THE UNIVERSITY OF MANITOBA

MELTWATER MIXING
IN SMALL ARCTIC LAKES

by

MARTIN ALEXANDER ERNEST BERGMANN

A DISSERTATION

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ABSTRACT

Meltwater mixing in small arctic lakes at Saqvaqjuac, $63^{o}68'$N was studied in 1980 and 1981 to evaluate the applicability of theoretical lake water renewal times, which assume 100% mixing of all inputs with the lake water, to ice-covered lakes. Two 370 GBq tritium additions were made to 7.09 ha P & N Lake. One was mixed with the unfrozen water at the time of maximum lake-ice thickness (May 1980) and the other was mixed with the lake immediately after freezing (October 1980). Dye experiments were also performed at four lakes to define the spatial and temporal distribution of the inflow and icemelt layers. Results from the tritiated water, conductance, temperature and dye profiles indicated that, during ice-on, the cold, low-density meltwater floated in a thin layer 0 to 100 cm beneath the ice, extended over the entire subice-surface area, and left the lake without mixing with the heavier subice water. These results imply that 1) lake models incorporating a lake flushing rate term need to be re-evaluated to accomodate the lack of meltwater mixing beneath spring ice, and 2) more attention should be given to the early spring meltwater chemistry and its distribution within the upper lake strata.

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INTRODUCTION

Lake water theoretical turnover time (TT) is an important term in lake management models (Vollenweider 1969, Dillon 1975, Dillon and Rigler 1974, Schindler 1978, Schindler *et al.* 1978, Fee 1979), and assumes that all inputs mix completely with the lake.

$$(1) \qquad TT = \frac{\text{lake volume } (m^3)}{\text{annual outflow volume } (m^3\ yr^{-1})}$$

However, results from a number of studies indicate that complete mixing probably does not occur in ice-covered lakes, but rather that inflows to ice-covered lakes can proceed directly to the outflow, mixing only with the upper layers of the lake (Groterud 1972, Schindler *et al.* 1974, Henrikson and Wright 1977, Hultberg 1977, Prentki *et al.* 1979, Hendrey *et al.* 1980). The purpose of this study was to make direct measurements of spring inflow and retention of icemelt by ice-covered lakes and thereby refine estimates of lake turnover time. More accurate estimates of lake turnover time would be useful in limnological problems concerned with acid and nutrient loading as well as metal accumulation.

Arctic lakes provide a good opportunity for such a study because wind energy transferred to the lake water is negligible during the long period of ice-cover.

In conjunction with whole-lake nutrient addition experiments in progress at Saqvaqjuac, a Canadian Department of Fisheries and Oceans research camp on Hudson Bay at $63^{o}68'N$ $90^{o}40'W$, a series of experiments was performed and measurments made to determine the fate of spring (snow and ice) meltwater beneath the ice. In P & N lake tritiated water (HTO) was used as a tracer in a two-stage experiment, with the subice labelled in 1980 and both ice and subice labelled in 1981. In separate dye experiments Rhodamine B was added to the melting ice surface (Turkey Lake) and to the inflows (P & N, Spring and Hawk Lakes) to determine the spatial pattern of stratification.

To better describe the mixing during springmelt, the lake was assumed to consist of three compartments (c): ice, subice (liquid water at the onset of melt) and in-outflow (all inputs to the lake which were discharged via the outflow).

The traditional theoretical turnover time model (Formula 1) is based on the lake hydrological budget (H_2O dataset). This model predicts a whole-lake turnover time where each of the above compartments would have the same turnover time. In this study two whole-lake mixing models were derived, and used to compare the traditional model with the tritium data. The tritium data were developed by integrating the lake morphometry with the lake water inputs and outputs to form the HTO dataset. The first model (M1) determined a theoretical turnover time (TT_c) for each of the three compartments, whereas the second model (M2) considered daily changes in the lake system and predicted outflow losses for the labelled watermass during the the ice-on period.

To provide an independent comparison of the traditional model and the observed data conductivity data were also used in conjunction with the water budget data to predict daily outflow concentrations and were compared with the measured data.

MATERIALS AND METHODS

Description of the study area.

Saqvaqjuac is situated 45 km north of Chesterfield Inlet N.W.T., on the west coast of Hudson Bay (Fig. 1). The four study lakes (P & N, Hawk, Spring and Turkey) occupy depressions in bedrock composed of Late Precambrian diorite, gneiss and granite. Vegetation is typical of the arctic tundra. Local relief is 20-50 m.

Saqvaqjuac lakes freeze about 1 October and ice depth averages 1.7-1.9 m by May. Slush ice, formed on the upper ice surface as a result of snow cover loading (Jones, 1970), does not occur in all years and is usually <10 cm thick. Annual precipitation (mean for 1977-1981) is 298 mm (Fig. 2). Spring melt begins by late May or early June when ambient temperatures are near $0^{o}C$, although freezing of inflows and outflows sometimes occurs during this period and may last several days. With increasing air and water temperature a moat forms and extends to a maximum depth of approximately 1.5 to 2.0 m before the ice-pan begins its rapid deterioration. P & N Lake temperature isopleths for 1980 and 1981 are shown in Fig. 3.

P & N lake (Fig. 4), the site of the whole-lake HTO experiments, is a headwater lake with a mean annual (1977-1980) theoretical turnover time (TT) of 2.2 years. Spring, Hawk and Turkey lakes (Fig. 5-7), used in dye experiments, have TT's of 1.1, 2.9 and 0.4 years respectively.

Figure 1: Saqvaqjuac research field camp and experimental lakes

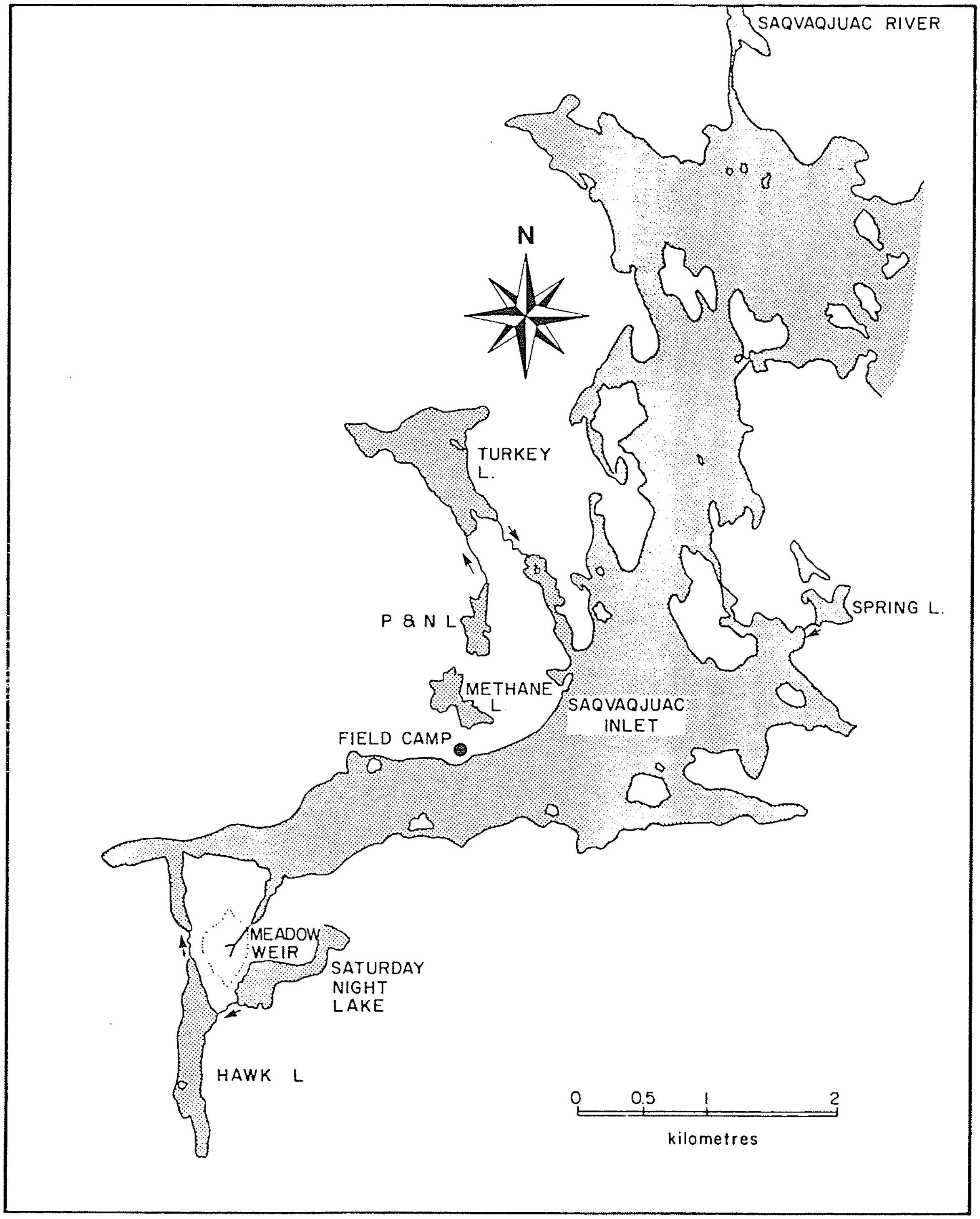

SAQVAQJUAC RIVER
N
TURKEY
L.
SPRING L.
P & N L
METHANE
L.
SAQVAQJUAC
INLET
FIELD CAMP
MEADOW
WEIR
SATURDAY
NIGHT
LAKE
HAWK L
0 0.5 1 2
kilometres

Figure 2: Mean annual precipitation at Saqvaqjuac/Chesterfield Inlet for 1977-1981. May - September from Saqvaqjuac data, else from Environment Canada

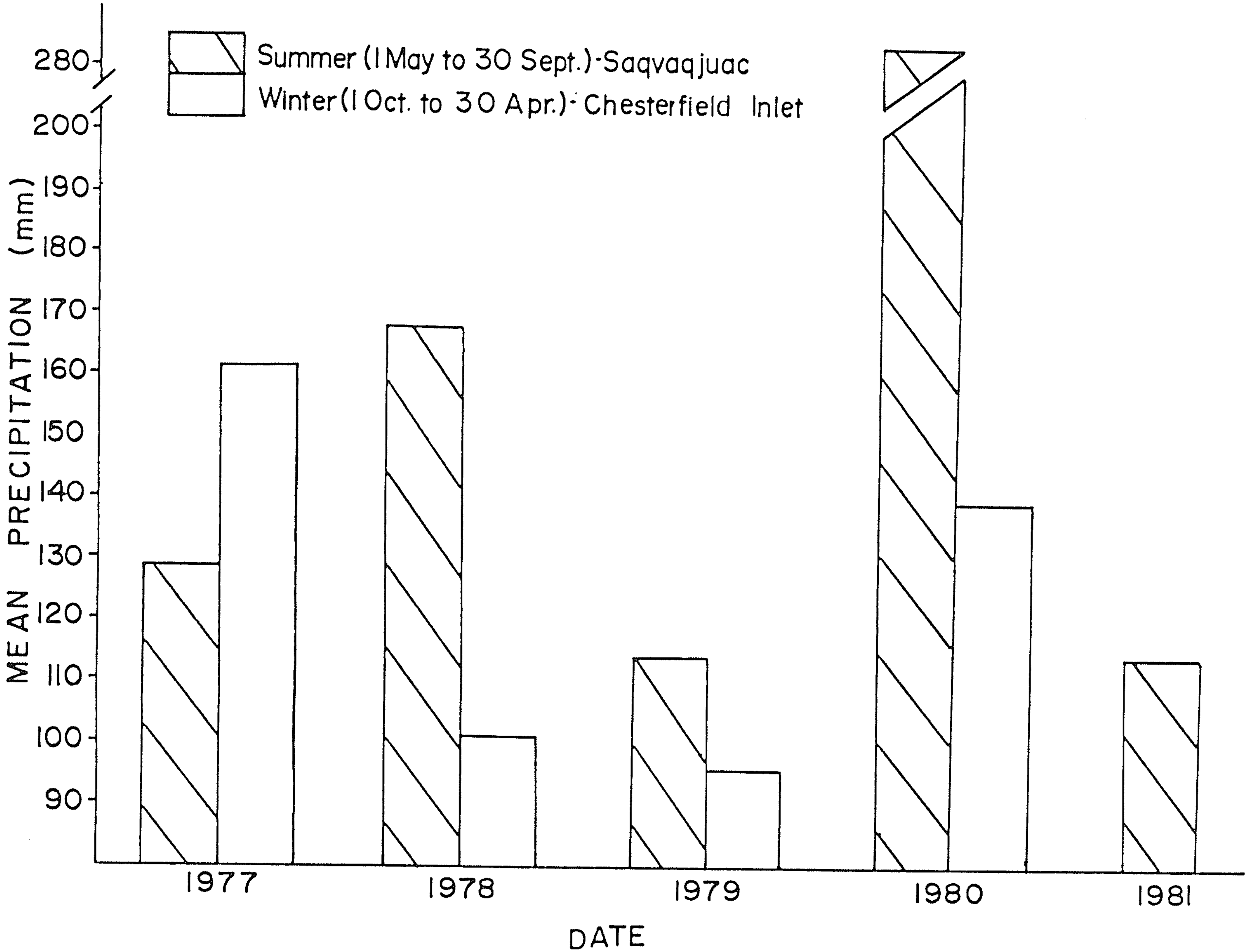

Summer (1 May to 30 Sept.)-Saqvaqjuac
Winter (1 Oct. to 30 Apr.)-Chesterfield Inlet
MEAN PRECIPITATION (mm)
280
200
190
180
170
160
150
140
130
120
110
100
90
1977
1978
1979
1980
1981
DATE

