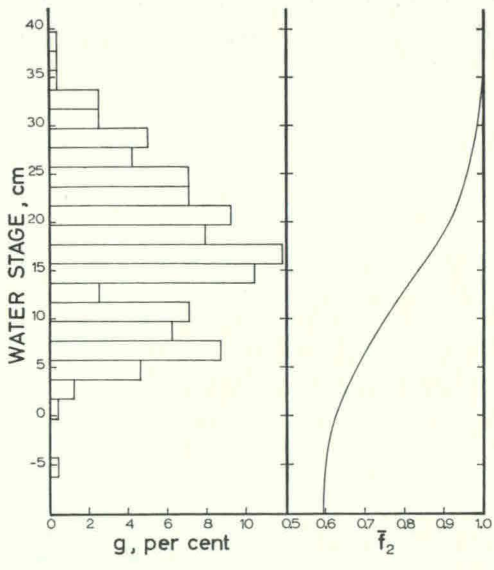

Fig. 12. The distribution g of the height of the bog surface in Luutasuo and the mean function $\bar{f}_2$ computed from the equation (9). Both g and $\bar{f}_2$ have been plotted against the water stage of gauge 13.

By using the measured distribution $g(W)$ and the local function $f_2(z)$ shown in Fig. 10, the mean function $\bar{f}_2(W)$ may be computed from equation (9). For the change ΔW the value 2 cm was used, and the result is presented in Fig. 12. It is observed that the computed curve is fairly similar to the curve for a raised bog presented in Fig. 9. However, the computed curve is more even, which may have been caused by the difference in the distribution g.

There may be several errors associated with the computed curve $\bar{f}_2(W)$. The disregard of the influence of different vegetation has been mentioned already. Another perhaps more essential defect is that the calculations are based on the assumption that the bog-surface height is fixed. However, in reality changes of bog-surface height were observed. After the 20th of June in 1960 at Korvanen a 4 cm fall of the bog-surface during 20 days was observed near the gauge. This may have been due to the melting of frost and to the 10 cm fall of the water level which occurred during the same interval. At Möksy no change in the height of the bog-surface was observed. Measurements at Loppi, 1962 and 1965, show that the bog-surface height in watery parts (in hollows) depends on the water stage. The dependence is such that a change of 1 cm in the water stage results in a 0.3 cm change in the bog-surface height.

The examples presented above show that the bog-surface height may change with the water stage. However, this will not be taken into account in the following computations.

It is clear from Fig. 12 that when the water stage of gauge 13 is between 0 cm and 20 cm, the function $\bar{f}_2$ is linear. The observed water stage in summer 1962 lay within this interval. From equation (8) one may now obtain

$$E_z = (e_1'z + e_2')E_{z_0}\,, \tag{10}$$

or

$$E(W,E_{z_0}) = (e_1W + e_2)E_{z_0}\,. \tag{11}$$

It is not always possible to compute the mean function $\bar{f}_{z_0}$ by the method presented earlier. In these cases the parameter e_1 was estimated directly from the results of two different sample containers. This parameter has been computed from

$$e_1 = \frac{\bar{E}_{z_1} - \bar{E}_{z_0}}{(z_0 - z_1)\bar{E}_{z_0}}\,, \tag{12}$$

where z_1 and z_0 are the water depths in two sample containers and $\bar{E}_{z_1}$ and $\bar{E}_{z_0}$ the corresponding time averages of evapotranspiration. Equation (12) may be derived from equation (10) since $e_1' = -e_1$.

The estimation of the error involved in determined areal evapotranspiration will be discussed at a later stage in connection with the water budget calculations. For this purpose equation (11) will be differentiated with respect to the quantities e_1, e_2 and E_{z_0}. The examination refers to systematic errors only, since no error is assumed in the measurement of the water stage W. Furthermore it is assumed that systematic errors in determination of the evapotranspiration E_{z_0} are directly proportional to this evapotranspiration. Now the error in the evapotranspiration for a certain area may be expressed in the form

$$\Delta E(W,E_{z_0}) = pE_{z_0} + qE_{z_0}W \,, \tag{13}$$

with

$$p = \Delta e_2 + e_2 d$$

and

$$q = \Delta e_1 + e_1 d \,,$$

where Δe_1 and Δe_2 indicate errors in the quantities e_1 and e_2, and d the relative error in the evapotranspiration E_{z_0}. The present purpose is to seek a method for determining the quantities p and q by means of water budget calculation.

On relations between results from different sample containers

According to equation (8) the results from two different sample containers are directly proportional. Whether such a direct proportionality really exists, will next be studied with the aid of measurements carried out at Möksy, 1961, and at Loppi, 1962.

From the values in Table 5 it is seen that the constant term a is greater for results from those sample containers with a deep water level. This in itself already refutes the validity of equation (8). This behaviour is examined by comparing results from different sample containers. The regression equation

$$E_z = bE_2 + a \tag{14}$$

is calculated by the method of least squares. In this case E_z denotes the evapotranspiration from a sample container with a water depth z (cm). Out of the data from Loppi every five-day total was used. Out of the data from Möksy results from days with precipitation less than 4 mm were used and the computations were carried out with daily values. The results are presented in Table 7.

Of the values in Table 7 that of the constant a is the most interesting. With the aid of t-test it was verified that, except for the smallest value of 1.6 mm/five-day, every constant a was significant at the level 0.01. The smallest value was significant at the 0.05 level.

The results may be taken as indicating that when the evapotranspiration E_2 is small, local evapotranspiration is independent of the water depth. Above a fixed value $(E_2)_0$ of the evapotranspiration the increase of the evapotranspiration E_2 is greater than the increase of the local evapo-

Table 7. Regression line of E_z versus E_2 (equation (14)). Evapotranspiration is in units of mm per five-day. Water depth is z, r the correlation coefficient and n the number of observations.

Station	Year	z cm	b	a mm/5-day	r	n
Möksy	1961	15	0.55	2.6	0.82	43
Loppi	1962	4	0.65	1.6	0.95	20
	1962	11	0.35	3.1	0.90	20
	1962	16	0.41	3.1	0.91	20

transpiration from a point with water depth greater than 2 cm. The evapotranspiration would be independent of the water depth up to an evapotranspiration of about $(E_2)_0 = 5$ mm/five-day or 1 mm/day.

Thus in order to clarify in a direct way the dependence of evapotranspiration on water depth with small evapotranspiration, days with evapotranspiration E_2 smaller than 1 mm/day should be examined. Unfortunately such values are scarce. Therefore, disregarding the critical value 1 mm/day the 12 days with least evapotranspiration E_2 were picked from the data from Loppi. The results of evapotranspiration of these days were arranged in two groups in such a way that one group included days with precipitation greater than 0.4 mm, and the other included days with precipitation less than 0.4 mm. There were 7 cases in the first group and 5 in the other. The greatest daily precipitation was 3.6 mm. Group means expressed in units mm/day have been computed for each sample container, and are presented in the following:

	z, cm			
	2	4	11	16
E_z in mm/day for				
– days with small precipitation	1.73	1.37	0.76	0.93
– days with abundant precipitation	0.68	0.74	1.09	0.99

It is observed that the values for days with small precipitation differ from each other by about the same ratio as the totals of evapotranspiration in Table 4. On the other hand, for days with abundant precipitation the situation is quite different. Even the order of magnitude of the evapotranspiration has changed. This change may be accounted for by the proposal that during rainy days the moisture content of surface peat may have increased, and the increase may be greater for results from sample containers with deep water levels. Had the daily evapotranspiration been 0.7 mm from each of the different sample containers during rainy days, then for instance the 16 cm thick peat layer would have retained about 0.3 mm of the amount of precipitation.

On the average, the mean water depth in the bogs was greater than 10 cm. For this reason the effect of the non-proportionality between local evapotranspiration and potential evapotranspiration may be minimized by selecting the evapotranspiration measured from the sample container with a deep water level for the standard evapotranspiration E_{z_0} in equation (8).

4. COMPUTATION OF WATER BUDGET

4.1 Runoff from bogs

Runoff and water level

Runoff is one term of the water budget. It may be defined as the amount of water travelling in the ground in unit time per unit catchment area. A particular differentiation is often made between surface, subsurface and ground-water runoff. In this investigation surface and ground-water runoff will be discussed, the latter being defined as runoff in deep layers of the bog.

According to the investigations to be presented later, the travelling of water would occur only in the surface layer of the bog. Thus on this basis we shall discuss the possibility of computing the runoff with the aid of the water level.

The system of co-ordinates presented in Fig. 11 will be applied, so that the co-ordinate y denotes the distance of the water level from a fixed horizontal level. Through a vertical strip of width dl an amount of water

$$dQ = dl \, \frac{dy}{dx} \int_{W_a}^{W} k(W) \, dW \tag{15}$$

travels per unit time. This equation is based on DUPUIT's assumption that the flow direction is almost completely horizontal. The direction of l is normal to both x and y, where x is in the direction of the horizontal projection of the flow. The quantity $k(W)$ indicates the hydraulic conductivity of the peat corresponding to the water stage W. The quantity W_a is the water stage when no water is flowing in the bog.

The quantity $\frac{dy}{dx}$ is equal to the water level inclination I, and it may be assumed that the inclination and the value of the integral in equation (15) does not change over a short distance Δl along the contour line. The amount of water travelling through a vertical cross section with a width Δl may now be computed from

(16) $$\Delta Q = \Delta l\ I\ K(W)\ .$$

That the inclination I is either a constant or depends only on the water stage W will be presented later. Further it will be presented in the light of available literature that the hydraulic conductivity and also the function K may be regarded as dependent on only the peat layer. From this argument it follows that the right hand side of equation (16) depends only on the water stage W. We are considering the catchment area for the part Δl of the contour line. When both parts of equation (16) are divided by this area the result gives us the runoff R. Taking into account what has been said above, the surface runoff may now be written in the form

(17) $$R = R(W)\ .$$

Hydraulic conductivity of peat

IVANOV (1957, p. 129) has presented measurements carried out in the Soviet Union. Several investigations (DUBACH, EVSTAFJEV, GERMANOV, GETMANOV) have shown that the following factors affect the hydraulic conductivity: 1) degree of humification, 2) vegetation cover, 3) pressure, 4) moisture content of sample at beginning of measurement, 5) flow duration, and 6) direction of flow.

In Finland HUIKARI (1959a) has measured relative hydraulic conductivity in field conditions. The auger in a hole method was used. It became evident that the travelling of water occurs mainly in the surface layers of an open fen.

SARASTO (1961, 1963) has measured the conductivity with the aid of peat samples. One result showed that the conductivity depends greatly on the degree of humification and the kind of peat. Furthermore the measured conductivity depends on the duration of the flow in such a way that the conductivity decreases during 1—4 days and stays almost constant after that. In addition the result was obtained that the conductivity depends on the pressure and direction of flow.

Owing to the large number of factors, it is evident that the conductivity of peat cannot be single-valued even at a point. However, as the travelling of water in a bog takes place mainly in poorly humificated surface layers, then according to IVANOV (1957, p. 146) in actual applications the conductivity at a point may be regarded as dependent on only the distance z from the surface to that point and on the type of bog and peat. By investigating peat samples of different bogs it has been obtained that the conductivity in the horizontal direction may be computed from

(18) $$k(z) = \frac{A}{(z + 1)^n}\ ,$$

where A and n are parameters dependent on the type of wet peatland and on the type of peat. With the aid of equations (15) and (18) IVANOV has derived a formula for the dependence of amount of flow on water depth.

According to IVANOV the function K in equation (16) may at a certain spot be assumed to be dependent on only the water stage W.