Figure 3: P & N Lake temperature isopleths for 1980 and 1981

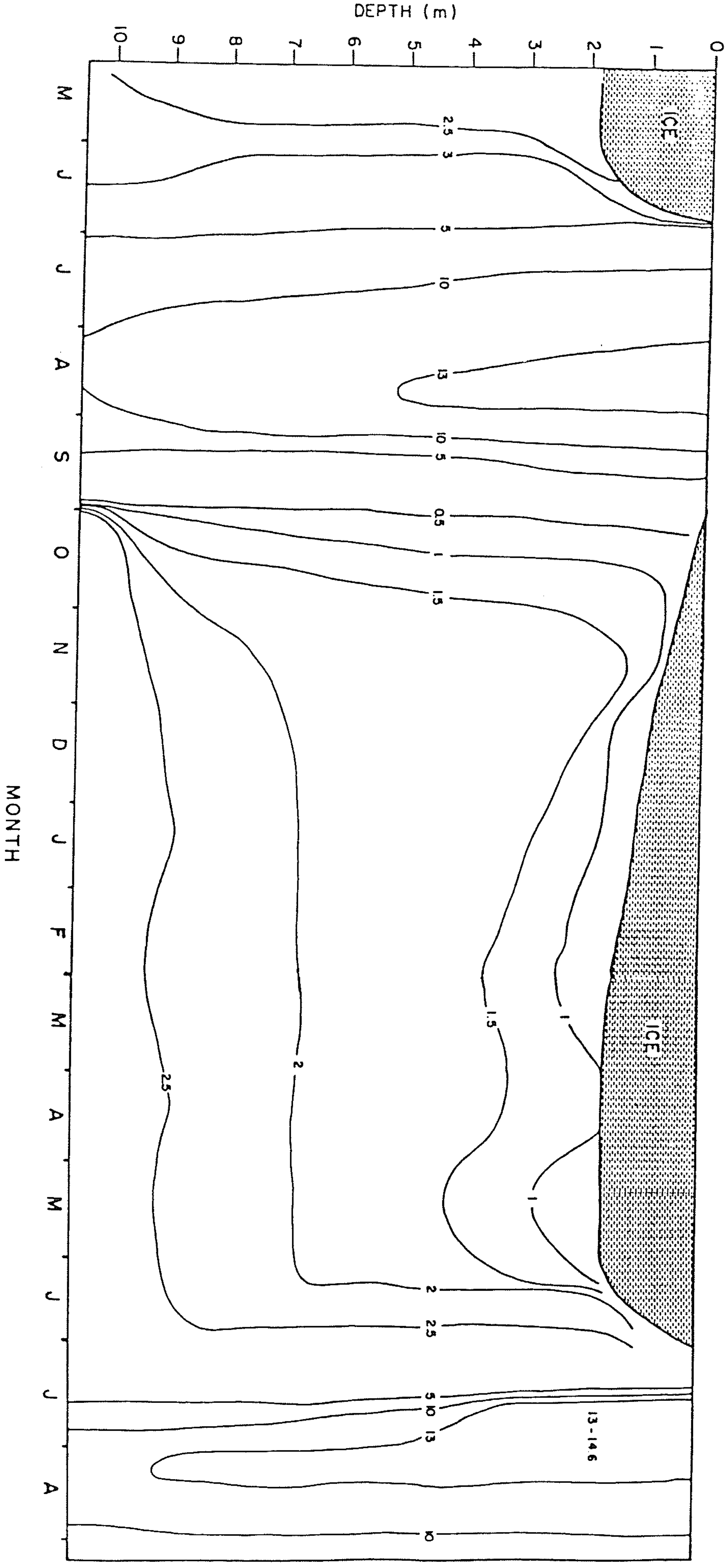
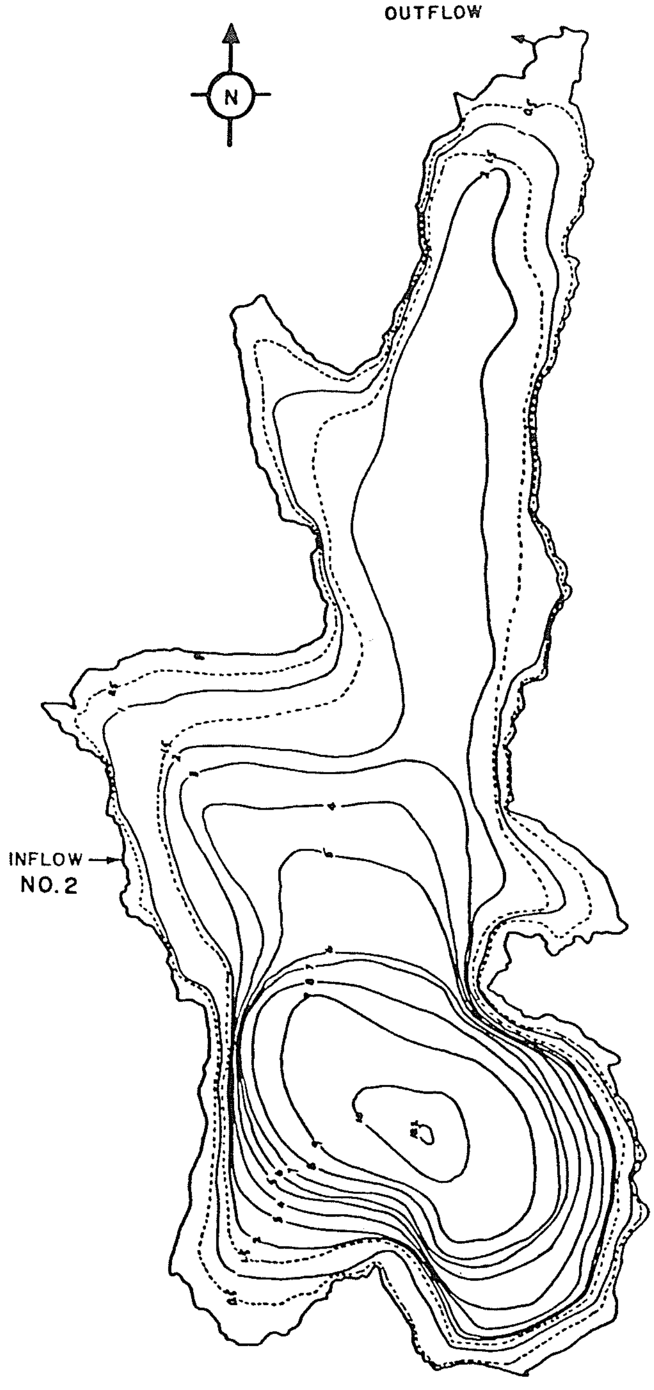

DEPTH (m)
0
1
2
3
4
5
6
7
8
9
10
ICE
ICE
MONTH
M
J
J
A
S
O
N
D
J
F
M
A
M
J
J
A
2.5
3
5
10
13
10
5
0.5
1
1.5
1
1.5
2
2.5
1
2
2.5
5
10
13
13-14.6
10

Figure 4: P & N Lake, morphometric map showing dye addition site at inflow no.2



P & N LAKE

CONTOUR INTERVAL
ONE METER

SCALE METERS
0 50 100

DEPTH (m)	AREA ($10^4 m^2$)	VOLUME ($10^4 m^3$)
0	7.09	3.34
0.5	6.26	2.97
1.0	5.62	2.59
1.5	4.76	2.13
2.0	3.79	3.16
3.0	2.57	2.35
4.0	2.13	1.91
5.0	1.70	1.48
6.0	1.27	1.19
7.0	1.12	1.02
8.0	0.93	0.78
9.0	0.63	0.34
10.0	0.12	0.01
10.2	0	

TOTAL VOLUME = 232,700 m^3

MEAN DEPTH = 3.3 m

EXTREME LAKE LEVELS
1977-1979
15.73 , 16.18 meters A.S.L.

Figure 5: Spring Lake, morphometric map showing dye addition site at inflow

SPRING LAKE

0 20 40 60 80 100 200
SCALE IN METERS

CONTOUR INTERVAL
ONE METER

DEPTH (m)	AREA ($10^4 m^2$)	VOLUME ($10^4 m^3$)
0	6.94	6.07
1	5.25	4.31
2	3.44	2.89
3	2.37	2.05
4	1.74	1.53
5	1.33	1.13
6	0.95	0.60
7	0.32	0.11

TOTAL VOLUME 186,900 m^3
MEAN DEPTH 2.69 m

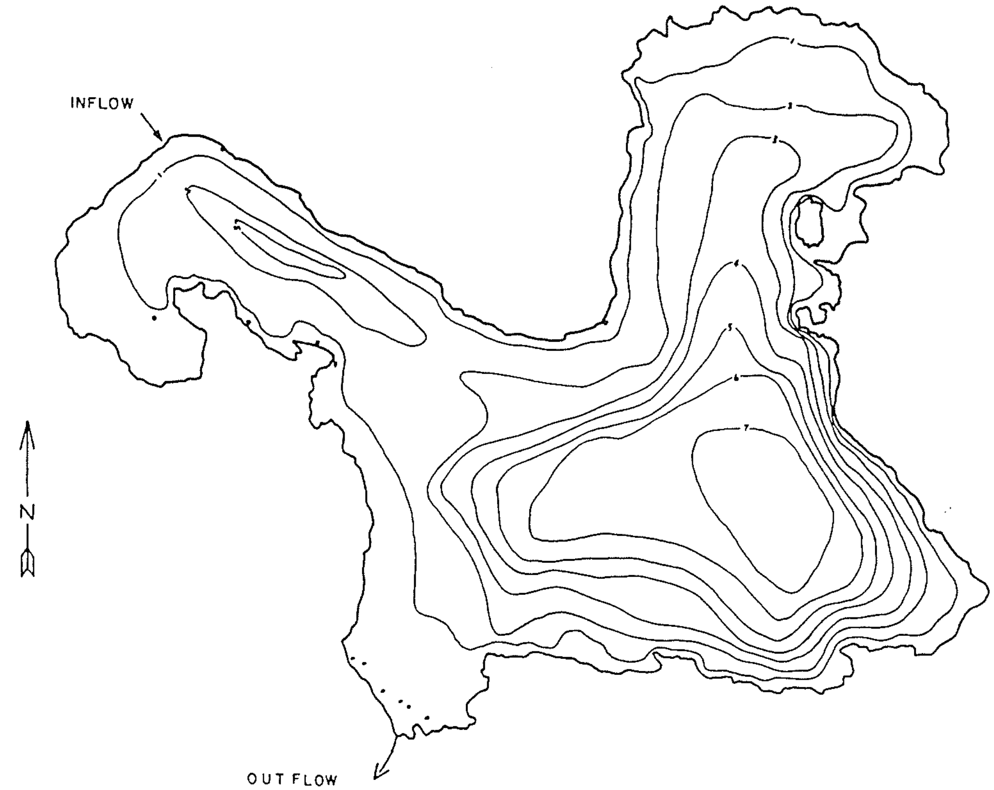


Figure 6: Hawk Lake, morphometric map showing dye addition site at inflow

HAWK LAKE

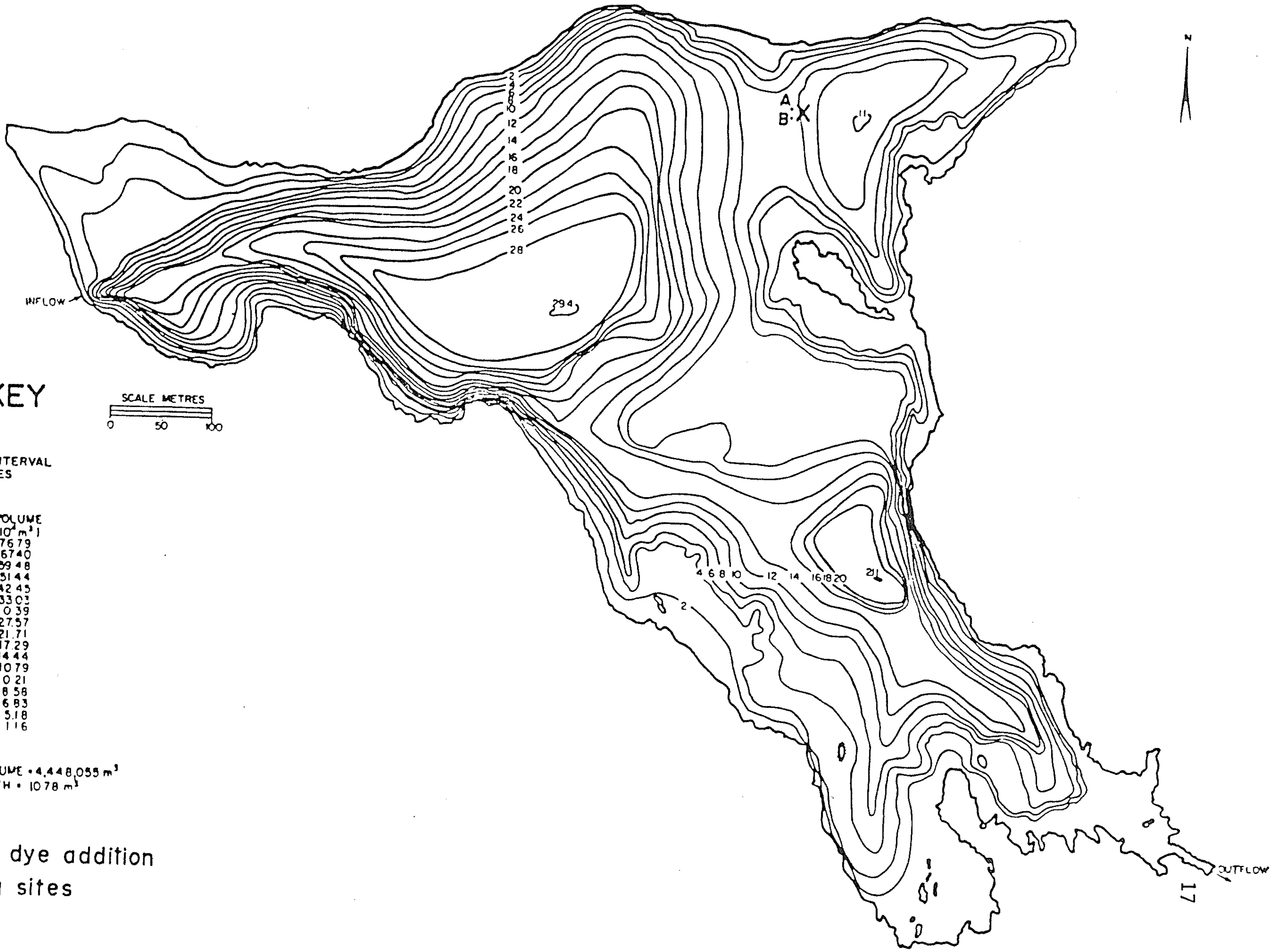
CONTOUR INTERVAL
TWO METRES

SCALE METRES
0 50 100

DEPTH (m)	AREA ($10^4 m^2$)	VOLUME ($10^4 m^3$)
0	24.34	45.28
2	20.98	39.55
4	18.59	34.77
6	16.21	29.83
7.4	0.004	0.11
8	13.85	25.84
10	12.01	22.42
12	10.43	19.37
14	8.96	16.13
16	7.20	12.18
18	5.04	8.31
20	3.33	5.22
20.4	0.02	.09
22	2.19	4.01
24	1.83	3.20
26	1.38	2.53
28	1.15	2.07
30	0.93	1.51
32	0.60	.67
34	0.13	.01
34.2	0.03	

TOTAL VOLUME 2,834,370 m^3
MEAN DEPTH 11.65 m

Figure 7: Turkey Lake, morphometric map showing site of surface dye addition



SCALE METRES
0 50 100

TURKEY LAKE

CONTOUR INTERVAL
TWO METRES

DEPTH (m)	AREA ($10^4 m^2$)	VOLUME ($10^4 m^3$)
0 0	41 25	76 79
2 0	35 61	67 40
4 0	31 82	59 48
6 0	27 71	51 44
8 0	23 78	42 45
10 0	18 77	33 03
11 0	0 02	0 39
12 0	15 23	27.57
14 0	12 39	21.71
16 0	9 39	17.29
18 0	7 93	14 44
20 0	6 54	10 79
21 1	0 01	0 21
22 0	4 76	8 58
24 0	3 84	6 83
26 0	3 05	5.18
28 0	2 15	1 16
29 4	0 04	

TOTAL VOLUME = 4,448,055 m^3
MEAN DEPTH = 10 78 m^3

X Surface dye addition
A,B Sampling sites

Complete water flow data for P & N lake for 1980 and 1981 are provided by Dalton (1981) and Welch (pers. comm.).

Whole-lake HTO addition experiments.

The first addition of HTO to P & N lake was on 16 May 1980, at the point of maximum water depth (Z_{MAX}). A 370 GBq (10 Ci) vial (New England Nuclear) was added 0.3 m below the bottom of the ice, using the apparatus shown in Fig. 8. At 1950h the HTO vial was crushed in the stream of the water pump (flow directed horizontally). The pump was turned off at 2330h, restarted (flow aimed downward) at 0840h and continued intermittently from 17 - 19 May. Air was then pumped at 0.5 $l\ min^{-1}$ through airstones at 8 m depth for 12 hours over the next two days, after which profiles showed that the HTO was completely mixed throughout the lake. Thermal stratification was re-established within 3 days.

On 3 October 1980, another 370 GBq of HTO was added to P & N lake under 13 cm ice. The vial was crushed, diluted in 200 l of lake water and siphoned into the lake beneath the ice at 10 $ml\ s^{-1}$. The airstones were then lowered as before and air pumped for 20 hours over six days. An outboard motor, the boat anchored to the ice-hole edge, was also run intermittently to achieve complete mixing of HTO with the lake.

Sampling and analytical methods.

Duplicate outflow samples from P & N lake were taken at least daily over the entire flow periods of 1980 and 1981. To determine the stratification of meltwater beneath the ice, a suction sampler (Fig. 9) which drew water (25 $ml\ s^{-1}$) from a layer about 1 cm thick was used to sample water through grids of holes. A 2 l Van Dorn water bottle was used to sample the lake during ice-off period.

Figure 8: Apparatus used for first HTO addition (P & N Lake 16 May 1980)

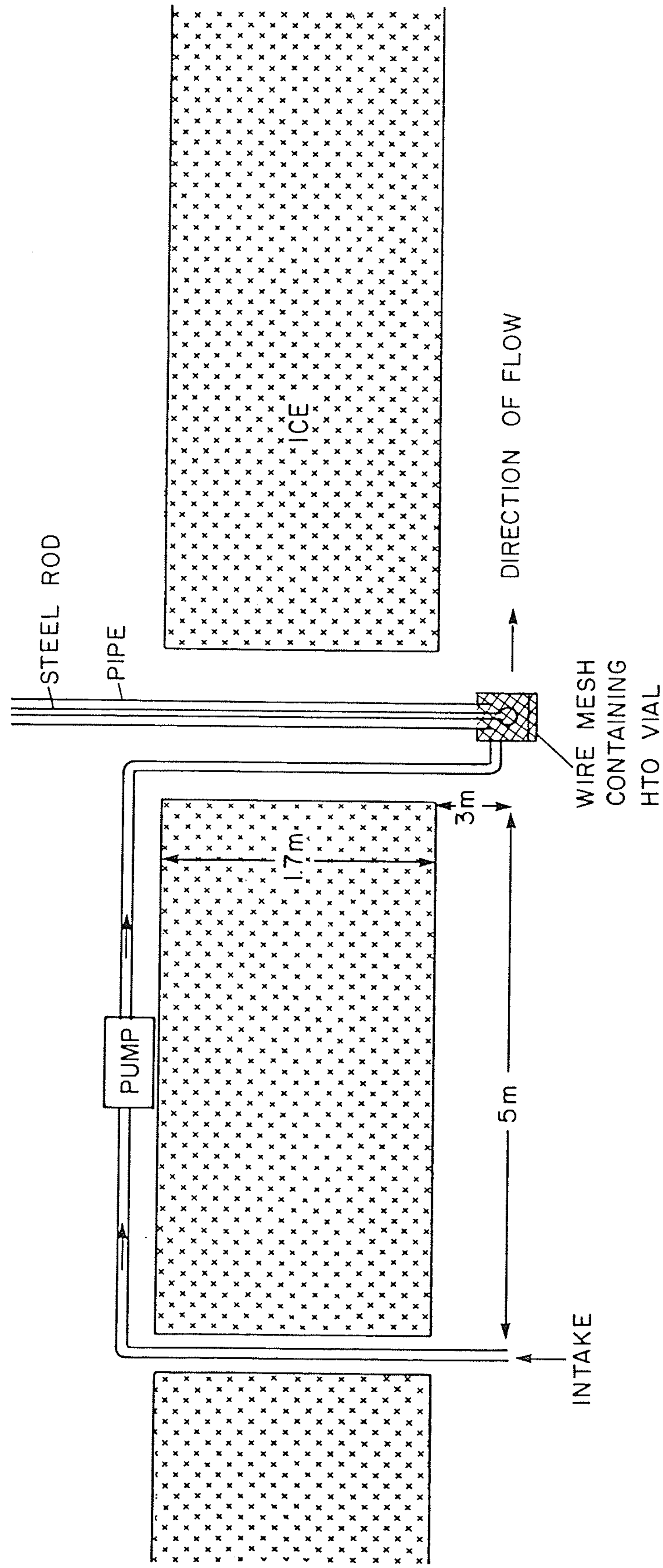
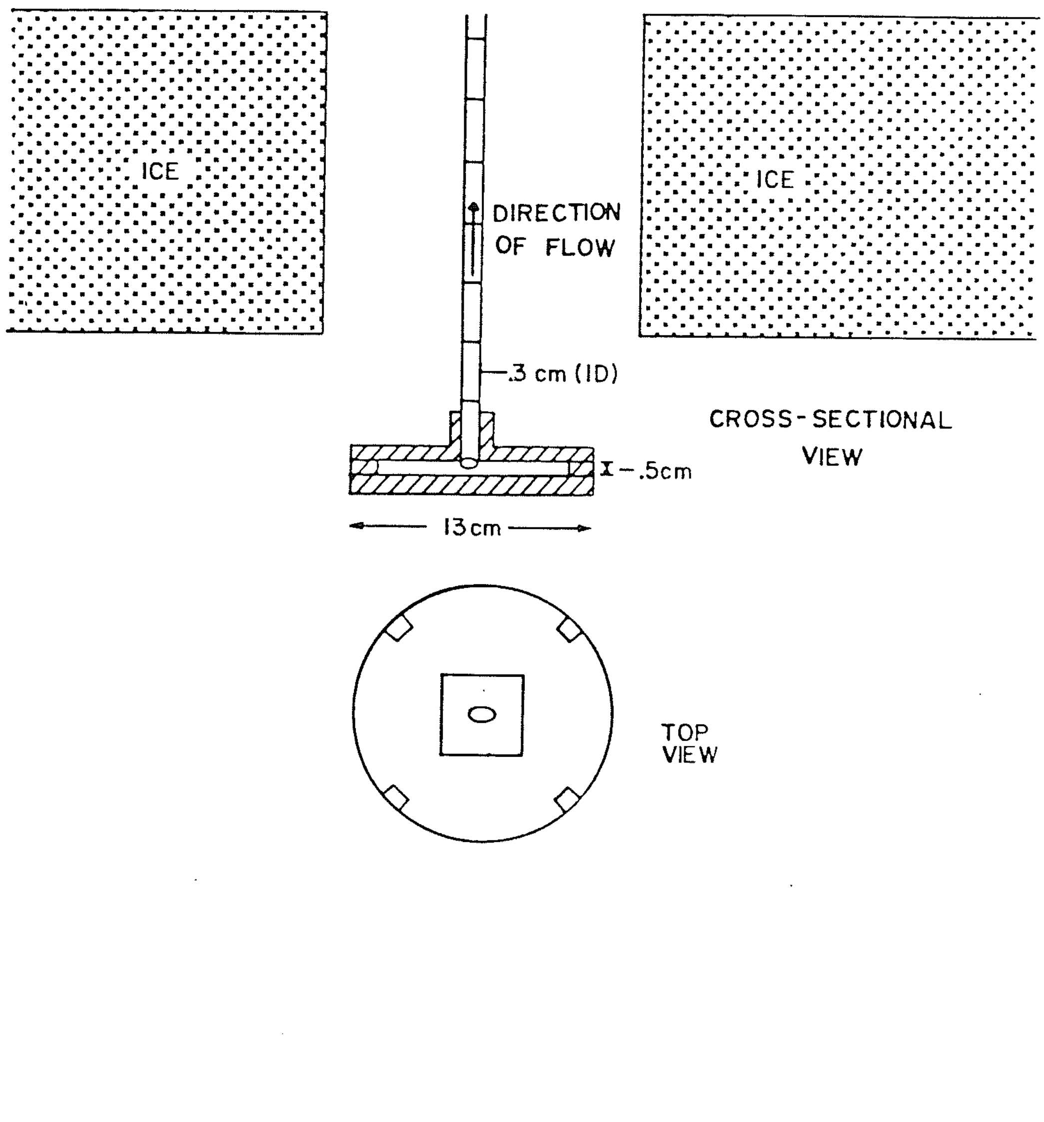

STEEL ROD
PIPE
ICE
DIRECTION OF FLOW
WIRE MESH
CONTAINING
HTO VIAL
3m
1.7m
5m
PUMP
INTAKE

Figure 9: Suction sampler, used for sampling water at small depth intervals ($\geq$ 1 cm)


ICE
DIRECTION
OF FLOW
ICE
.3 cm (ID)
CROSS-SECTIONAL
VIEW
.5cm
13cm
TOP
VIEW

Conductance (μmhos cm^{-1} at $25^{o}C$) was measured using a Radiometer conductivity meter calibrated against a standard 0.01 KCl solution, with a precision of 5% on samples from P & N lake, inflows and outflows. Each subsample was transferred to a liquid scintillation vial containing a 10 ml aliquot of fluor (Beckman All-Purpose scintillation cocktail). A liquid scintillation counter (Beckman LS100) with an efficiency for tritium of 23 $\pm$ 3% (glass vials) and 21 $\pm$ 3% (plastic vials) was used to determine the activity of each sample (counting efficiency did not change throughout the study period). The samples were counted to 5% (2 σ) statistical counting error. Including sample volume error, the precision of each HTO analysis was probably about $\pm$ 8%.