Inclination of water level

IVANOV (1957, p. 216) found that the fluctuations of water level in different parts of the bog are simultaneous and of almost similar order of magnitude. This would indicate that the inclination I of the water level is either a constant or depends on only the water level. This possibility may be examined for the measurements carried out at Möksy and Loppi. In these stations the water level was measured at many points. If changes of the inclination do occur, there should be changes in the difference between the water level elevation in two nearby water gauges and this ought to be detectable from the measurements.

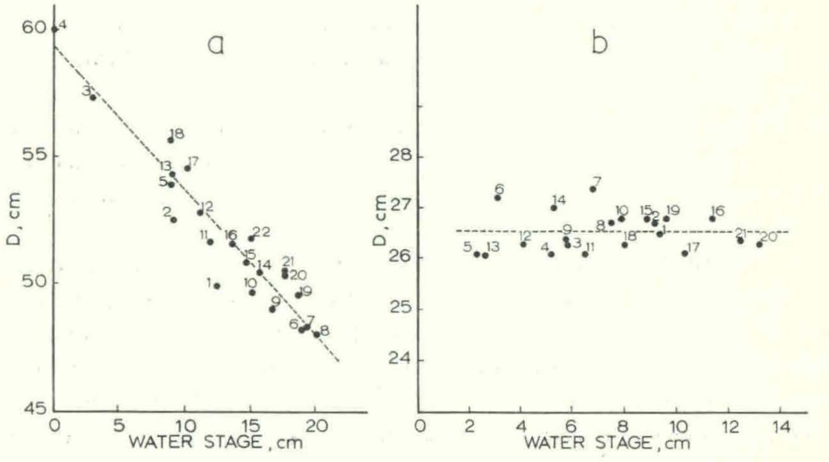

Fig. 13. a) Möksy 1961. The difference D between water-surface elevation at gauges 2 and 3 as compared to the water stage of gauge 3. b) Loppi 1962. The difference D between water-surface elevation at gauges 12 and 11 as compared to the water stage of gauge 11.

In Fig. 13a the difference D of the water level elevation at gauges 2 and 3 is presented as a function of the water stage of gauge 3. The distance between these gauges was 230 m. Water level readings were taken in the afternoon of every fifth day. The numbers in Fig. 13 indicate the order of measurement. It is seen that the points are evenly distributed. At least in this case the water level difference or inclination is not a constant but depends on the water stage. It is further observed that the inclination increases during the summer; this may be due to the development of vegetation.

In Fig. 13b the results from Loppi 1962 are shown. In this case the difference D of the water-surface elevation at gauges 12 and 11 as compared to the water stage at gauge 11 is presented. The distance between these gauges was 190 m. The water level was read here once a day and each point denotes an average of five measurements. It is possible in this case that the water level difference does not at all depend on the water stage. It is seen that these means of five measurements differ more from the curve $D =$ constant than the corresponding momentary values at Möksy differ from the curve $D = D(W)$. If one examines differences D for only periods with monotonically decreasing water stage, the distribution of the points is more even. These periods have the ordinals 4, 5, 11, 13, 17 and 20. During these periods the difference D varies between 26.1 cm and 26.3 cm.

On the basis of measurements presented previously, it may seem that the inclination of the water level at a fixed point is either a constant or depends on the water stage. The maximum deviation of the measured difference of water-surface elevation from the mean curve $D(W)$ is about 5 per cent. Thus equation (17) may be derived from equation (16).

On ground-water runoff from bogs

Equation (17) has been derived for only application to the surface runoff of a bog. We shall next consider in the light of the literature the ground-water runoff from bogs.

In several examinations (KIVINEN, 1948, p. 158) it has been concluded that the travelling of water in deep layers of a bog is very slow. If any flow occurs, the water is then travelling in hollows around the submerged tree trunks.

In the Soviet Union (IVANOV, 1957, p. 275) minimum runoff from areas with a large coverage of peatland have been compared with that for areas with less peatland. It was noticed that the mean annual minimum runoff for areas with many raised bogs was smaller than in other areas. These investigations were carried out in the European part of the Soviet Union.

According to IVANOV (1957, p. 284) the cessation of runoff in peatland is due to the fact that water may travel only in surface layers. When the water level falls below a critical value, the flow will cease. Due to evapotranspiration the water level may yet fall below this value. The flow will restart when due to precipitation the water level has risen sufficiently. In different bogs the water level at which cessation of flow occurs is 1—10 cm below the water level of the renewed flow. Besides, the critical water levels may vary. These just mentioned examples refer to raised bogs.

In Germany (EGGELSMANN, 1960) the ground-water runoff from a raised bog has been studied. Determination of the runoff was carried out by four methods: by measurement of 1) the minimum runoff of a small streamlet from the bog, 2) the minimum runoff from a small experimental basin, 3) the change of the moisture content of the bog during winter and 4) by the measured minimum hydraulic conductivity of peat. All these methods gave about the same value 1.2 $l/sec/km^2$ or 0.1 mm/day for the ground-water runoff.

Runoff measurements have been carried out by the Hydrotechnical Research Bureau of the Board of Agriculture in southeastern Finland at Ruoholahti (MUSTONEN and LAIKARI, 1961) in an area called Huhtisuo. About 29 per cent of this area is covered by open fen and 16 per cent by wet pine peat-moor, both in the natural state. Only twice, in 1954 and 1956, during the period 1936—1956 has the annual minimum runoff been greater than 0. In these two years the minimum runoff was 0.07 $l/sec/km^2$ or 0.01 mm/day. Since the peatland was drained in 1957, the minimum runoff has been much greater.

In Sweden ODÉN (1964) has compared the tritium content of water at different depths in a raised bog with that of rain water. Since it was noticed that the tritium content of water at the depth of 50 cm was very small, it was concluded that water can travel only in the surface layers of this bog.

At Möksy during the summers since 1961 runoff measurements have been carried out in the streamlet flowing from Pohjoisneva. A minimum runoff of 0.5 $l/sec/km^2$ or 0.05 mm/day was obtained during every summer except 1962, when the minimum runoff was 0.2 mm/day.

The above examples indicate that the annual minimum runoff for an area with much peatland is generally less than 0.2 mm/day. This fact indicates that if there is ground-water runoff in the bog its amount is small. From the data given by MUSTONEN (1965a) it may be found that in general the minimum runoff of small basins for the period from June to August is only slightly greater than the annual minimum runoff. This fact that the minimum runoff during a summer is small as compared to evapotranspiration will be used later in the water budget calculation.

The runoff computed from equation (17) refers to only a natural catchment. As the measurements carried out explain the hydrological properties of a peatland, this equation may be used particularly for entirely peated areas, which are generally found only in raised bogs. In the computation of the runoff from an open fen (ROMANOV, 1960), the runoff from the surrounding forest and the eventual ground-water runoff has to be taken into account.

4.2 Change of water content of bogs

Water level and water content of surface peat

In order to facilitate water budget calculation, the possibility of computing the change of water content of peatland by water level observations will now be discussed.

In Germany the relation between peat moisture content and water depth has been studied (EGGELSMANN, 1957). For an uncultivated, drained, raised bog a regression line $y = 927 - 0.872\,x$ was obtained between water content y (mm) of a 1 m thick peat layer and water depth x (cm). For the correlation coefficient the value $-\,0.895$ was obtained. The moisture content was determined by taking peat samples.

JÄRVINEN (1962) has measured water depth and water content of peat in Finland. No clear dependence was observed between water content and water depth.

Results by PAAVILAINEN (1963) indicate that with an increase of 1 cm in the water depth the volumetric moisture content of a 30 cm thick peat layer was reduced by 0.41 %. The correlation coefficient between the water level depth and the moisture content was found to be $-\,0.646$. Samples for determination of water content were taken after at least two preceding precipitationless days.

HEIKURAINEN (1964) and HEIKURAINEN, PÄIVÄNEN and SARASTO (1964) have presented results obtained by weighing peat samples buried in the soil. The sample had an area of 500 cm² and a height of 20 cm. It was found that for water depths from 8 cm to 48 cm the mean volumetric change of the moisture content was 0.5 % per 1 cm change in water level. A typical correlation coefficient between the water depth and sample moisture content was $-\,0.95$. The weighings were carried out at fixed time

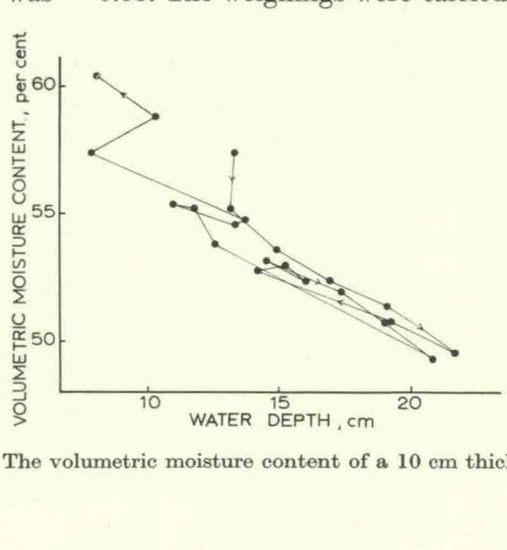

Fig. 14. Loppi 1962. The volumetric moisture content of a 10 cm thick peat sample and the water depth.

intervals regardless of the preceding precipitation. Moreover, it was obtained that in a drained bog, even below the ground-water level, the moisture content of the sample was dependent on the height of the water level.

At Loppi, 1962, similar measurements were carried out as by HEIKURAINEN (1964). Three brass net containers with a height of 10 cm and a surface area of 500 cm², were employed. The containers were weighed every fifth day. Fig. 14 shows results for one container. It is seen that the moisture content of this peat sample evidently depends on the water depth. The maximum deviation of points from the mean curve is 2 cm in water depth and 2 per cent in the moisture content. The latter figure corresponds to an amount of 2 mm water in a peat layer 10 cm thick.

Storage coefficient

According to the investigations presented in the previous chapter, the moisture content of the surface peat appears to be dependent on the water depth. Thus the storage coefficient (IVANOV, 1957, p. 114; TODD, 1959, p. 63) may be used to indicate the water storage both in the saturated zone and in the unsaturated zone. If S is the water content of the bog, then the coefficient of storage is

$$s(W) = \frac{dS}{dW} \, . \tag{19}$$

When the water stage changes from W_b to W_e the change of water content ΔS is obtained

$$\Delta S = \int_{W_b}^{W_e} s \, dW \, . \tag{20}$$

As in this analysis the terms of the water budget will be expressed in mm and water depth in cm, the unit of the storage coefficient is mm/cm.

The storage coefficient has been determined in bogs by computing the regression coefficient between water level change and precipitation less evapotranspiration (PRYTZ, 1932), by the water level change caused by precipitation (IVANOV, 1957; HUIKARI, 1959b; VOROB'EV, 1963) and by studying peat samples (IVANOV, 1957; HEIKURAINEN, 1963; PÄIVÄNEN, 1964). The drawback of the method based on the precipitation and water level change is that an unknown amount of precipitation will be retained by the surface peat. VOROB'EV has tried to account for this unknown by plotting the water level change against the precipitation and drawing a straight line through the points. In general, because of the precipitation amount retained by the surface peat, the straight line does not go through the origin. The storage coefficient is equal to the slope of this straight line.

For the water budget computation equation (20) will be rewritten in a more practical form. In general the water level did not change very much during a five day period. In this case it could be assumed that during the period the storage coefficient depended linearly on the water stage, or in other words that

$$s = s_1 W + s_2 . \tag{21}$$

When this expression is substituted into equation (20) the following will be obtained

$$\Delta S = s_1 \frac{W_e^2 - W_b^2}{2} + s_2 (W_e - W_b) = s\left(\frac{W_b + W_e}{2}\right)(W_e - W_b) . \tag{22}$$

In such cases as where equation (21) is valid, the change of water content of the bog will be found from equation (22) by using the water stage measurements at the beginning and at the end of the period in question. However, it appeared that it was not always possible to use this equation. In some cases the curve $s(W)$ will be complicated, and in such cases this curve is replaced by a broken line and the change in water content will be computed from equation (20).

4.3 Computation of water budget

Methods for computing water budget

The water budget equation for unit area of a catchment is

$$\bar{P} = \bar{E} + \bar{R} + \frac{\Delta S}{\Delta t} . \tag{23}$$

The quantities $\bar{P}$, $\bar{E}$ and $\bar{R}$ are time averages of precipitation, evapotranspiration and runoff, ΔS the change of water content of the whole area and Δt the length of the period. As regards to the runoff R this includes here both surface and ground-water runoff.

The water budget equation may be treated by several methods. In this investigation only methods where the water content is associated with the ground-water level will be examined.

WHITE in 1932 and TROXELL in 1936 (see TODD, 1959, p. 155) have used the daily course of the ground-water level for the determination of evapotranspiration. The daily evapotranspiration was computed by employing the storage coefficient, the maximum rate of rise of the ground-water level, and the change of the ground-water level during one day.

REMSON and RANDOLPH (1958) in the United States have treated the water budget equation in the following manner: evapotranspiration was calculated by the method of THORNTHWAITE (1948), the ground-water runoff was considered as linearly dependent on the ground-water level, and the change in the water content was considered proportional to the change of ground-water level height. The parameters of the water budget equation were determined by the method of least squares.

In the Netherlands VISSER and BLOEMEN (1959) have determined the water budget of an area from observations of water level in soil, the level in ditches, precipitation and potential evapotranspiration. The dependence of the terms of the water budget on the water level and precipitation were determined graphically.

In the Soviet Union ROMANOV (1960) has carried out water budget calculations for several bogs in order to solve the terms of the water budget for different years. The dependence of the runoff and the storage coefficient on the water level, i.e. the curves $R(W)$ and $s(W)$, were determined from peat samples. Evapotranspiration was computed from the radiation balance and the depth of the water level. The radiation balance was again determined by an approximate method. With these and the known precipitation it was possible to compute the water budget for different years.

In Finland HEIKURAINEN (1963) has applied the water budget method to the determination of the evapotranspiration from a forest in a case where there is no runoff. The ground-water coefficient (inverse value of the storage coefficient) was determined from soil samples. The evapotranspiration was determined by the daily change of the ground-water level.

Computation of water budget for bogs

In this section the method used in this investigation for computing the water budget for bogs will be discussed. The method is similar to that used by REMSON and RANDOLPH (1958) and VISSER and BLOEMEN (1959). The water budget equation (23), with the aid of equations (8), (13), (17), and (22), may be rewritten in the form

$$(24) \quad \bar{P} = \bar{f}_{z_0}(\bar{W})\bar{E}_{z_0} + \overline{R(W)} + s\left(\frac{W_b + W_e}{2}\right)\frac{W_e - W_b}{\Delta t} + p\bar{E}_{z_0} + q\bar{E}_{z_0}\bar{W} \, .$$

The first term on the right hand side represents the areal evapotranspiration, the second term the runoff, the third the change of water content of the bog and the last two terms take care of systematic errors in the measurement of evapotranspiration. The quantities $\bar{P}$, $\bar{E}_{z_0}$ and W have been measured. The function $\bar{f}_{z_0}$ was estimated from observations. The functions $R(W)$ and $s(W)$ and the parameters p and q ought to be solved from the water budget calculation.

In equation (24) the term corresponding to the areal evapotranspiration has been written as $\bar{f}_{z_0}(\bar{W})\bar{E}_{z_0}$ instead of $\overline{\bar{f}_{z_0}(W)E_{z_0}}$. The use of the former simplified expression does not introduce any significant error.

Were the curves $R(W)$ and $s(W)$ fairly simple, for instance the former a parabola and the latter a straight line, every parameter in equation (24) could perhaps be solved by the method of least squares, for example. As in general the observations cover only one summer it is impossible to use such a direct manner. More often the function $s(W)$ is inclined to be complicated.

For the determination of parameters p and q an additional condition should be known. In chapter 4.1 it was pointed out that the minimum runoff of an area with much peatland is rather small for almost every summer. Therefore the value of the function $R(W)$ ought to be small for small water heights and always positive. Making use of this fact the value of the parameter p may be determined.

The value of the parameter q, however, cannot be determined by this method. In the water budget computations the 0-level of the water gauge has been selected in such a way that this level corresponds approximately to the lowest measured water level during the period of measurement. Thus the last term in equation (24) is small for a small water stage. The influence of the term is not very great even during a period of high water stage, which fact will become evident from the following calculation. From equation (13) it appears that q is dependent on the systematic errors of evapotranspiration E_{z_0} and the parameter e_1. On the assumption that the values of both d and $\frac{\Delta e_1}{e_1}$ are 0.2, the error of the areal evapotranspiration can be computed. The values of the parameters used for Pohjoisneva at Möksy were $e_1 = 0.025/\mathrm{cm}$, $z_0 = 15$ cm, and the greatest measured water stage was 12 cm. The mean evapotranspiration E_{15} was 8 mm/five-day. In this case $qE_{z_0}W = 1$ mm/five-day will be obtained. Bearing in mind that the duration of the water stage 12 cm is short and that the values obtained from the water budget calculation are not very accurate, this calculated value is not great. For this reason it will be assumed that

$$q = 0 \,. \tag{25}$$

From equation (24) the functions $R(W)$ and $s(W)$ were calculated in the following manner. The observation material from the summer was grouped into five-day periods. If during the last day of any such period the precipitation exceeded 2 mm then this and the following periods were combined by computing the averages for five days. This procedure ascertained that the eventual local water currents in the bog after precipitation had no influence on the results of computations and that the water level at the end of the period was always falling. The significance of the last mentioned fact is that the surface peat is about in the same condition at the beginning and at the end of the period in question. The function $R(W)$ and $s(W)$ were solved by the method of successive approximations.

Initially it was assumed that $p = 0$. Since it was observed that the function $R(W)$ gave negative results for small water stages, the value -0.1 was given to the parameter p, and the computations presented earlier were repeated. It is possible that this value of p was still too great. Continuing in this way a series of curves $R(W, p)$ was obtained. At a certain value of p a small positive value of the runoff for a low water stage was obtained. From the corresponding curve the runoff for the bog was calculated and this value of p was used to adjust the areal evapotranspiration.

If the functions $R(W)$ and $s(W)$ proved to be simple, for instance polynomials of first or second degree, the constants of these curves were calculated by the method of least squares.

Calculation of water budget from meteorological observations

By employing the functions $R(W)$ and $s(W)$ it is possible to estimate the terms of the water budget even for periods without any particular observations in the bog in question. This type of computation has been carried out for several peatlands in the Soviet Union (IVANOV, 1957; ROMANOV, 1960; NOVIKOV, 1964).

In this investigation the following procedure will be used. By taking the water stage at the beginning and at the end of the period, the water budget equation (24) can be written in the form

$$(26) \quad \bar{P} = [\bar{f}_{z_0}(W_b) + \bar{f}_{z_0}(W_e)] \frac{\bar{E}_{z_0}}{2} + \frac{R(W_b) + R(W_e)}{2} + \frac{S(W_e) - S(W_b)}{\Delta t} .$$

Here the adjusted mean function $\bar{f}_{z_0}$ must be used. S denotes the water content of the bog and this may be derived by integrating the function $s(W)$ or $S(W) = \int^{W} s(W) dW$.

For calculating the water stage W_e at the end of the period from the water stage W_b at the beginning of the period, from the precipitation P and from the evapotranspiration E_{z_0} during the period, equation (26) can be rewritten in three parts

$$(27) \quad \begin{aligned} u &= \frac{\bar{f}_{z_0}(W_b)\bar{E}_{z_0} + R(W_b)}{2} - \frac{S(W_b)}{\Delta t} \\ v &= \bar{P} - u \\ v &= \frac{\bar{f}_{z_0}(W_e)\bar{E}_{z_0} + R(W_e)}{2} + \frac{S(W_e)}{\Delta t} . \end{aligned}$$

u and v are auxiliary variables.

It is possible to represent the group of equations (27) by a nomograph. Calculations presented later have been carried out for Δt equal to a five-day period.

The water stage for the entire summer may be calculated from the nomograph when the initial water stage has been estimated at the beginning of May. With the calculated water stage it is then possible to estimate the terms of the water budget.

5. APPLICATION OF WATER BUDGET CALCULATION

5.1 Water budget in Luutasuo at Loppi

Computation of functions $R(W)$ and $s(W)$

In Luutasuo 15 water gauges were in action during June, July and August 1962. These gauges were levelled in July 17 and Sept. 11 with reference to a fixed point at the edge of the bog. It was observed that some of the gauges had moved more than 1 cm during this time period; results from these gauges were neglected.

The gauges were placed in two lines. The water-surface elevations of these lines on July 17 are presented in Fig. 15. Also the positions and the height of the 0-levels of the gauges are marked in the Figure. At gauges 11, 12, and 13 the water level measurements were carried out once a day in the evenings. The other gauges were observed once every five days.

In Fig. 16 the difference $D_{13,i}$ between the water level elevation of gauges 13 and i are presented as a function of time. The gauges 3, 4, 7, 10, 11 and 12 are denoted by the index i. It is seen that the momentary fluctuation of the water level difference is generally less than 3 cm. If we compare the two topmost curves with the lowest curve, the last representing the water stage of gauge 13, it may be seen that the greatest changes in the water level difference often occur in connection with a sudden change of the water stage. This may be due to the fact that the method used for water level measurements did not give the true water level immediately after the precipitation. Provided that the water level difference D changes smoothly with time, it may be concluded from the topmost curve that an error greater than 3 mm in the water level difference is rare. This is important when estimating the errors associated with the water budget calculations.

The whole period of measurements consisted of 20 groups of five-day periods. On account of precipitation some of these periods had to be combined as described earlier. The remaining 14 periods may be used for water budget calculations. In addition to the five-day periods the readings include two periods covering 10 days each and two covering 15 days.

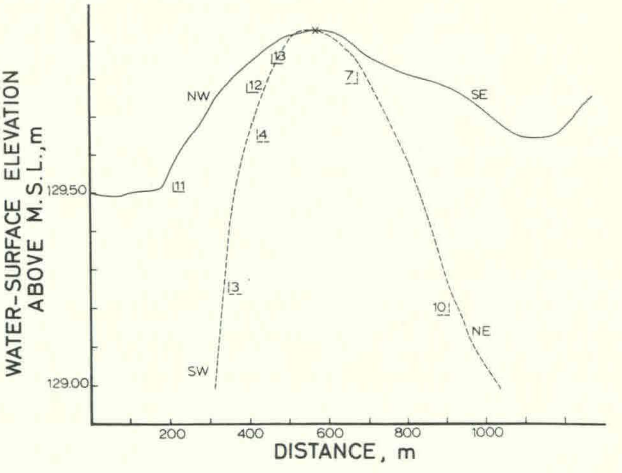

Fig. 15. The water-surface elevation of Luutasuo. Positions and 0-levels of water gauges 3, 4, 7, 10, 11, 12, and 13.