In addition to routine monitoring of ice thickness and moat width (the distance between lake edge and ice-pan) for 1980 and 1981, in May 1981 marked rods (3 mm X 90 cm) mounted on a 9 cm^2 clear plastic base were placed in 10 cm diameter holes with the base resting on the ice in the bottom of the hole 40 to 80 cm deep. The hole was filled with water and left to freeze, leaving the rod extending above the ice surface. After the beginning of melt, the increase in rod length above ice surface was used to measure the surface icemelt.

Lake temperature (T) measurements were made regularly from May 1980 to August 1981 using two different systems, a Hydrolab model RT-125 ($\pm$ $0.1^{o}C$) thermistor and a bead thermistor ($\pm$ $0.03^{o}C$).

Dye methods.

On 27 May and 4 June 1980 and 12 June 1981, 20 l of Rhodamine B (dye) solution was added over 0.5 hours to P & N inflow no.2 (Fig. 4), Spring (Fig. 5) and Hawk (Fig. 6) lake inflows respectively to determine the inflow path. Rhodamine B was also sprayed over 100 m^2 of Turkey lake ice (Fig. 7) on 14 June 1980 to follow icemelt as it entered the lake. Rhodamine B fluorescence was measured in a Turner model 111 fluorometer.

Pore-water methods.

The loss of HTO to the sediments was measured using pore-water peepers (Hesslein, 1976a) filled with low HTO water (<10 dpm) and covered with .21 μm (Nucleopore[R]) dialysis membrane. Each acrylic pore peeper was 70 cm long and consisted of 36, 1.2 cm by 7.6 cm chambers arranged serially on a plane, each separated by a 0.7 cm wall. The pore peepers were placed by diver on three occasions and left to equilibrate for two weeks, after which they were pulled by attached float-lines to the surface (elapsed time 10 s) where 8.1 ml of water from each 11 ml chamber was withdrawn by syringe and counted for HTO.

The HTO flux to the sediments was determined in two steps. First, a best-fit time-dependent numeric model was used to describe the sediment HTO profiles (Hesslein, 1980b); and for each depth, the concentration was given by:

$$C^{t+\Delta t} = C^{t} + \frac{\left(C^{t}_{z+\Delta z} - 2C^{t}_{z} + C^{t}_{z-\Delta z} \right)}{\Delta z} \quad (tD)$$

where: C^t = HTO concentation at depth z (cm) at model starting time

t = 1 hour

D = diffusion coefficient of HTO into sediments

Second, the model-generated curve was then integrated over the profile depth and applied to the area >1 m depth to give a flux for the entire lake.

Gross evaporation.

The difference between the HTO predicted for the whole lake from the water budget and that actually measured (both corrected for sediment loss) during the ice-off period was assumed to be due to HTO loss from the lake surface via vapor exchange (gross evaporation minus net evaporation). These losses were used together with the other measured losses to correct the HTO mass balance.

Mass balance for HTO.

Mass (M) of HTO is determined by:

$$M = \int_{s=0}^{s=max} A_s C_s d_s + \int_{i=0}^{i=max} A_i C_i d_i$$

where A is the area of the midpoint of each contour interval and C is the concentration at depth for the subice (s) and ice (i) mass.

Models.

In P & N Lake the volume of labelled subice water lost by the outflow during 1980 was determined by dividing the outflow HTO mass by the corrected subice HTO concentration. In 1981, when both ice and subice were labelled, the following equation was used:

$$V = T_O \Bigg/ \frac{T_S + T_I}{V_S + V_I}$$

where: V = volume of subice and ice lost via outflow (m^3); labelled water

T_O = mass of HTO lost via outflow (dpm); measured

T_S = subice mass of HTO (dpm); at time of maximum ice thickness

V_S = subice volume (m^3); at time of maximum ice thickness

T_I = ice mass of HTO (dpm); at time of maximum ice thickness

V_I = ice volume (m^3 - water equivalents); at time of maximum ice thickness

For model M1, the HTO and H_2O datasets for 1980 and 1981 were combined with the assumption that for the outflow volume measured or predicted, the proportions for each the lake compartment did not change from one year to the next. From these pooled data it was possible to obtain estimates (m^3) for the proportion of each compartment to the total outflow volume. The turnover time for each compartment (TT_c) was predicted by:

$$TT_c = \left(\frac{VW_c}{\sum_{c=1}^{3} VW_c} \times VL \right) \Bigg/ \left(\frac{VH_c}{\sum_{c=1}^{3} VH_c} \times VO \right)$$

where: TT_c = annual turnover time (yr)

VW_c = mean annual volume of compartment c (m^3); as determined from the water budget

V_L = mean annual volume of lake (232,700 m^3)

VH_c = mean annual volume (m^3) of compartment c as determined from HTO and H_2O datasets

V_0 = mean annual volume of compartment c lost to the outflow (m^3)

Daily HTO losses from P & N Lake outflow provided estimates of the losses (m^3) from the labelled subice compartment (1980) or subice/ice compartments (1981). In order to compare these results with those predicted by the traditional model (formula 1), a second model was developed to predict daily HTO losses from the labelled compartment(s).

Model M2 predicted V_P as:

$$V_p = PT_O \Big/ \frac{PT_S + PT_I}{PV_S + PV_I}$$

where: V_p = volume lost from labelled compartment (m^3)

$$PT_O = \frac{(M_L + M_I - M_O)}{(V_L + V_i + V_I - V_O)} \times \text{outflow discharge}$$

PT_S = predicted HTO of subice (dpm)

PV_S = predicted volume of subice (m^3)

PT_I = predicted HTO of ice (dpm); 1981 only

PV_I = predicted volume of ice (m^3); 1981 only

M_L = total lake HTO of previous day (dpm)

M_I = mass of HTO melted from ice (dpm); 1981 only

M_O = mass of HTO lost via outflow (dpm)

V_L = total lake volume of previous day (m^3)

V_i = inflow volume (m^3)

V_I = ice volume (m^3 - water equivalent); 1981 only

V_O = outflow volume (m^3)

To provide and independent comparison of the losses predicted by the traditional model and the observed data, measured conductance was compared with those estimates obtained using the water budget data on a daily basis.

RESULTS

Profiles.

The shape of the tritium, conductivity and temperature profiles during the ice-on melt period were similar in 1980 and 1981 (Fig. 10), except for the tritium profile. Profiles showed stratification, with a thin layer of less dense, colder water with lower salt concentration, located just below the ice. As springmelt continued, both the icemelt and snowmelt runoff layers beneath the icepan became thicker, with only a small encroachment into the deeper layer, as shown by the HTO data (Fig. 11). Simultaneously a shallow moat, which remained completely mixed, as determined by HTO and conductance measurements, formed between the icepan and the lake edge (not shown).

Dye experiments.

Four dye experiments, one ice and three inflow additions, showed the pattern of subice spring meltwater. All inflow Rhodamine B additions provided similar results; the dye extended from the inflow in a narrow plume beneath the ice and became wider and deeper before reaching the outflows.

At P & N Lake (27 May - 6 June 1980) the inflow rate was about 5 l s^{-1} and the dye was carried out in all directions from the source, forming a thin layer beneath the ice and becoming deeper over the durarion of the experiment (Fig.12 and 13).

Figure 10: Tritium (HTO), conductivity and temperature profiles for P & N Lake, one site shown for 6 June 1980 and one site for 16 June 1981

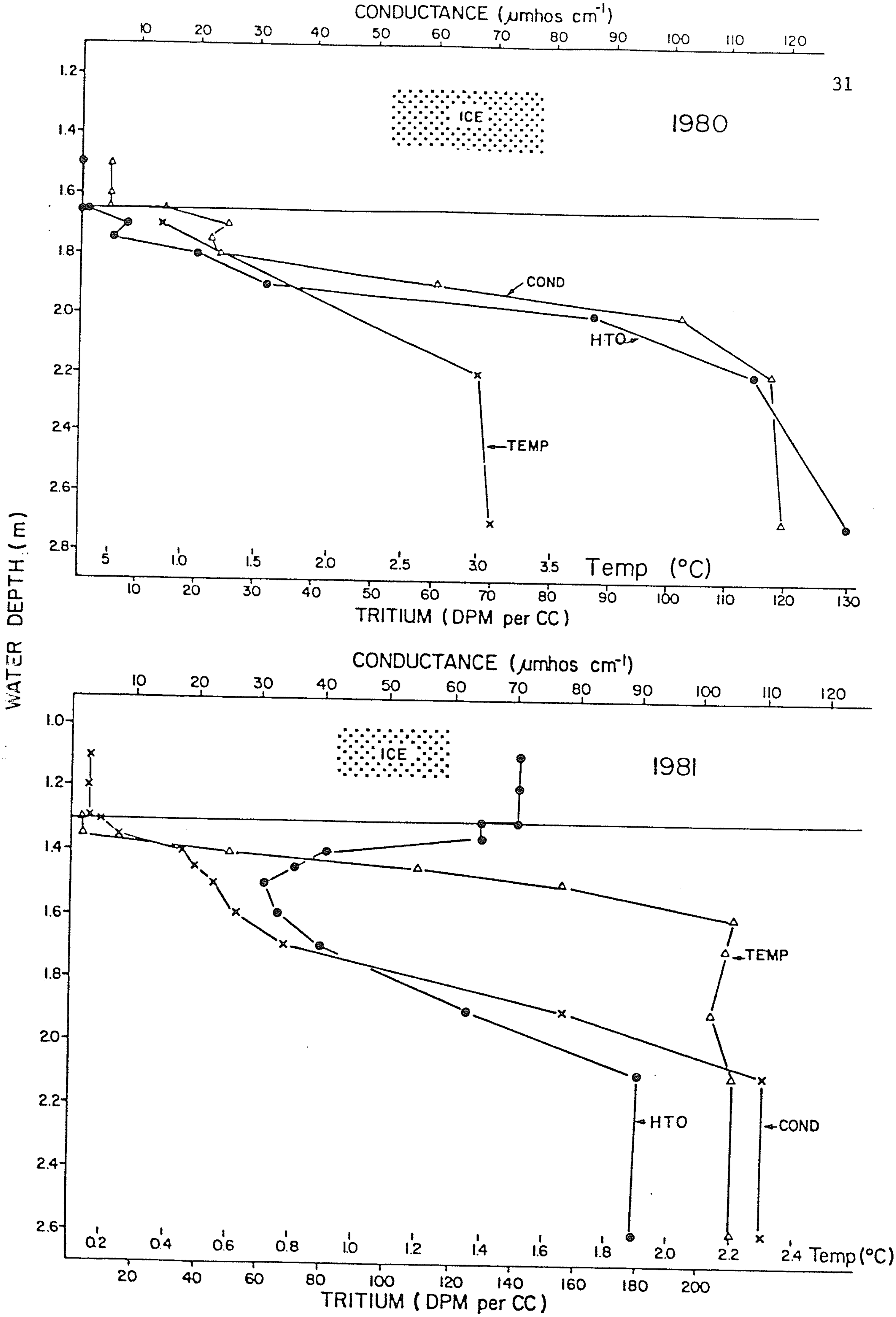

CONDUCTANCE (μmhos cm⁻¹)
ICE
1980
COND
HTO
TEMP
Temp (°C)
TRITIUM (DPM per CC)
WATER DEPTH (m)
CONDUCTANCE (μmhos cm⁻¹)
ICE
1981
TEMP
HTO
COND
Temp (°C)
TRITIUM (DPM per CC)