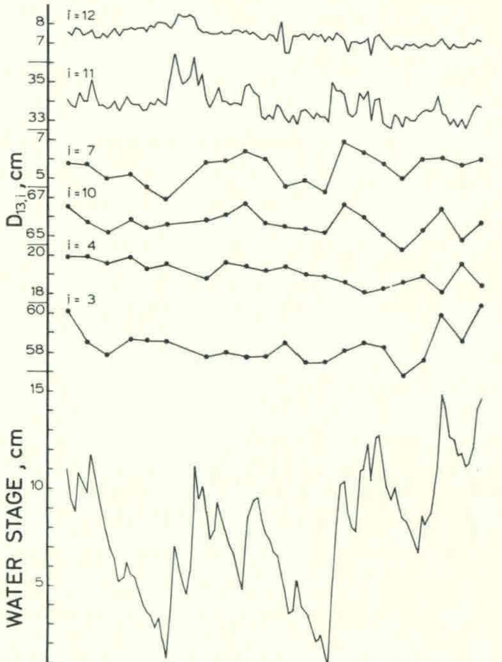

Fig. 16. The difference $D_{13,\,i}$ between the water-surface elevation at gauges 13 and i, and the water stage of gauge 13 as a function of time.

The calculations presented in chapter 4.3 were carried out initially with reference to gauge 13. It was observed that the functions $R(W)$ and $s(W)$ did not have a complicated form. Thus it might be possible to carry out the computations by the method of least squares. For this purpose the following formulae will be substituted into equation (24),

$$
\begin{aligned}
\overline{R(W)} &= a_1 \sum_i^n v_i W_i^2 + a_2 \sum_i^n v_i W_i + a_3 \,, \\
s\left(\frac{W_b + W_e}{2}\right) &= s_1 \frac{W_b + W_e}{2} + s_2 \,,
\end{aligned}
\tag{28}
$$

where n is the number of water stage observations W_i during the period. A smaller weighting factor v_i has to be allowed for the first and the last water stage observation than for other observations, as these were measured at the beginning and at the end of the period. For instance, in a normal case, when the period covered five days, there were six water level observations and the weights given were 0.1, 0.2, 0.2, 0.2, 0.2 and 0.1.

As in this case there are 14 equations and the number of unknown parameters is 5, namely a_1, a_2, a_3, s_1 and s_2, it is evident that the measured values have to be fairly well determined and the functions $R(W)$ and $s(W)$ should agree well with the model presented in equations (28). Although in principle it should be possible to compute also p and q by the method of least squares, on account of the insufficient observation material this cannot be applied. Thus these parameters will be estimated by the method presented earlier in chapter 4.3.

In the next part only results for gauge 11 will be discussed. These results may perhaps characterize best the average conditions of the entire bog. In this case the runoff may be considered to be dependent on only the water stage of gauge 11. The same water stage might be applied to the storage coefficient, too, as a clear dependence was observed between the water stages measured at different gauges. On the other hand, the change of water level should be calculated from observations of every gauge of the catchment. The mean change of water level was calculated from the observations of gauges 11, 12 and 13. The $\bar{f}_{16}$-function used for the calculation of the areal evapotranspiration was obtained from the $\bar{f}_2$-function in Fig. 12 multiplied by the ratio $\Sigma E_2 / \Sigma E_{16}$.

Table 8 indicates that, in spite of the scarcity of observation data, the computed parameters are in general significant. The significance level of 5 per cent corresponds to the t-value 2.3. Thus the significance of only parameter a_2 is uncertain. The multiple correlation coefficient R is high, because of the employed independent variable $P - E$, which has a large standard deviation.

4

Table 8. The values of parameters in equation (28) and the corresponding t-test with different values of parameter p. The standard deviation of residuals is s_e and R is the multiple correlation coefficient. The figures refer to gauge 11 in Luutasuo.

p	Coefficient *t-test*					s_e	R
	s_1 $\frac{\text{mm}}{\text{cm}^2}$	s_2 $\frac{\text{mm}}{\text{cm}}$	a_1 $\frac{\text{mm}}{\text{cm}^2\ \text{5-day}}$	a_2 $\frac{\text{mm}}{\text{cm 5-day}}$	a_3 $\frac{\text{mm}}{\text{5-day}}$	$\frac{\text{mm}}{\text{5-day}}$	
0.0	0.176	2.92	0.120	−0.48	−0.3	0.49	0.999
	6.6	*17.9*	*5.7*	*1.7*			
−0.1	0.171	2.92	0.122	−0.56	1.0	0.52	0.999
	6.1	*17.0*	*5.5*	*1.9*			
−0.2	0.165	2.91	0.125	−0.65	2.3	0.57	0.998
	5.4	*15.4*	*5.1*	*2.0*			

In the case at hand the value −0.1 of the parameter p gives a suitable curve $R(W)$. The minimum value 0.4 mm/five-day of the function $R(W)_{p=-0.1}$ will be obtained by the water stage 2.3 cm. During the periods covered by the calculation the mean water stage has varied between 2.3 cm and 10.3 cm.

We shall later calculate the runoff of the bog from the computed parabola for days when the water stage was greater than 2.3 cm. For days with the water stage between 2.3 cm and 0.0 cm the runoff is assumed to be 0.2 mm/five-day and for days with the water stage less than 0.0 cm the runoff is assumed to be 0.

The method used to adjust the measured evapotranspiration has various weaknesses. First, the runoff will have to be assumed to be small for low water stages. This has not been indicated in the present examination, but the assumption rests entirely on sources in the literature. Another weakness is that the value of the parameter p has been selected first of all on the basis of only a few periods of low water stage. It would be beneficial to study the reliability of the parameter p. From equation (24) it may be seen that the independent variable $P - E$ as employed is equal to the runoff R when the water level change is $\Delta W = 0$. The reliability of p may thus be estimated by computing the standard error of the computed $(P - E)_{W, \Delta W=0}$ for a small value of W. This standard error may be determined from the variances and covariances of the estimated parameters a_1, a_2, a_3, s_1 and s_2 (CRAMÉR, 1961, p. 223). The computations were carried out for water stages 2 cm and 4.6 cm; the former water stage corresponds to the minimum of the $R(W)$-curve and the latter to the water stage with $R_{p=0} = 0$. The selection of the value of the parameter p is based first of all on the measurements carried out within this water stage interval. The results were $s_e = 1.3$ mm/five-day and 0.7 mm/five-day with 9 degrees of freedom. The first mentioned result is equal to the difference $R(2 \text{ cm})_{p=-0.1} - R(2 \text{ cm})_{p=0}$. From this it may be concluded that the error in the paramameter p may be of the order of 0.1.

Terms of the water budget during measurement period

Table 9 presents the terms of the water budget of each five-day period calculated from results presented earlier, as well as the residuals computed from the water budget equation (24). The last column in Table 9 shows the ordinal j which indicates the period of the water budget calculation including the five-day period in question.

From the residuals it may be seen that they are in general small. Greater errors occur only in the combined periods. It has been mentioned earlier

Table 9. Precipitation P, areal evapotranspiration E, runoff R, change in water content ΔS and residual Δ for gauge 11 in Luutasuo 1962. The ordinal j indicates the period of the water budget calculation which includes the five-day period in question.

Period	P mm/5-day	E mm/5-day	R mm/5-day	ΔS mm/5-day	Δ mm/5-day	j
May 26–30	10.4	9.4	5.9	−5.1	0.2	1
31–June 4	8.3	9.5	6.2	−7.7	0.3	2
5–9	2.1	10.0	1.9	−7.6	−2.2	3
10–14	1.8	7.9	1.0	−8.1	1.0	3
15–19	3.8	11.3	0.4	−8.3	0.4	4
20–24	19.6	10.7	0.5	8.3	0.1	5
25–29	27.8	5.8	3.3	18.6	0.1	6
30–July 4	10.8	11.2	3.7	−4.2	0.1	7
5–9	9.3	10.2	1.9	−3.2	0.4	8
10–14	14.6	5.9	4.3	5.7	−1.3	8
15–19	0.1	9.1	2.6	−11.6	0.0	9
20–24	5.6	9.9	0.8	−5.8	0.7	10
25–29	2.9	9.3	0.4	−5.9	−0.9	11
30–Aug. 3	43.2	7.8	2.6	31.5	1.3	12
4–8	17.4	8.7	5.8	3.7	−0.8	12
9–13	21.8	7.4	10.4	4.3	−0.3	12
14–18	3.8	8.1	8.2	−12.2	−0.3	13
19–23	6.4	8.4	4.4	−1.8	−4.6	14
24–28	44.4	4.7	7.3	29.9	2.5	14
29–Sept. 2	10.0	7.5	15.1	−14.0	1.4	14
3–7	22.6	(4.0)	13.2	12.5	−7.1	–
Sum	286.7	176.8	99.9	19.0	−9.0	

that in the water level change greater errors than 0.3 cm hardly ever occur. This corresponds on the average to 1 mm water. In this case the change of the water stage has been computed as a mean from observations of three gauges, and the greatest error may perhaps be 0.5 mm of water. There are in Table 9 only five combined periods during which the absolute value of the residual is greater than that. Accordingly it may be concluded that the functions $R(W)$ and $s(W)$ are applicable to the model presented in equations (28).

It is possible to compare the computed values of runoff with the measured ones. However, such a comparison is uncertain because the runoff measurements were carried out in a streamlet flowing from Luutasuo, and the raised bog is only a small part of the catchment of the measurement place.

From a topographic map (scale 1:20 000) the catchment area of the measuring weir was estimated as 3.4 km^2. This area consists of 50 per cent pine and spruce forest and arable land, 20 per cent marginal slope and lagg of the raised bog, 8 per cent ponds and 22 per cent of the actual raised bog suitable for water budget calculations.

The runoff measurements were carried out with a THOMPSON weir from June 20 to September 7. During this period the measured runoff was 26 mm. During the same period the amount of water in the ponds increased by 11 mm. The total runoff is thus 37 mm. The computed value of 84 mm for the runoff from the bog during the same period is considerably greater than the measured value.

The difference between the measured and computed runoff values may be explained by several factors. For instance, the runoff of only the raised bog has been computed, whereas the runoff of other parts of the catchment area is not known; nor is the evaporation from the ponds known. Moreover, the ground around Luutasuo consists of a coarse type of soil which allows water to penetrate easily. Part of the precipitation may have reached the ground-water directly without passing through the runoff measurement place.

Calculation of water budget from meteorological observations

A nomograph was constructed in accordance with equation (27) for calculating the water stage of gauge 11. The nomograph is presented in Fig. 17.

When employing the nomograph the initial water stage must be known. In the following computations this has been estimated as 20 cm at the beginning of May. The possible error in this estimated water stage has only a small effect on the water stage during the summer. Let us illustrate this by taking a sample calculation. On 25th May, 1962, the measured water stage was 11.0 cm. By means of the nomograph the computed water stage of 4.0 cm after 20 days is obtained. Taking 16.0 cm for the initial water stage the computed water stage is 4.7 cm after 20 days. The difference is now only 0.7 cm. From this it may be concluded that the influence of the eventual error in the initial water stage will disappear in course of time, and it is possible that the computed water stages during June, July and August are almost independent of the error in the initial water stage.

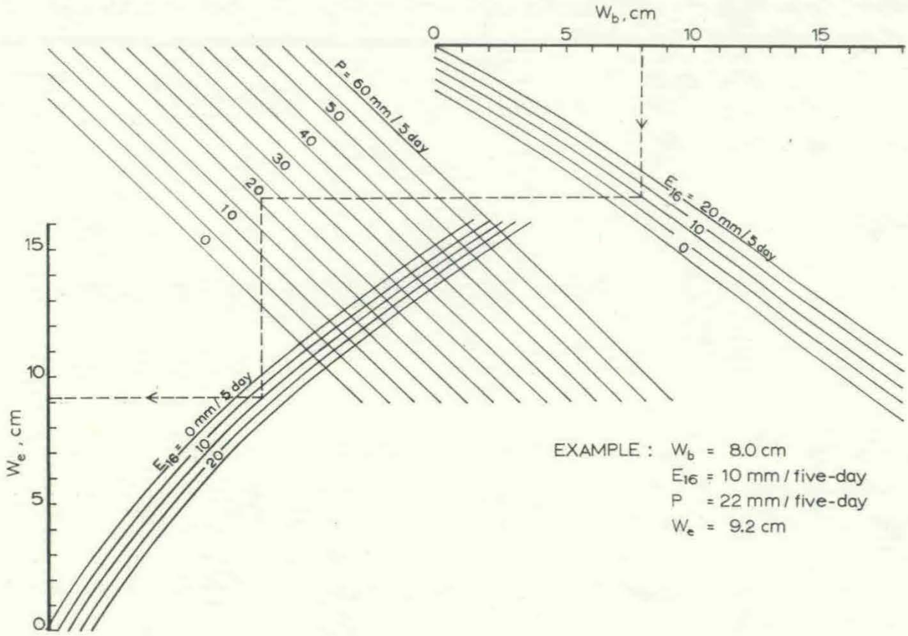

Fig. 17. Nomograph for computing water stage. W_e of gauge 11 at the end of a five-day period. The water stage W_b at the beginning of the period and the five-day precipitation P and evapotranspiration E_{16} have to be known.

During several summers the computed water stages were often below 0 cm. In such events the computation of the water stage was based on three assumptions, 1) $R(W < 0) = 0$, 2) $s(W < 0) = s(0)$ and 3) $\bar{f}_{z_0}(W < 0) = \bar{f}_{z_0}(-4 \text{ cm})$. The reasoning behind assumption 1 has been discussed earlier. By consideration of Fig. 12 assumption 3 may be made. On the basis of these assumptions the water budget of the bog is not dependent on the water stage with $W < 0$. In this case the storage coefficient s has no influence on the evapotranspiration and runoff; it affects only the water stage during a period with $W < 0$. Hence also assumption 2 may be justified.

The total precipitation for five-day periods required in the calculation have been obtained from regular precipitation stations of the Finnish Hydrological Office and Meteorological Office not far from Luutasuo. The nearest of these stations is situated about 10 km southeast of Luutasuo.

This considerable distance to the precipitation stations may cause errors in the computed water stage.

For the estimation of evapotranspiration E_{16} the measured values of this quantity have been compared with the quantities calculated by the meteorological observations carried out at Jokioinen observatory in 1962. This observatory is situated about 50 km westward from Luutasuo. For the quantities to be compared the MAKKINK variable calculated from the monthly weather report data of the Finnish Meteorological Office, and the evaporation measured with the U.S. Weather Bureau Class A land pan given by the Finnish Hydrological Office were chosen. The following regression lines were calculated

$$E_{16}^{M} = 0.37\,\frac{\Delta R_m}{\Delta + \gamma} + 2.1 \tag{29a}$$

$$n = 20\,,\quad r = 0.89\,,\quad s_e = 1.0 \text{ mm/five-day}\,,$$

$$E_{16}^{pan} = 0.36 E_{pan} + 3.2 \tag{29b}$$

$$n = 20\,,\quad r = 0.85\,,\quad s_e = 1.1 \text{ mm/five-day}\,.$$

E_{pan} is the evapotranspiration from the Class A pan and other symbols are defined in chapter 3.2. Both the MAKKINK variable and E_{pan} have to be expressed in mm/five-day.

The above results refer to only one summer. Therefore it is uncertain whether the same dependence may be obtained during other summers. From Table 6 it is observed that the ratio $\Sigma E_2/\Sigma E^P$ is almost a constant at different measurement stations and for different years. Also in Finland the computation method for potential evapotranspiration of U.S. Weather Bureau based on the PENMAN-formula has been shown (MUSTONEN, 1964) as capable of indicating evaporation from a Class A pan. Thus it may be concluded that evaporation from the Class A pan indicates well also evapotranspiration E_2. It has still to be examined how well the MAKKINK variable could indicate this evapotranspiration during different years. Therefore the ratios $\Sigma E_{16}^{M}/\Sigma E_{16}^{pan}$ for different summers will be calculated. The summations include the period from June to August. The results are

	1961	1962	1963	1964	1965
$\Sigma E_{16}^{M}/\Sigma E_{16}^{pan}$	1.03	1.00	0.96	0.96	0.98

It can be seen that these values do not differ much from unity. Accordingly the evapotranspiration E_{16} may be calculated from either of equations (29).

With the help of the nomograph the water stage of gauge 11 has been calculated from measurements carried out in Luutasuo in 1962. It was noticed that the absolute value of the deviation of the computed water stage from the measured value was in three cases greater than 1 cm. Altogether there were 21 cases.

Later an attempt was made to test the method for computing the water stage. In the following water stage measurements and the corresponding calculated values for gauge 11 will be presented.

	Nov. 5 1964	June 6 1965	June 29 1965	Aug. 13 1965	Sept. 9 1965	Oct. 15 1965
Computed W_{11}, cm	7.5	−4.0	−12.5	11.5	5.0	5.5
Observed W_{11}, cm	4.5	−2.7	−13.4	6.6	0.6	4.1

According to the computed results it is evident that there may be some systematic error. The eventual source of error is in the estimated precipitation and evapotranspiration E_{16}.

In summer 1966 two rain gauges situated about 1.5 km northwest and southeast from the centre of the bog were used to estimate the precipitation of the bog. Taking the observed water stage 6.3 cm on May 27 (calculated water stage was 4.0 cm) for the initial water stage a better consistency between measured and computed water stages is obtained as may be seen from the following values:

	Water stage, cm				Water stage, cm		
Day	Obs.	Comp.	Diff.	Day	Obs.	Comp.	Diff.
27 May	6.3			1 July	−9.5	−7.6	−1.9
1 June	4.3	4.7	−0.4	6	−7.2	−6.0	−1.2
6	5.5	5.5	0.0	11	−10.8	−7.9	−2 9
11	1.9	2.3	−0.4	16	5.7	7.2	−1.5
16	−3.5	−2.6	−0.9	21	1.1	5.1	−4.0
21	−7.4	−7.6	0.2	26	−3.2	0 4	−3.6
26	−9.3	−8.8	−0.5				

From the water stage computed for gauge 11, evapotranspiration and runoff may be computed. Table 10 presents the results for different years. Only the results of summer 1962 are based on the measured water stage.

Table 10. The sums of precipitation P, evapotranspiration E and runoff R for periods from June to August in Luutasuo.

Year	ΣP mm	ΣE mm	ΣR mm
1961	320	170	110
1962	250	150	70
1963	200	160	10
1964	160	160	0
1965	220	160	50

5.2 Water budget in Pohjoisneva at Möksy

Computation of functions $R(W)$ *and* $s(W)$

The computations for the water budget determination for Pohjoisneva have been presented earlier (VIRTA, 1962). However, then the method for adjustment of the measured evapotranspiration presented in chapter 4.3 was not used.

It appeared that the functions $R(W)$ and $s(W)$ were more complicated than those in Luutasuo. Therefore the water budget equation could not be solved by the method of least squares. The computations were carried out by the method of successive approximations.

In summer 1961 the water stage observations were carried out at five gauges (Fig. 2b). The calculations were carried out separately for gauge 1 and gauges 2—5. The functions $R(W)$ and $s(W)$ associated with the catchment of gauge 1 were determined relative to the water stage of this gauge. The water level change in that area was considered equal to the change of water stage of gauge 1. The function $R(W)$ associated with the catchment of gauge 3 was determined relative to the water stage of this gauge and the function $s(W)$ was computed from the mean change of the water stages of gauges 2—5.

For computing the areal evapotranspiration equation (11) was used. The parameter e_1 was determined from equation (12). 15 cm was chosen for the depth z_0 and 2 cm for z_1. The parameter e_2 was determined by the water budget calculation. It then follows that

$$E(W,E_{15}) = (0.80 + 0.025W)E_{15} \,, \tag{30}$$

where W is the water stage (cm) of gauge 1.

This type of method of determining the equation for the computation of the areal evapotranspiration is subject to several errors. The surface of Pohjoisneva is more even than that of Luutasuo. For this reason the form of the function $\bar{f}_{z_0}$ may be nearer to curve f_2 in Fig. 10 than to a straight line. Moreover, the systematic error in the evapotranspiration E_2 discussed in chapter 3.3 affects the results. As measured this evapotranspiration was too small. However, it was considered better to use this too small value, as the corresponding function $\bar{f}_{z_0}$ agrees better with the curve in Fig. 10.

The levelling of the fen surface has not been carried out in the same manner as was used in Luutasuo. It is known only that the surface near the gauge 1 is at a height corresponding to water stage 13 cm.

The whole measurement period consisted of 21 five-day periods. Because precipitation occurred during the last day of some of the five-day periods some successive periods had to be combined. Thus in the water budget calculation 15 periods may be used. Besides five-day periods there are three covering 10 days and one covering 20 days.

Fig. 18 presents the computed curves $R(W)$ and $s(W)$ for gauge 1. The points corresponding to the runoff have been plotted from the mean water stage of the period. It is seen that there are great deviations from the mean curves. One factor affecting the deviations is that in this case one single gauge illustrates the fluctuations of the water content of a great area. The gauge is situated in one side of the area, which consists of both peatland and forest.

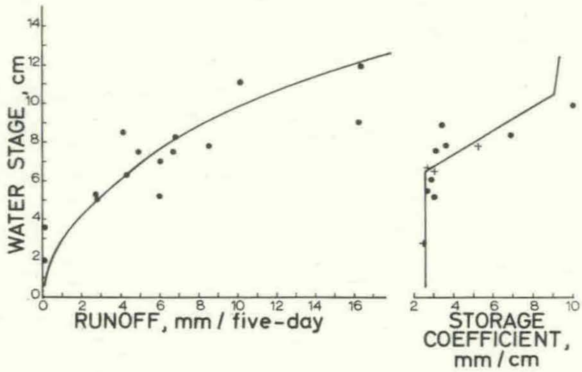

Fig. 18. The dependence of runoff and storage coefficient on the water stage of gauge 1. Storage coefficient with $W_e - W_b \geq 1.0$ cm is denoted by (+).

Calculations were carried out also for gauges 2—5. It appeared that the runoff in August was smaller than for the same water stage during June and July. This has been explained by the damming caused by the development of the vegetation cover. The runoff was fixed relative to the water stage of gauge 3 situated in an area with many open channels between the strings.

Terms of the water budget during measurement period

By using the curves presented in Fig. 18, the terms of the water budget have been calculated for the catchment of gauge 1 for each five-day period in question and are presented in Table 11.

Here the error of the parameter p may be greater than that computed for Luutasuo. The selection of the value of this parameter is based mainly on the periods 2 and 3.

Calculation of water budget from meteorological observations

For water stage computation considering gauge 1 a nomograph was constructed based on equation (27). In principle the nomograph is similar to

Table 11. Precipitation P, areal evapotranspiration E, runoff R, change in water content ΔS and residual Δ for gauge 1 in Pohjoisneva 1961. The ordinal j indicates the period of the water budget calculation which includes the five-day period in question.