Figure 11: P & N Lake tritium (HTO) profile for Site 12 (1980) and Site 2 (1981) located approximately 150m south of outflow

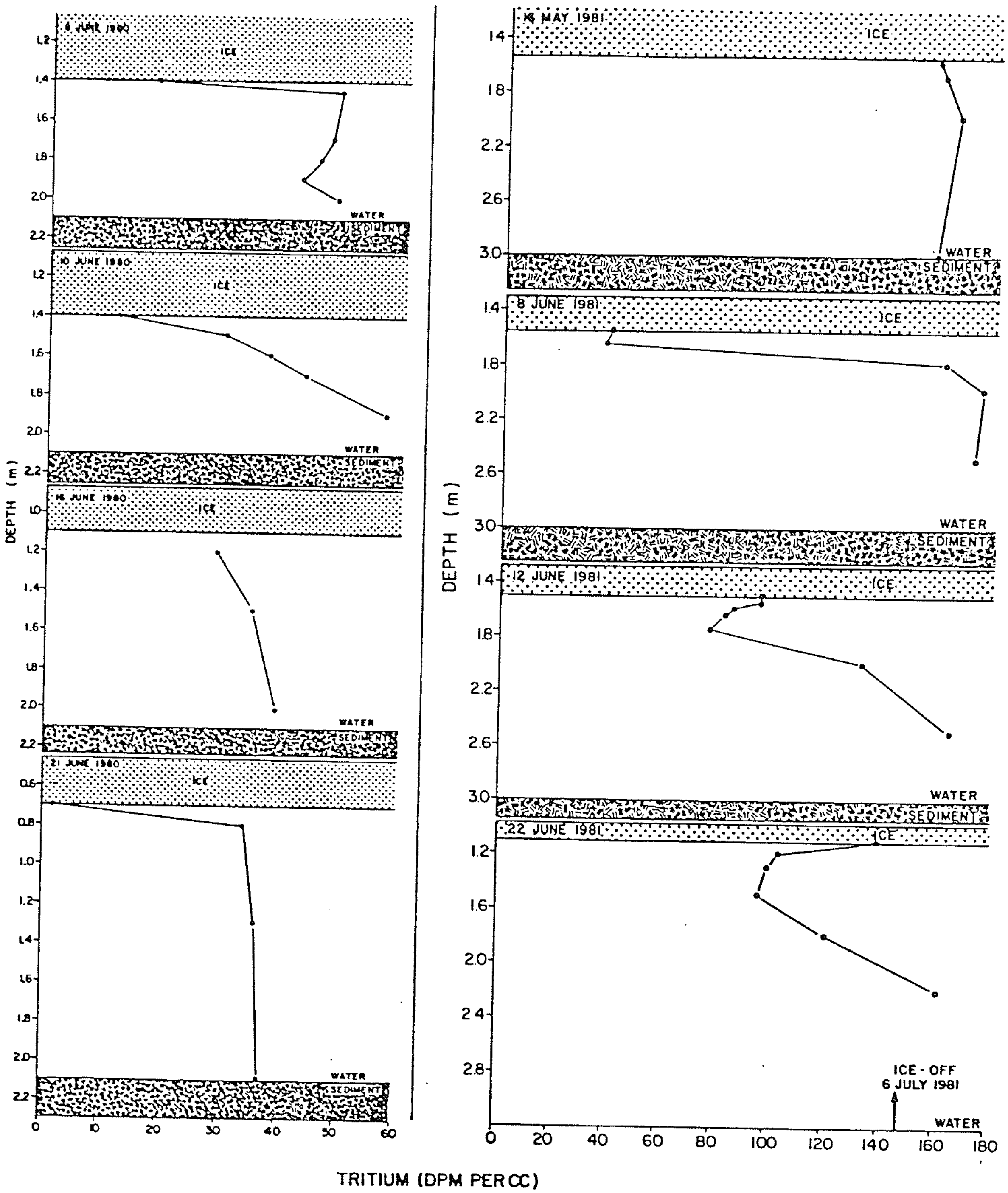

8 JUNE 1980
10 JUNE 1980
16 JUNE 1980
21 JUNE 1980
16 MAY 1981
8 JUNE 1981
12 JUNE 1981
22 JUNE 1981
ICE
WATER
SEDIMENT
DEPTH (m)
ICE - OFF
6 JULY 1981
TRITIUM (DPM PER CC)

Figure 12: P & N Lake dye distribution, plan view, 27 May-6 June 1980

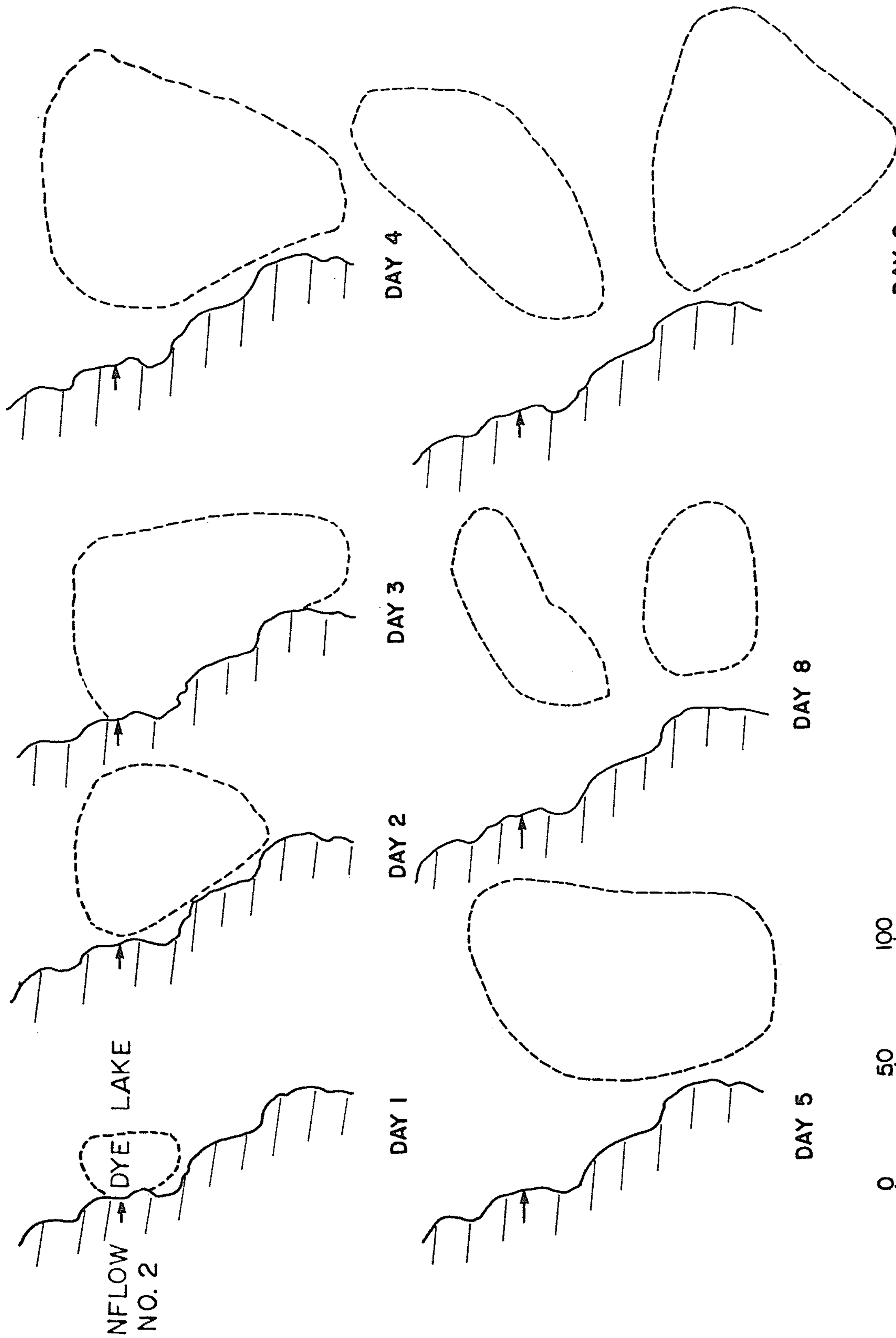
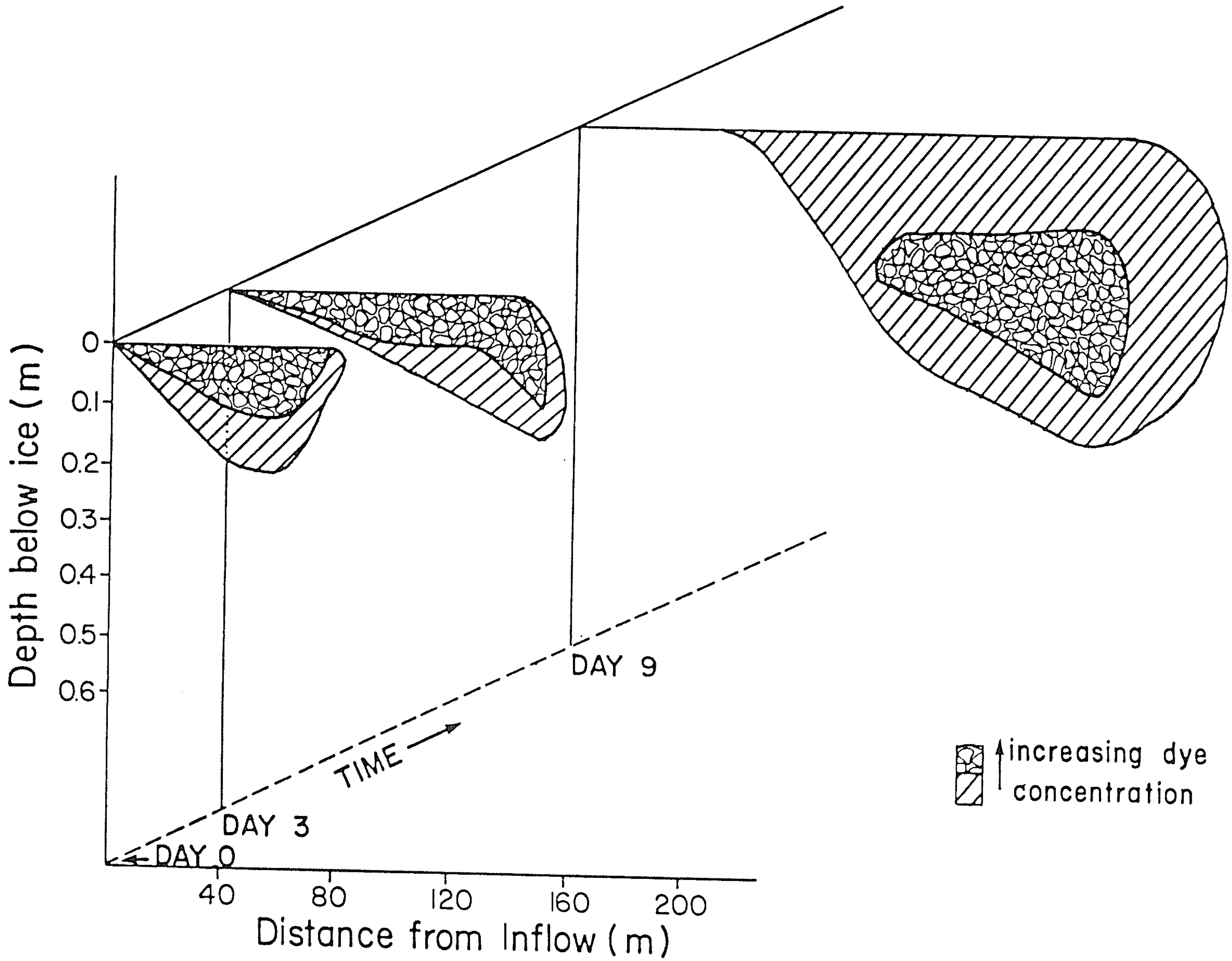

INFLOW
NO. 2
DYE
LAKE
DAY 1
DAY 2
DAY 3
DAY 4
DAY 5
DAY 8
DAY 9
0
50
100
metres

Figure 13: P & N Lake dye distribution, plume depth and length, 27 May -6 June 1980 (very low flow rate)


Depth below ice (m)
0
0.1
0.2
0.3
0.4
0.5
0.6
DAY 0
DAY 3
DAY 9
TIME
40
80
120
160
200
Distance from Inflow (m)
increasing dye concentration

The initial dye distribution from Spring Lake (4-7 June 1980) inflow (about 20 l s^{-1}) was in a narrow band, located a few cm below the bottom of the ice (Fig. 14). The dye reached the outflow four days after the addition, with the highest concentrations in the lake still just below the ice.