Period	P mm/5-day	E mm/5-day	R mm/5-day	ΔS mm/5-day	Δ mm/5-day	j
May 26 – 30	1.6	4.4	5.2	−8.8	0.8	1
31 – June 4	1.0	10.2	1.5	−9.3	−1.4	2
5 – 9	18.9	13.9	0.3	4.9	−0.2	3
10 – 14	30.0	6.0	3.1	15.5	5.4	4
15 – 19	40.1	10.7	10.3	24.4	−5.3	4
20 – 24	36.0	11.2	15.6	8.1	1.1	5
25 – 29	22.8	12.3	15.6	−5.3	0.2	5
30 – July 4	8.1	13.5	13.1	−15.5	−3.0	6
5 – 9	9.6	10.2	8.3	−16.8	7.9	7
10 – 14	3.6	7.6	4.3	−8.3	0.0	8
15 – 19	14.0	8.3	3.0	−0.3	3.0	9
20 – 24	14.6	12.6	2.8	−0.8	0.0	10
25 – 29	36.4	10.2	4.6	18.5	3.1	11
30 – Aug. 3	7.3	8.0	7.4	−2.8	−5.3	11
4 – 8	8.7	8.8	6.4	−8.6	2.1	12
9 – 13	0.8	5.9	3.1	−7.8	−0.4	13
14 – 18	10.0	7.6	1.6	2.3	−1.5	14
19 – 23	43.9	5.1	8.1	28.3	2.4	14
24 – 28	15.6	6.5	12.1	3.7	−6.7	14
29 – Sept. 2	2.6	5.9	10.6	−14.5	0.6	14
3 – 7	0.7	7.2	7.3	−10.6	−3.2	15
Sum	326.3	186.1	144.3	−3.7	−0.4	

that in Fig. 17. The water stage of gauge 3 was determined from the water stage computed for gauge 1 with the gauge relation curve. From the water stages thus obtained the runoff from the bog was then computed.

For several summers the calculations show the water stage to have been below 0. For such periods the calculation of the water stage was based on the assumptions 1) $R(W < 0) = 0$, 2) $s(W < 0) = s(0)$ and 3) $\bar{f}_{z_0}(W < 0) = \bar{f}_{z_0}(0)$. These assumptions are about the same as in the calculation for Luutasuo.

Except for the measurement periods in 1960 and 1961, the five-day sums of precipitation required in the calculation have been obtained from the regular precipitation station of the Finnish Meteorological Office situated about 2 km west of the evapotranspiration station. The precipitation data for 1960 and 1961 at the regular station have been compared with the corresponding data obtained from a rain gauge situated in Pohjoisneva.

The comparison material consisted of 34 such five-day periods during which the precipitation exceeded 5 mm at the regular station. It was observed that the absolute value of the difference between the five-day sums of precipitation in the two rain gauges was smaller than 1 mm in 35 per cent of the cases and smaller than 5 mm in 94 per cent of the cases.

A Class A pan of the Finnish Hydrological Office situated 2 km west of the evapotranspiration station was used for estimating evapotranspiration E_{15}. A coefficient 0.59 was used to convert the pan evaporation to the evapotranspiration E_{15}; this coefficient was obtained from measurements in summer 1961. When using this method for computing the evapotranspiration, errors of 2—3 mm may occur in a five-day value of E_{15}.

The water stage for gauge 1 has been computed from precipitation and evapotranspiration E_{15} measurements in 1961. It was observed that the maximum and minimum difference between the computed and measured water stages was 1.8 cm and −1.4 cm, respectively.

In 1962, during the period from May 5 to September 27, water stage measurements at gauge 1 were carried out. The 0-level of gauge 1 was determined by levelling in the summers of 1961 and 1962. However, since the levellings were not sufficiently accurate, the observed water stages have been compared with the water stage measurements of Poikkijoki streamlet during rainless periods in the summers of both 1961 and 1962. Due to this comparison a difference of scarcely more than 0.5 cm occurs between the 0-levels of gauge 1 during these two summers.

When computing the water stage for summer 1962 the observed water stage 12.9 on May 4 has been taken for the initial water stage. In the following the observed and the computed water stages are presented.

Day	Water stage, cm			Day	Water stage, cm		
	Obs.	Comp.	Diff.		Obs.	Comp.	Diff.
4 May	12.9			19 July	2.7	4.2	−1.5
7	13.2	12.5	0.7	24	4.5	6.5	−2.0
11	11.4	10.7	0.7	29	1.0	4.0	−3.0
16	8.4	7.2	1.2	3 Aug.	5.3	6.8	−1.5
21	7.4	6.6	0.8	8	7.6	6.5	1.1
26	6.6	6.8	−0.2	13	8.5	9.7	−1.2
31	7.3	8.0	−0.7	18	6.4	7.8	−1.4
6 June	5.4	2.6	2.8	23	7.7	6.0	1.7
10	4.8	2.5	2.3	28	8.9	9.7	−0.8
15	4.7	6.4	−1.7	2 Sept.	9.7	9.8	−0.1
20	2.0	5.5	−3.5	7	10.7	11.7	−1.0
25	1.8	2.0	−0.2	12	10.9	11.5	−0.6
30	7.9	9.4	−1.5	17	10.6	10.3	0.3
4 July	8.8	8.8	0.0	22	7.6	9.3	−1.7
9	4.8	5.8	−1.0	27	6.2	7.0	−0.8
14	4.1	6.5	−2.4				

Large differences are sometimes established between the measured and the computed water stages. From this it may be concluded that the calculated water stage is unsuitable for determining the terms of the water budget for a short period. Such great systematic differences as in Luutasuo do not occur here.

The water stage has been calculated also for the years 1960, 1963, 1964 and 1965. The initial water stage has been estimated as 15.0 cm on May 1. From these and the observed water stages in 1961 and 1962, the three month totals of evapotranspiration and runoff have been computed. These totals are given in Table 12.

Table 12. The sums of precipitation P, evapotranspiration E and runoff R for periods from June to August in Pohjoisneva.

Year	ΣP mm	ΣE mm	ΣR mm
1960	290	170	40
1961	320	170	120
1962	250	140	70
1963	150	150	5
1964	150	160	10
1965	230	160	60

When comparing the values in Table 12 with the corresponding values for Luutasuo in Table 10, it has to be taken into account that the catchment in the latter case consists of only peatland. In Pohjoisneva the catchment of gauges 1 and 3 covers an area of 5 km^2 of which only 2.8 km^2 is peatland.

It is possible to compare the computed runoff with the runoff measured in the streamlet flowing from Pohjoisneva. The area of the catchment of this measuring site was determined from a 1:10 000 scale topographic map. The catchment area is 10.0 km^2, and the area of the peatland in the natural state is 2.8 km^2, as stated earlier. 1.5 km^2 of the area is drained fen and the remaining part is covered by forest.

For the runoff sum of the period from June to August the following values for different years were obtained:

	1961	1962	1963	1964	1965
ΣR, mm	110	70	6	10	45

5.3 Water budget in Naarasaapa at Korvanen

Computation of function $s(W)$

Measurements in Naarasaapa at Korvanen were carried out during June, July and August in 1959 and 1960. The measuring station was situated

(Fig. 2a) at one side of the fen and owing to this situation the results obtained are not considered applicable to the fen as a whole. Moreover, a further weakness was introduced since the evapotranspiration was measured from sample containers with a water depth of only 2 cm. Thus the dependence of the real evapotranspiration on the water depth remains uncertain.

The evapotranspiration for Naarasaapa can be calculated from equation (30) by replacing the variable E_{15} by the evapotranspiration E_2. The ratio E_{15}/E_2 was calculated from measurements carried out at Loppi using the expression $E_{15}/E_2 \approx (\Sigma E_{11} + \Sigma E_{16})/2\Sigma E_2$. The constant term e_2 in equation (11) was again determined from the water budget calculation. Thus we obtain

$$(31) \qquad E(W,E_2) = (0.53 + 0.014W)E_2 \,.$$

The computations of the $R(W)$ and $s(W)$ curves for Naarasaapa have been carried out previously (VIRTA, 1962). There the evapotranspiration was computed from the equation obtained for Pohjoisneva, in which the difference of the 0-level of each gauge in relation to its respective fen surface was eliminated by adjusting the parameter e_2. This method gave large amounts of evapotranspiration and a negative value for the runoff. This fact was explained by the situation of the water gauge near the side of the bog where runoff from the forest might have affected the results. However, here the results will be computed by assuming that the runoff of the fen is zero for low water stage.

The results showed that the runoff would be zero also for a high value of the water stage. It may be that the runoff is greater than zero only during three periods with maximum water stage. For this reason only the computed curve $s(W)$ has been presented in Fig. 19. In computations of function $s(W)$ all five-day periods with the absolute value of the change of the

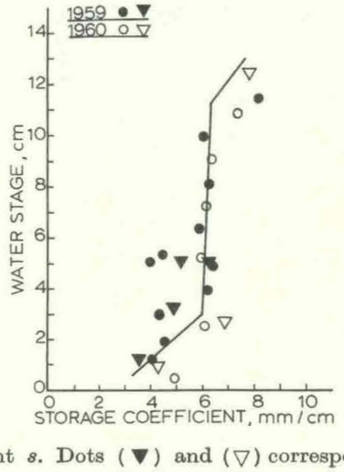

Fig. 19. The storage coefficient s. Dots (▼) and (▽) correspond to period with $W_e > W_b$.

water stage $|W_b - W_e|$ greater than 1.0 cm have been used. The three periods with runoff greater than zero were neglected.

The water gauge was situated in the »rimpi», a watery area of the fen. In 1960 the surface of the peat near the gauge corresponded to the gauge height 13 cm at the beginning of the observations and 9 cm at the end of observations.

Terms of the water budget during measurement period

It has been stated earlier that according to the calculation no runoff appears in the measurement site. For this reason only evapotranspiration, change of water content and residuals are presented in Table 13. In some cases the residual may include runoff.

It is seen that the absolute value of the residual is sometimes rather high. This occurs often in periods with precipitation during the last day, and in these cases the residuals are marked with an asterisk (*). The periods

Table 13. Precipitation P, areal evapotranspiration E, change in water content ΔS, residual Δ and mean water stage $\overline{W}$ in Naarasaapa in 1959 and 1960.

Period	P mm/5-day		E mm/5-day		ΔS mm/5-day		Δ mm/5-day		$\overline{W}$ cm	
	1959	1960	1959	1960	1959	1960	1959	1960	1959	1960
May 26–30		0.6		16.8		−18.9		2.7		13.2
31–June 4		20.9		10.0		10.2		0.7		12.4
5–9	4.7	8.1	18.8	10.4	−15.4	−6.8	1.3	4.5	13.5	12.9
10–14	9.2	12.8	19.9	11.8	−8.0	0.7	−2.7*	0.3	11.3	12.1
15–19	3.0	4.5	13.3	11.3	−11.2	−6.1	0.9	−0.7	10.0	12.2
20–24	0.3	3.1	14.1	12.0	−13.9	−7.8	0.1	−1.1	7.9	11.1
25–29	3.8	0.8	10.3	16.7	−6.9	−16.2	0.4	0.3	6.0	8.8
30–July 4	0.0	8.2	11.3	14.4	−10.9	−6.2	−0.4	0.0	4.5	7.2
5–9	21.5	0.0	10.6	18.7	12.7	−18.8	−1.8*	0.1	4.0	5.2
10–14	3.2	0.3	9.9	14.4	−9.2	−12.2	2.5	−1.9	5.4	2.6
15–19	3.6	4.8	11.6	11.5	−7.4	−2.9	−0.6	−3.8*	4.5	0.4
20–24	0.2	6.8	13.2	8.6	−13.5	−2.9	0.5	1.1	2.0	0.3
25–29	13.4	19.7	8.4	8.6	5.4	9.9	−0.4	1.2	0.5	0.0
30–Aug. 3	0.7	8.7	6.4	8.1	−5.4	−1.9	−0.3	2.5	1.6	2.4
4–8	33.5	5.8	6.7	7.1	29.6	−1.3	−2.8	0.0	3.0	1.5
9–13	1.8	9.7	9.7	7.6	−12.2	1.6	4.3	0.5*	5.3	1.1
14–18	0.1	2.6	9.9	2.6	−12.4	−0.4	2.6	0.4	2.5	1.8
19–23	4.3	16.6	7.2	4.8	0.0	9.6	−2.9	2.2	1.6	3.0
24–28	40.3	9.4	3.3	4.6	37.6	5.2	−0.6	−0.4	4.0	4.0
29–Sept. 2		3.0		4.6		−3.0		1.4		4.0
Sum	143.6	146.4	184.6	204.6	−41.1	−68.2	0.1	10.0		

June 5—9, 1959, May 26—30, 1960 and June 5—9, 1960 represent periods with a maximum water stage. It is doubtful whether there is runoff during these periods.

The following sums of precipitation P and evapotranspiration E were obtained for periods from June to August:

	ΣP mm	ΣE mm
1959	150	190
1960	140	190

The total runoff is approximately equal to 0.

Here the results of evapotranspiration are greater than the values obtained for bogs situated more to the south. This may be due, in addition to a difference in type of peatland, also to the favourable meteorological conditions for evapotranspiration in Lapland during the summers 1959 and 1960. This may be observed in the values of potential evapotranspiration computed by MUSTONEN (1964). The sum of potential evapotranspiration for the three summer months at Sodankylä in Lapland was 262 mm in 1959 and 255 mm in 1960. During the years 1961—1963 the corresponding value at Jokioinen in Southern Finland exceeded the above mentioned values only once, when it amounted to 268 mm in summer 1963.

6. COMPARISON WITH VALUES OBTAINED BY OTHERS

Comparison of evapotranspiration

NIINIVAARA (1955) has computed monthly evapotranspiration for some small basins situated in the southeastern part of Finland. The mean total evapotranspiration for June, July and August varied from 178 mm to 199 mm in different basins. The first value refers to a basin with part of the area a drained fen and the second value to a basin with part an open fen in the natural state. The values of evapotranspiration presented in Tables 10 and 12 are somewhat smaller. In this comparison attention should be paid to the fact that a raised bog, such as Luutasuo, is not represented in these basins.

HUIKARI (1959b) has given values of evapotranspiration computed for drained peatlands with different strip widths situated at Parkano, about 130 km southwest of Pohjoisneva. The evapotranspiration was greatest from an area with the maximum strip width 100 m. The results for this area are presented in the following.

Period	Evapotranspiration mm/day	Evapotranspiration mm/three-month
July 30—Sept. 12 1954	1.79	165
July 13—Sept. 6 1955	1.38	127

The mean ground-water depth was 17 cm during the first period and 43 cm during the second. These values of evapotranspiration are somewhat smaller than the values computed for Pohjoisneva. However, it has to be taken into account that the above values of evapotranspiration may have been affected by drainage.

ROMANOV (1960) has computed monthly totals of evapotranspiration and runoff in the Soviet Union for a raised bog called Lammin-Suo situated at Zelenogorski. Fig. 20 presents the evapotranspiration and runoff for a period from June to August plotted against the corresponding precipitation values. It is noticed that the evapotranspiration is greater than the values presented in Table 10. Accordingly the runoff is smaller than the corresponding values in Table 10.

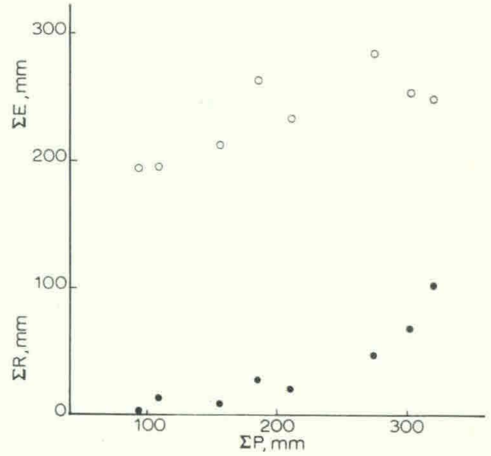

Fig. 20. Total runoff ΣR and total evapotranspiration ΣE plotted against total precipitation ΣP for periods from June to August in a raised bog in Zelenogorski, after ROMANOV (1960).

In northwestern Germany evapotranspiration has been measured from a raised bog in the natural state (EGGELSMANN, 1963). During the period from 1951 to 1957 the mean total evapotranspiration for June, July and August was 220 mm. The scattered values varied from 210 mm to 240 mm.

ODÉN (1964) has estimated the mean annual evapotranspiration from a raised bog in Southern Sweden by the chemical budget method. The result of 140 mm in a year is lower than the three month evapotranspiration presented in Table 10.

Comparison of runoff

In the following the calculated runoff will be compared with the runoff of some experimental basins of the Board of Agriculture. These basins are situated between latitudes 61°16'N and 63°26'N and between longitudes 21°31'E and 26°2'E. The areas of these basins vary from 3 km² to 79 km². The basins are without lakes and the percentage of open bog varies from 0 to 25. Runoff data has been given by MUSTONEN (1965a) and precipitation data were obtained from the year-books of the Finnish Meteorological Office. The same precipitation stations were used as by MUSTONEN (1965b) in his examination. Total runoff of these basins for periods from June to August in 1960, 1961 and 1962 has been plotted against total precipitation in Fig. 21. In this figure results of computations of Pohjoisneva are also presented, the points being marked by (C).

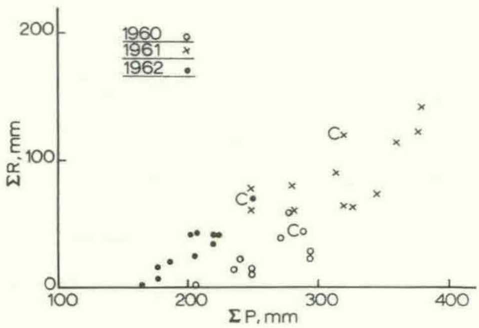

Fig. 21. Total runoff ΣR of some experimental basins of the Board of Agriculture as well as of Pohjoisneva (C) plotted against total precipitation ΣP for periods from June to August.

It is seen that the computed values for Pohjoisneva agree with the measured ones, although perhaps the computed runoff for summer 1961 is somewhat too large.

SUMMARY

The main purpose of this investigation has been to determine and clarify the evapotranspiration from different types of open peatlands. The measurements of evapotranspiration have been carried out using lysimeters. It has been attempted to adjust the measured evapotranspiration by means of the water budget calculation. The measuring sites were situated in Lapland, East Bothnia and Southern Finland the measurements in each bog being carried out during one or two summers.

A summary of the most important results will be presented:

1) On comparing the measured evapotranspiration with the evapotranspiration calculated by different approximate equations, it was found that the ratio between the total measured evapotranspiration for three months and the corresponding amount computed by the PENMAN-formula

varied by only 5 per cent from the mean value of different periods of measurements. The best correlation was obtained between measured five-day evapotranspiration and the MAKKINK variable.

2) The measurements indicate that meteorological factors and water depth have the greatest effect on the local evapotranspiration. The water depth has proved to be even more significant than the type of vegetation. The results obtained from measurements pertain, of course, to only the type of vegetation used in the sample containers. On the above mentioned basis areal evapotranspiration may be determined from the distribution of the bog-surface height in relation to the bog water level, as well as by consideration of the meteorological factors.

3) It was noticed that the least complicated model, which assumed proportionality between evapotranspiration and some meteorological factor, applies to only the potential evapotranspiration. In other cases the daily evapotranspiration may be calculated from a linear equation. However, in the water budget calculations the model based on proportionality has been used, as the influence of the constant term in the linear equation is small.

4) In computing the water budget a method was used in which the areal evapotranspiration is considered proportional to the evapotranspiration measured by the lysimeter. The proportionality coefficient and the runoff are regarded as dependent on only the water stage, and the change of water content of the bog during a period is taken as proportional to the change of water stage during the same time. The method was observed to be applicable to open peatlands, in particular to a raised bog.

5) The above method provides a possibility for adjusting the measured evapotranspiration. Here it must be assumed that the runoff is small for small water stage. The error in the areal evapotranspiration adjusted by this method may amount to 10 per cent.

6) Results of water budget calculations provide a means for constructing a nomograph for computation of the water stage. In order to utilize the nomograph, precipitation and evapotranspiration determination from meteorological factors for five-day periods have to be known. From the computed water stage it is possible to estimate the terms of the water budget. The estimation of these terms has to be carried out for periods longer than one month. Calculations for Luutasuo indicate that there may be some systematic error factor in this method.

7) The reliability of the water budget calculation is reduced by the fact that the computations are based on observations carried out during only one or two summers. The accuracy of the method may be improved by observations covering a greater number of summers including both rainy and rainless periods.

8) No clear difference in respect of evapotranspiration between the peatlands situated in Southern Finland and East Bothnia was observed. The evapotranspiration from the fen situated in Lapland was found to be greater. However, since this result is related to measurement at the side of the fen, it is possible that the mean evapotranspiration from the fen is actually smaller than reported in this paper.

DISCUSSION

The method used to adjust the measured evapotranspiration in this work is based wholly on the assumption that the flow of water in deep layers of the bog is extremely slow, although this assumption has not been explicitly justified in this investigation.

The opinion that the water exchange between the bog and its surroundings is negligible during dry periods is associated with the above assumption. However, it may be thought that there is possibly water flow from the surroundings through the bottom of the bog, for example. In this case the runoff from the bog during dry periods may even be negative, and the computed values of evapotranspiration may be too small.

Further study and clarification of any such eventual water exchange is thus left to future research projects.

REFERENCES

ANDERSON, ERNEST R., 1952: Energy-budget studies. *Water-loss investigations: Volume 1 — Lake Hefner studies technical report*, U. S. Geol. Surv. Circ. 229, 71—119.

BAC, STANISLAW, 1936: Moorforschungen in Polen. *V. Hydr. Konf. der Balt. Staaten*, Bericht 3 E, Helsinki.

BADEN, W. & EGGELSMANN, R., 1958: Über den Einfluss der Vegetation leistungsfähigen Hochmoorgrundlandes auf den Wasserhaushalt. *Publication N:o 44 of IASH*, Gen. Assemb. of Toronto, vol. II, 387—396.

BLANEY, HARRY F., 1954: Evapo-transpiration measurements in Western United States. *Publication N:o 38 de AIHS*, Assemb. Gén. de Rome, tome III, 150—160.

BLOMQVIST, EDV., 1917: Haihtumismittauksia Pyhäjärvessä Tampereen luona vuosina 1912 ja 1913. *Suomen tie- ja vesirakennusten ylihall. Hydrografisen toimiston tiedonantoja*, III, Helsinki.

BUDYKO, M. I., 1958: *The heat balance of the earth's surface* (translated from the Russian). Washington D.C..

CRAMÉR, HARALD, 1961: *Sannolikhetskalkylen och några av dess användningar*. Uppsala.

DUBACH, A. D., 1936: Moorforschungen in der UdSSR. *V. Hydr. Konf. der Balt. Staaten*, Bericht 3 G, Helsinki.

EGGELSMANN, RUDOLF, 1957: Zur Kenntnis der Zusammenhänge zwischen Bodenfeuchte und oberflächennahem Grundwasser. *Die Wasserwirtschaft*, 47:11, 283—287.

—»— 1960: Über den unterirdischen Abfluss aus Mooren. *Die Wasserwirtschaft*, 50:6, 149—154.

—»— 1963: Die potentielle und aktuelle Evaporation eines Seeklima-Hochmoores. *Publication N:o 62 of IASH*, Gen. Assemb. of Berkeley, committee for evaporation, 88—97.

Finnish Meteorological Office: Precipitation and snow cover data, 1960, 1961, 1962.