Dye was added to Hawk Lake inflow (12-14 June 1981) at peak flow (about 50 l s^{-1}) and was found at the outflow after 33 hours. Highest Rhodamine B concentrations were again found a few cm beneath the ice (Fig. 15).

These three experiments indicate that the meltwater inflows: 1) proceed toward the outflows as vertically discrete masses, 2) tend to concentrate a few cm below the subice surface rather than immediately adjacent to it, and 3) tend to mix more vertically as inflow volume increases.

The addition of dye to the ice surface of Turkey Lake (17 June 1980) illustrated that the dye poured down through the ice with the highest concentrations found contiguous to the ice (Fig 16). Because the amount of dye added was small, the dilution was rapid and no dye was found >3 m from the addition site. Thirty-six hours after addition, the dye was no longer detectable. This experiment showed that the icemelt water remains immediately below the ice.

HTO mass balance and compartment renewal times.

The daily P & N Lake discharge record for 1980 showed a spring melt

Figure 14: Spring Lake dye distribution, plume depth and length, 4-7 June 1980 (low flow rate)

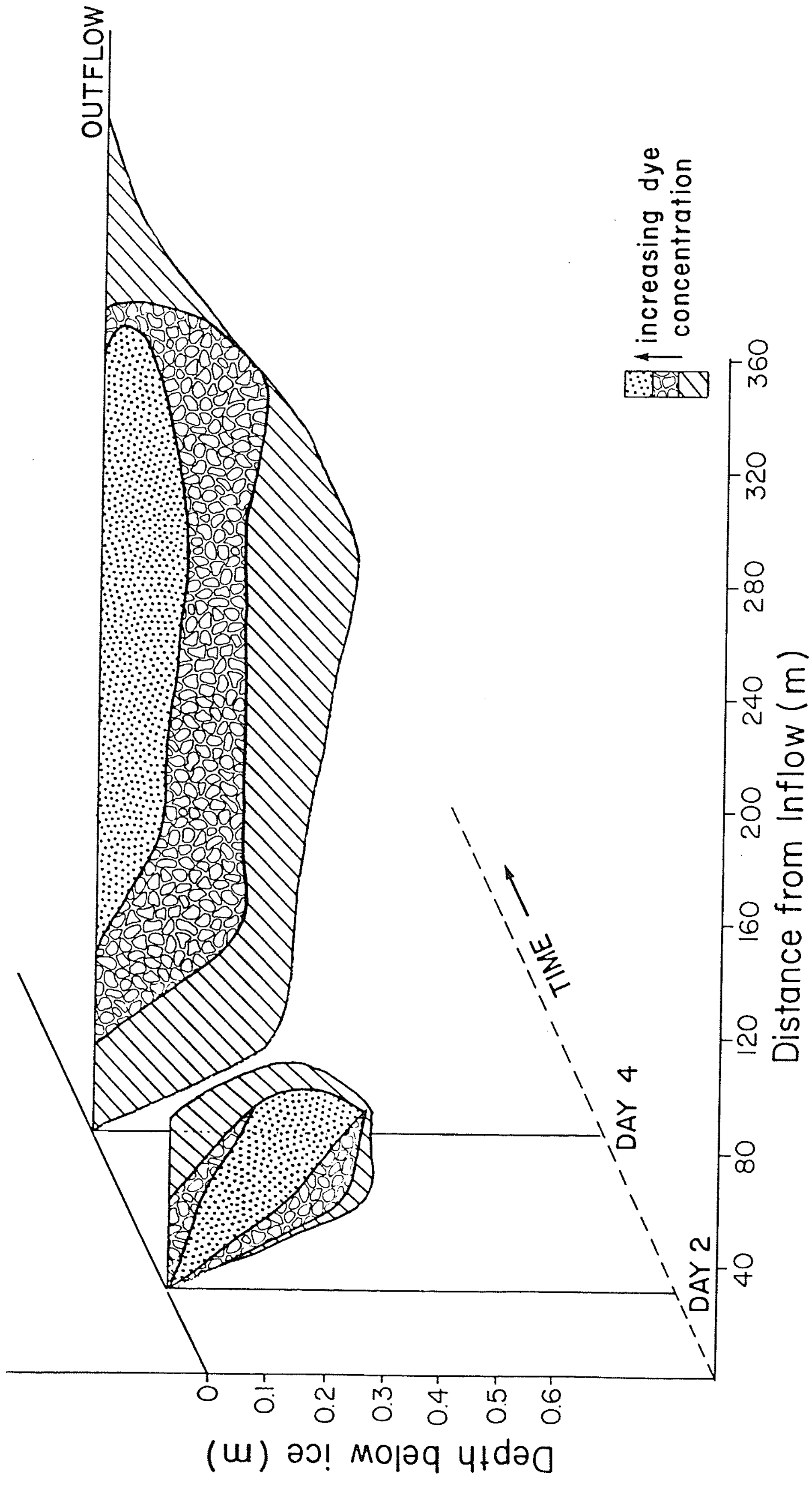

OUTFLOW
increasing dye concentration
TIME
DAY 2
DAY 4
Depth below ice (m)
0
0.1
0.2
0.3
0.4
0.5
0.6
40
80
120
160
200
240
280
320
360
Distance from Inflow (m)

Figure 15: Hawk Lake dye distribution, plume depth and length, 12-14 June 1981 (high flow rate)

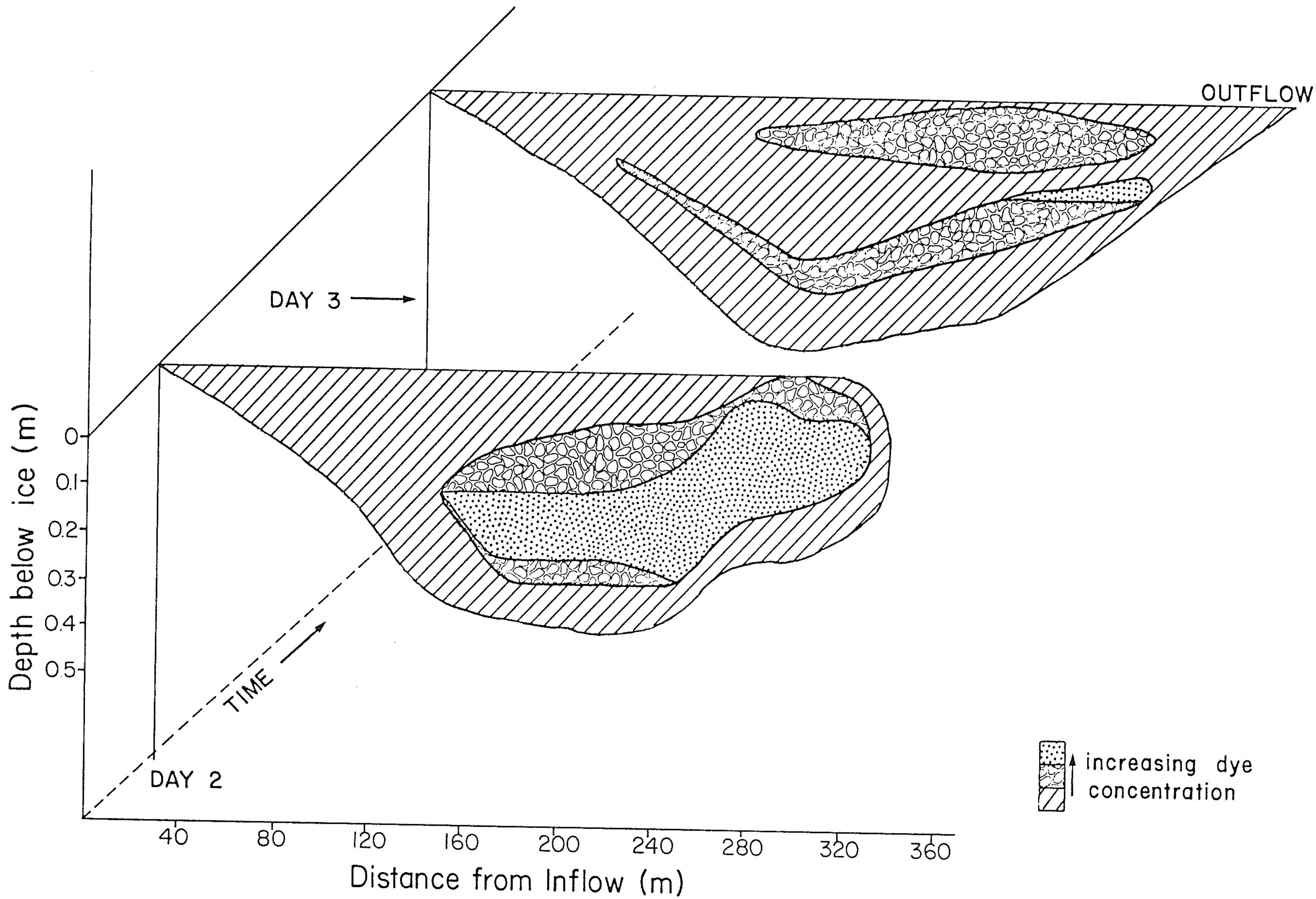
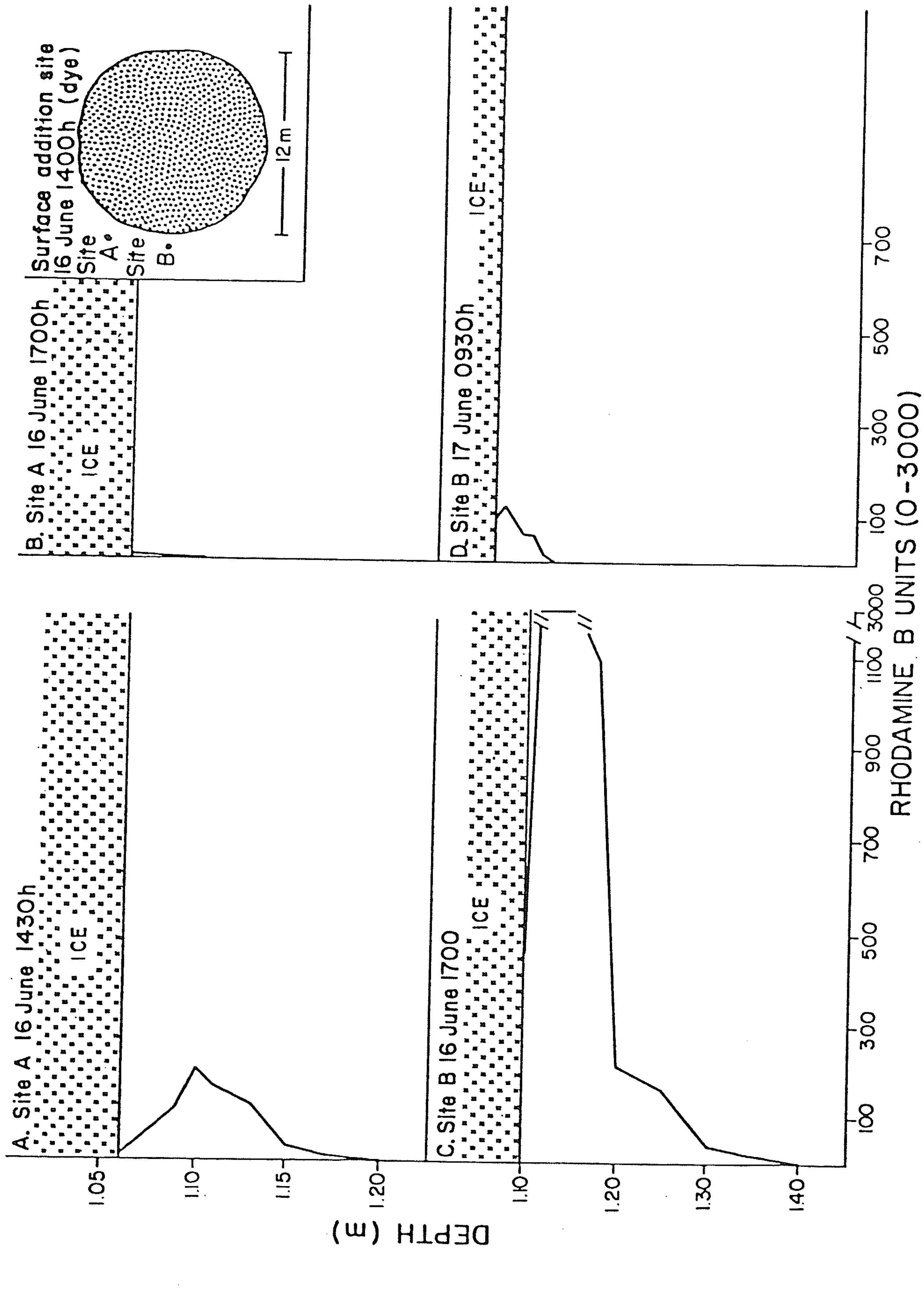

OUTFLOW
DAY 3
DAY 2
TIME
Depth below ice (m)
0
0.1
0.2
0.3
0.4
0.5
Distance from Inflow (m)
40
80
120
160
200
240
280
320
360
increasing dye concentration

Figure 16: Turkey Lake dye profile, 16-17 June 1980


A. Site A 16 June 1430h
ICE
B. Site A 16 June 1700h
ICE
Surface addition site
16 June 1400h (dye)
Site A
Site B
12 m
C. Site B 16 June 1700
ICE
D. Site B 17 June 0930h
ICE
DEPTH (m)
1.05
1.10
1.15
1.20
1.10
1.20
1.30
1.40
100
300
500
700
900
1100
3000
100
300
500
700
RHODAMINE B UNITS (0-3000)

peak prior to ice-off (Fig. 17), but 66% of the total discharge occurred between ice-off and freeze-up (108 d or 80% of the total outflow period), because of several large summer precipitation events. For 1981 the daily discharge record also showed a spring melt peak before ice-off (Fig. 18), but in contrast to the previous year, 95% of the total discharge occurred during the ice-on period (25 d or 41% of the total outflow period). These and other physical data comparing differences between the 1980 and 1981 water budgets are shown for P & N Lake and are presented in Table 1.