FRANSSILA, MATTI, 1940: Zur Frage des Wärme- und Feuchteaustausches über Binnenseen. *Soc. Sci. Fenn. Comm. Phys.-Math.*, X. 14, Helsinki.

—»— 1960: On the measurement of soil temperature in forests and swamps. *Finnish Meteorological Office contributions*, N:o 52, Helsinki.

—»— 1962: On the temperature conditions in a large aapa bog area in Finnish Lapland. *Finnish Meteorological Office contributions*, N:o 53, Helsinki.

HEIKURAINEN, LEO, 1963: On using ground water table fluctuations for measuring evapotranspiration. *Acta Forestalia Fennica*, 76.5, Helsinki.

—»— 1964: Ajatuksia turvemaiden vesitaloudesta. Abstract: Thoughts on the water economy of peat lands. *Suo*, 15:2, 37—43.

—»— PÄIVÄNEN, JUHANI and SARASTO, JUHANI, 1964: Ground water table and water content in peat soil. *Acta Forestalia Fennica*, 77.1, Helsinki.

HOMÉN, THEODOR, 1893: *Om nattfroster*. Stockholm.

HUIKARI, OLAVI, 1959a: Kenttämittaustuloksia turpeiden vedenläpäisevyydestä. Deutsches Referat: Feldmessungsergebnisse über die Wasserdurchlässigkeit von Torfen. *Communicationes Instituti Forestalis Fenniae*, 51.1, Helsinki.

—»— 1959b: Metsäojitettujen turvemaiden vesitaloudesta. Deutsches Referat: Über den Wasserhaushalt waldentwässerter Torfböden. *Communicationes Instituti Forestalis Fenniae*, 51.2, Helsinki.

Ilmatieteellinen keskuslaitos: Kuukausikatsaus Suomen sääoloihin, 53:6,7,8, 54:6,7,8, 55:6,7,8, 56:6,7,8, 57:6,7,8, 58:6,7,8, 59:6,7,8.

—»— Kuukausikatsaus Suomen ilmastoon, 1966, kesäkuu, heinäkuu, elokuu.

IVANOV, K. E., 1957 — ИВАНОВ, К. Е., 1957: Основы гидрологии болот лесной зоны. Ленинград.

JUUSELA, T., 1962: Ein zur unmittelbaren Messungen der Verdunstung der Bodenoberfläche benutztes Verfahren. *Zeitschrift für Kulturtechnik*, 3:3, 137—142.

JÄRVINEN, JUHANI, 1962: Pohjavesipinnasta kosteussadannesten ja pohjavesikaivojen valossa. *Suo*, 13:2, 25—27.

KAITERA, PENTTI, 1939: The evaporation from snow. *Publication de AIHS*, Réunion de Washington, tome II, comission des neiges, Question 2, Rapport 2.

KING, K. M., TANNER, C. B. and SUOMI, V. E., 1956: A floating lysimeter and its evaporation recorder. *Trans. Amer. Geophys. Union*, 37:6, 738—742.

KIVINEN, ERKKI, 1948: *Suotiede*. Porvoo.

MAKKINK, G. F., 1957: Ekzameno de la formulo de Penman. *Neth. Journ. Agr. Sci.*, 5:4, 290—305.

MATHER, JOHN R., 1950: Micrometeorology of the surface layer of the atmosphere. *The Johns Hopkins University, Laboratory of Climatology*, Supplement to Interim Report N:o 10, Seabrook.

MATTSSON, R. and RAPELI, P., 1960: The total radiation at Sodankylä observatory 1953—1957. *Mitteilungen der Meteorologischen Zentralanstalt*, N:o 50, Helsinki.

MUSTONEN, SEPPO E., 1964: Potentiaalisen evapotranspiraation määrittämisestä. English summary: Estimating potential evapotranspiration. *Acta Agralia Fennica*, 102.2, Helsinki.

—»— 1965a: Maataloushallituksen hydrologiset tutkimukset vuosina 1957...1964. Hydrologic investigations by the Board of Agriculture during the years 1957 to 1964. *Soil and hydrotechnical investigations*, 11, Helsinki.

—»— 1965b: Meteorologisten ja aluetekijöiden vaikutuksesta valuntaan. English abstract: Effects of meteorologic and basin characteristics on runoff. *Soil and hydrotechnical investigations*, 12, Helsinki.

—»— ja LAIKARI, HANNU, 1961: Ojituksen vaikutuksesta valuntaan Huhtisuon havaintoalueella. *Maatal. hallituksen insinööriosasto, maa- ja vesitekn. tutkimustoimisto*, tiedotus 2, Helsinki.

NIINIVAARA, K., 1953: Haihtumisesta pienehköillä vesistöalueilla Suomessa. Summary: Evaporation from watersheds in Finland. *Soil and hydrotechnical investigations*, 7, Helsinki.

—»— 1955: Haihtumismääristä eri kuukausina. Summary: On the amount of evapotranspiration in different months. *Maa- ja vesirakentaja*, 2, 77—82, Helsinki.

NOMALS, P., 1938: Die Verdunstung in einem Hochmoor. *VI. Balt. Hydr. Konf.*, Bericht 16 B, Berlin.

NOVIKOV, S. M., 1964: Computation of the water-level regime of undrained upland swamps from meteorological data. *Soviet Hydrology: Selected papers*, 1, 1—22.

ODÉN, SVANTE, 1964: II. C-14 och Tritium isotopernas förekomst över Skandinavien under senare år med tillämpning inom markläras och hydrologi. Summary: The occurrence of C^{14} and Tritium over Scandinavia with applications in Soil Science and Hydrology. *Grundförbättring*, 2, 122—142.

OVERBECK, FRITZ und HAPPACH, HILKA, 1957: Über das Wachstum und den Wasserhaushalt einiger Hochmoorsphagnen. *Flora*, 144:3, 335—402.

PAAVILAINEN, EERO, 1963: Turpeen vesipitoisuudesta ja pohjavesipinnasta. Summary: On water content of peat and ground water level. *Suo*, 14:1, 8—9.

PALMÉN, E., 1963: Computation of the evaporation over the Baltic Sea from the flux of water vapor in the atmosphere. *Publication N:o 62 of IASH*, Gen. Assemb. of Berkeley, committee for evaporation, 244—252.

PENMAN, H. L., 1954: Evaporation over parts of Europe. *Publication N:o 38 de AIHS*, Assemb. Gén. de Rome, tome III, 168—176.

—»— 1956: Estimating evaporation. *Trans. Amer. Geophys. Union*, 37:1, 43—50.

PESSI, YRJÖ, 1956: Studies on the effect of admixture of mineral soil upon the thermal conditions of cultivated peat land. *Publications of the Finnish State Agricultural Research Board* N:o 147, Helsinki.

—»— 1957: On the thermal conditions in mineral and peat soil at Pelsonsuo in 1955—1956. *Publications of the Finnish State Agricultural Research Board* N:o 159, Helsinki.

PORKKA, M. T., 1956: Results of measurements with Renqvist's evaporation recorder in South Finland in summer 1950. *Geophysica*, 5:2, 70—77, Helsinki.

PRYTZ, K., 1932: Der Kreislauf des Wassers auf unberührtem Hochmoor. *Ingeniørvidenskablige Skr.*, A 33, København.

PÄIVÄNEN, JUHANI, 1964: Menetelmä pohjavesikertoimen ja pintakasvillisuuden haihdunnan määrittämiseksi. Summary: A method to determine the ground water coefficient and the ground vegetation transpiration. *Suo*, 15:6, 88—91.

REMSON, IRWIN and RANDOLPH, J. R., 1958: Application of statistical methods to the analysis of ground-water levels. *Trans. Amer. Geophys. Union*, 39:1, 76—83.

RENQVIST, HENRIK, 1951: Evaporation recorder. *Publication N:o 34 de AIHS*, Assemb. Gén. de Bruxelles, tome III, 473.

ROMANOV, V. V., 1957 — Романов, В. В., 1957: Испарение с неосушенных и осушенных болот. Труды ГГИ, выпуск 60, 20—42.

—»— 1960: Изменение водного баланса болот в засушливые и влажные годы. Труды ГГИ, выпуск 89, 5—36.

SARASTO, JUHANI, 1961: Kokeita turpeen vedenläpäisevyydestä. Experiments on the permeability of peat. *Suo*, 13:2, 24—25.

—»— 1963: Tutkimuksia rahka- ja saraturpeiden vedenläpäisevyydestä. Summary: A study on the permeability to water of different kinds of peat. *Suo*, 14:3, 32—35.

SIMOJOKI, HEIKKI, 1948: On the evaporation from the Northern Baltic. *Geophysica*, 3, 123—126, Helsinki.

SIRÉN, ALLAN, 1936: Niederschlag, Abfluss und Verdunstung des Päijänne-Gebietes. *V. Hydr. Konf. der Balt. Staaten*, Bericht 1 F, Helsinki.

SIRÉN ALLAN, 1948: Die Bestimmung der Verdunstung und ihrer Einwirkung auf die Wasserläufen von Finnland. *Geophysica*, 3, 162–172, Helsinki.

SIRÉN, GUSTAV, 1955: The development of spruce forest on raw humus sites in Northern Finland and its ecology. *Acta Forestalia Fennica*, 62.4, Helsinki.

THORNTHWAITE, C. W., 1948: An approach toward a rational classification of climate. *Geogr. Rev.*, 38, 55–94.

TODD, DAVID KEITH, 1959: *Ground water hydrology*. New York.

TURC, LUCIEN, 1954: Calcul du bilan de l'eau. Evaluation en fonction des precipitations et des temperatures. *Publication N:o 38 de AIHS*, Assemb. Gén. de Rome, tome III, 188–202.

UHLIG, SIEGFRIED, 1956: Berechnung monatlicher Mittelwerte der Gebietsverdunstung nach einer Methode von Kalweit. *Mitteilungen des Deutschen Wetterdienstes*, 3:15, Bad Kissingen.

VIRTA, J., 1960: Evapotranspiration measurements in a string fen in Northern Finland. *Publication N:o 53 of IASH*, Gen. Assemb. of Helsinki, committee of evaporation and evapo-transpiration, 438–441.

—»— 1962: Suohydrologisista tutkimuksista Lapissa ja Pohjanmaalla. Summary: On the research of peat land hydrology in Lapland and Ostrobothnia. *Suo*, 13:3, 30–35.

VISSER, W. C. and BLOEMEN, G. W., 1959: The moisture flow technique for determining the water-balance. *Publication N:o 48 of IASH*, Symp. of Hannoversch-Münden, vol. I, 128–139.

VOROB'EV, P. K., 1963: Investigations of water yield of low lying swamps of Western Siberia. *Soviet Hydrology: Selected papers*, 3, 226–252.

VÄISÄNEN, AIMO, 1962: A computation of the evaporation over Finland during a rainless period based on the divergence of the watervapor flux. *Geophysica*, 8:2, 159–165, Helsinki.

WÄRE, MATTI, 1947: Maan vesisuhteista ja viljelyskasvien sadoista Maasojan vesitaloudellisella koekentällä vuosina 1939–1944. Referat: Über die Wasserverhältnisse des Bodens und die Erträge von Kulturpflanzen auf dem wasserwirtschaftlichen Versuchsfeld Maasoja in den Jahren 1939–1944. *Soil and hydrotechnical investigations*, 5, Helsinki.

ÅNGSTRÖM, A., 1928: Recording solar radiation. *Meddelanden från Statens Meteorologisk-Hydrologiska anstalt*, 4:3, Stockholm.

Communicated May 16, 1966 by H. Simojoki
Printed October, 1966

Keskuskirjapaino—Centraltryckeriet
Helsinki—Helsingfors