Lake HTO concentrations were corrected on a daily basis for the measured losses to the sediments (Fig.19) and the estimated losses via vapor exchange (Table 2). The theoretical turnover time (TTc) was predicted from model M1 for each of the three compartments (c); subice, ice and in-outflow. Table 3 shows the different results determined for 1) the combined hydrological and water budget data (H_2O dataset), and 2) the combined tritium (HTO) and water budget data (HTO dataset). Model M1 predicted that the subice compartment (TT_s) was underestimated and in-outflow (TT_i) overestimated for the annual period (first melt to outflow freeze-up). The volume lost from each lake compartment to the outflow was expressed as a percentage of the total outflow loss (Table 4).

In order to show the chronology of differences between the predicted and measured losses, first, estimates of the daily volume lost for the labelled compartment were obtained from the outflow HTO data (samples taken at the outflow). Second, these measured data were compared with those predicted using the HTO dataset and model M2 for the entire ice-on

Figure 17: P & N Lake daily outflow discharge, 1980

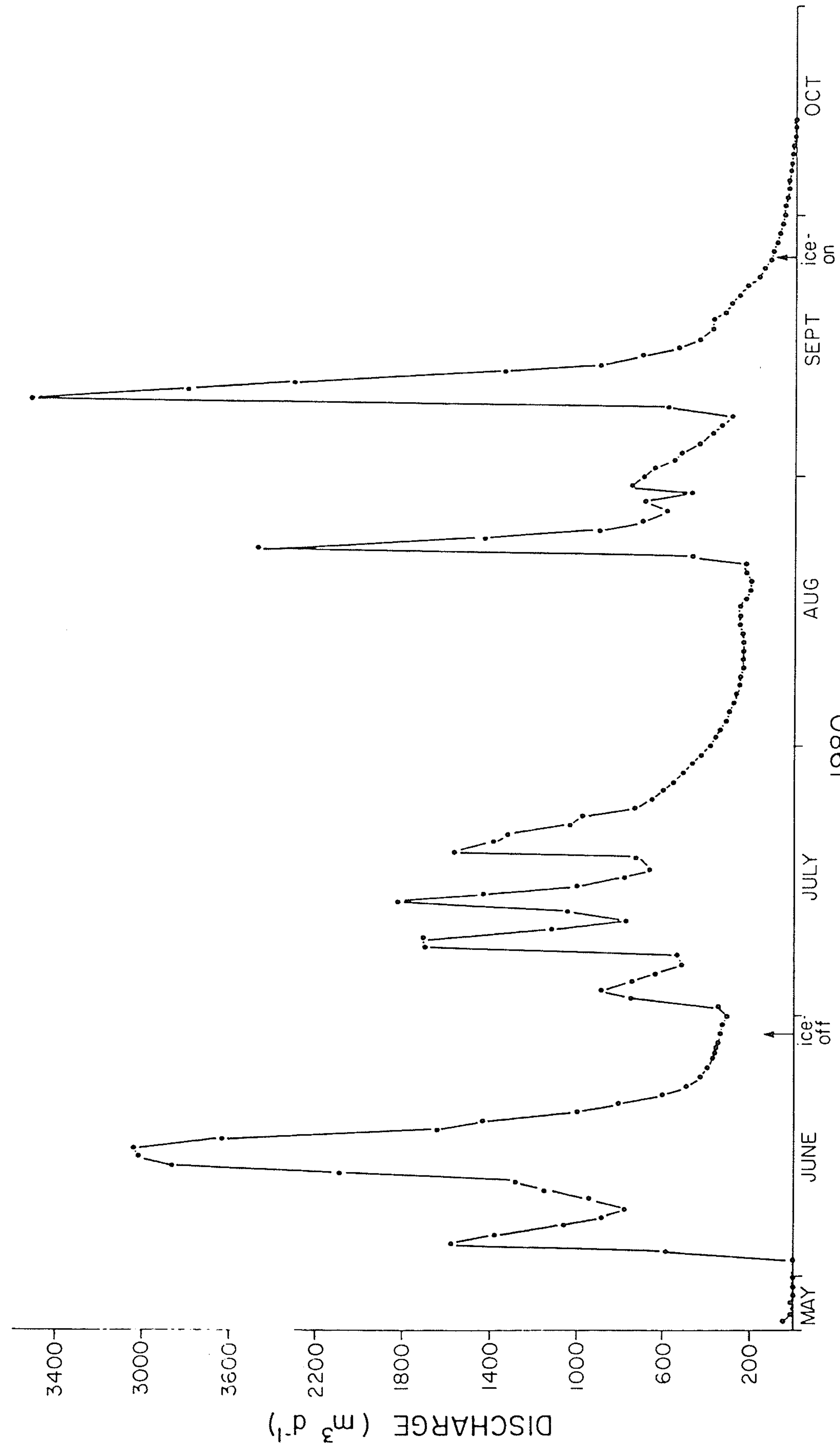
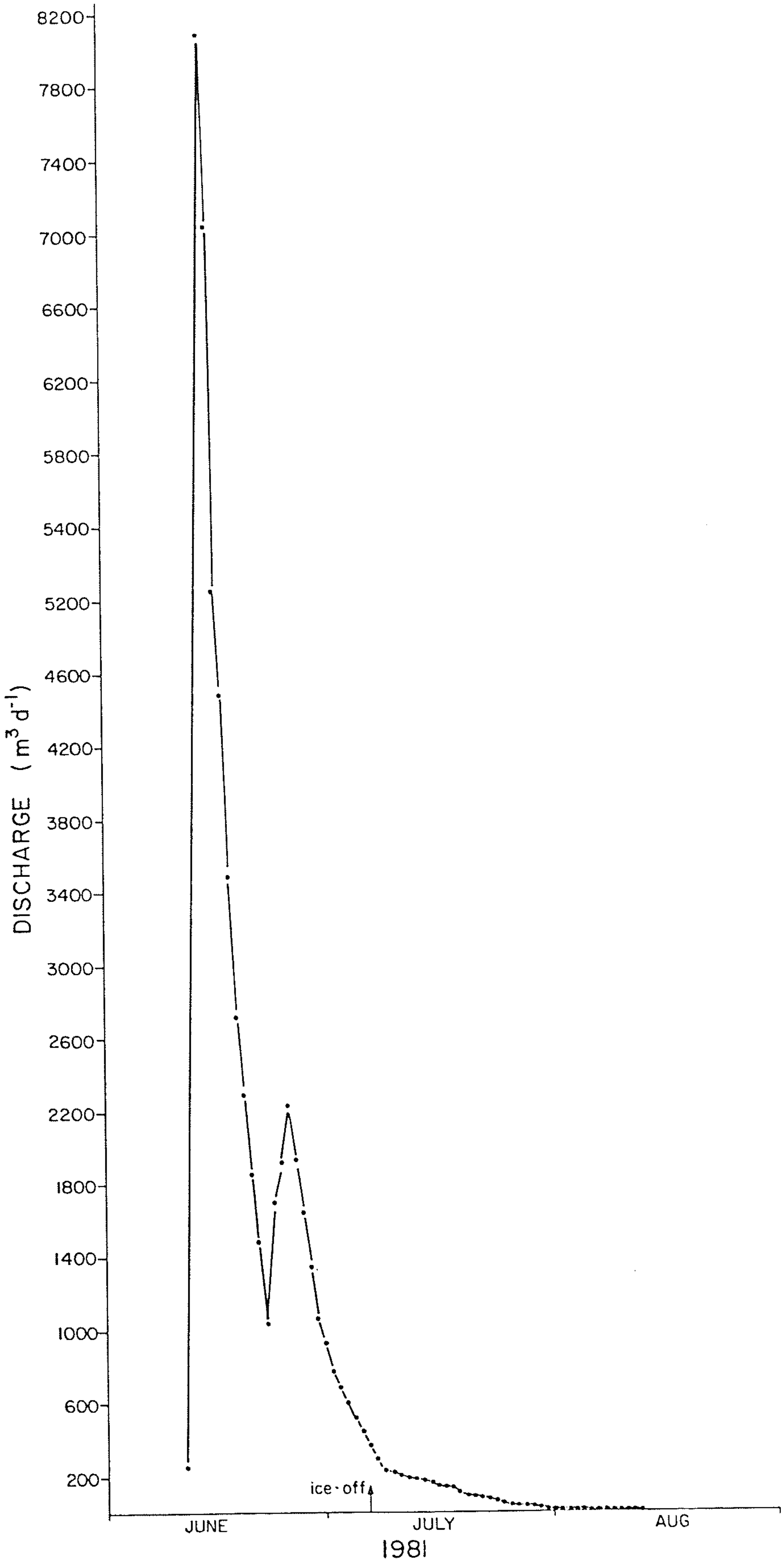

DISCHARGE (m³ d⁻¹)
3400
3000
2200
1800
1400
1000
600
200
1980
MAY
JUNE
JULY
AUG
SEPT
OCT
ice-off
ice-on

Figure 18: P & N Lake daily outflow discharge, 1981



8200
7800
7400
7000
6600
6200
5800
5400
5200
4600
4200
3800
3400
3000
2600
2200
1800
1400
1000
600
200
DISCHARGE (m3 d-1)
ice-off
JUNE
JULY
AUG
1981

Table 1. Physical data for P & N Lake, 1980 and 1981.

	1980	1981 (To 31 August)
Outflow Vol. (m^3)	9.42×10^4	6.70×10^4
Input Vol. (m^3)	1.77×10^5	7.10×10^4
Ice Vol. (m^3) *	8.91×10^4	8.83×10^4
Subice Vol. (m^3) *	1.35×10^5	1.36×10^5
% Annual Discharge from first melt to ice-off	33.6 (27 days)	95.0 (25 days)
% Annual Input from first melt to ice-off	33.7 (27 days)	88.4 (25 days)

* Note: At time of maximum ice depth

Figure 19: Sediment HTO profile as predicted for numeric model and actual pore peeper data, P & N Lake 22 September 1980 (diffusion rate = 1.3 X 10^{-5} cm^2 s^{-1}).

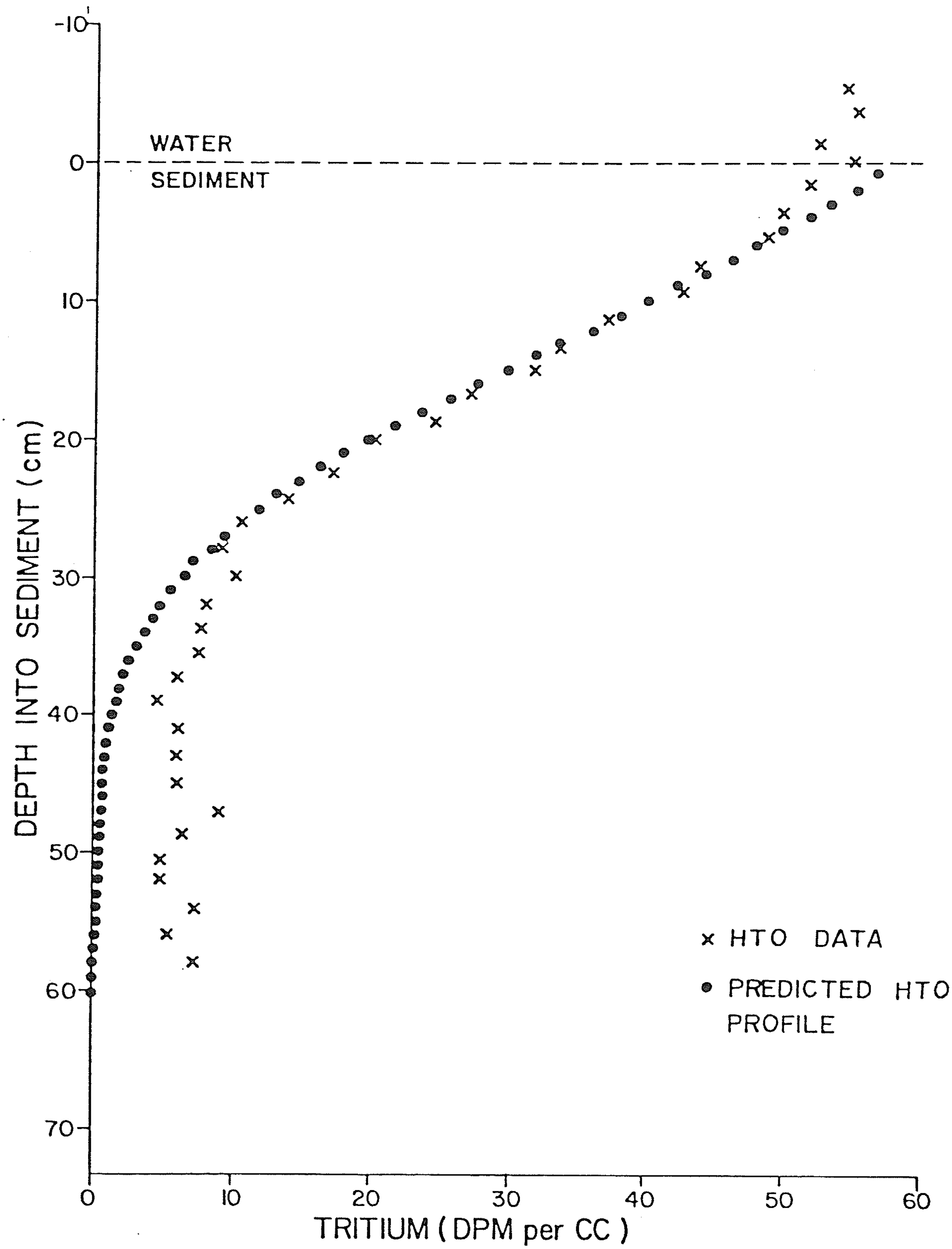

WATER
SEDIMENT
DEPTH INTO SEDIMENT (cm)
-10
0
10
20
30
40
50
60
70
TRITIUM (DPM per CC)
0
10
20
30
40
50
60
× HTO DATA
● PREDICTED HTO PROFILE

Table 2. HTO losses from P & N lake via the sediments and atmospheric vapour exchange, 1980 and 1981.

Loss to Sediments

Date	Flux (dpm $lake^{-1}$ d^{-1})
16 May 1980 - 2 Sept. 1980	5.65×10^{9}
3 Sept. 1980 - 22 Apr. 1981	1.04×10^{10}
23 Apr. 1981 - 24 Aug. 1981	1.03×10^{9}

* Loss to Atmosphere (V.E.)

Date	Flux (dpm $lake^{-1}$ d^{-1})
6 June 1980 - 26 Sept. 1980	3.03×10^{10}
20 June 1981 - 31 Aug. 1981	1.43×10^{11}

* Note losses of HTO to atmosphere via vapour exchange during ice-on were calculated as: ice-off X daily HTO-flux.

Table 3. Annual turnover times predicted by model M1 for the three P & N lake compartments based on the HTO and H_2O datasets.

Compartment	Annual Period (yr)	
	HTO	H_2O
Subice	3.5	2.9
Ice	3.4	2.9
In-Outflow	1.9	2.9

Table 4. Relative % and volume (m^3) of each compartment of P & N lake lost to outflow for the combined period 1980 - 1981, based on the HTO and H_2O budget data.

Period	Compartment	HTO	H_2O
Annual	Subice	(35.4) 57,052	(43.1) 69,487
	Ice	(25.9) 41,753	(30.7) 49,486
	In-Outflow	(38.7) 62,388	(26.1) 42,102
Ice-on	Subice	(23.0) 21,901	(47.7) 45,479
	Ice	(39.9) 37,994	(33.9) 32,307
	In-Outflow	(37.1) 35,328	(18.3) 17,468

period (Fig. 20). Differences between estimates were greatest at the onset of springmelt and became zero at ice-off.

When correcting the liquid P & N Lake water for all salt inputs, (icemelt, precipitation and runoff), the predicted conductivity values differed from that measured, again with the greatest differences at the onset of melt (Fig. 21).

Figure 20: Daily percent differences between data predicted from model M2 and the measured data for labelled compartment volumes lost via P & N Lake outflow, 1980 and 1981

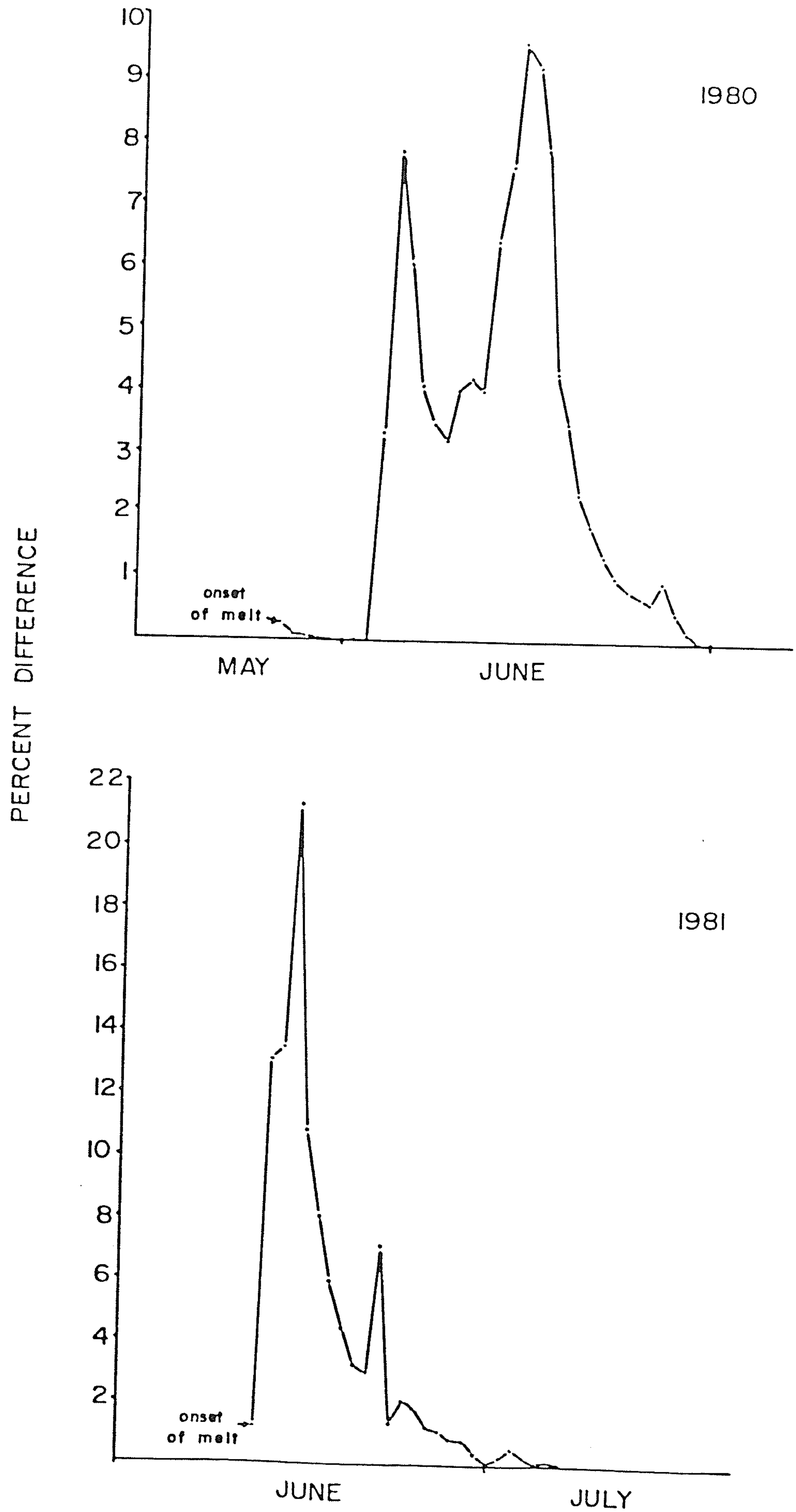
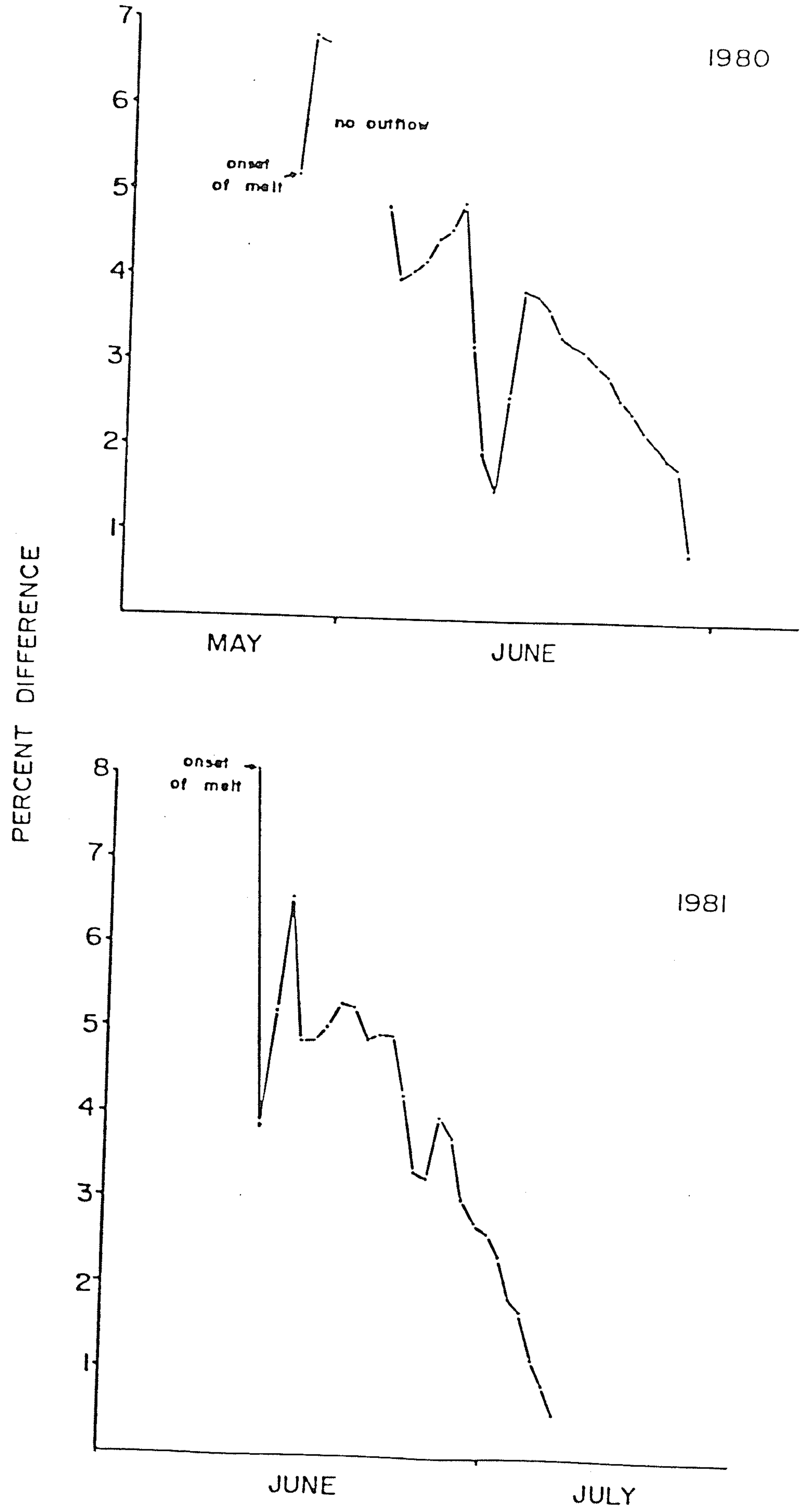

1980
10
9
8
7
6
5
4
3
2
1
onset
of melt
MAY
JUNE
PERCENT DIFFERENCE
1981
22
20
18
16
14
12
10
8
6
4
2
onset
of melt
JUNE
JULY

Figure 21: Daily percent differences between estimates predicted from the water budget data and measured estimates for conductivity at P & N Lake outflow, 1980


1980
no outflow
onset of melt
MAY
JUNE
PERCENT DIFFERENCE
1981
onset of melt
JUNE
JULY

DISCUSSION

Profiles.

In P & N Lake the subice stratification which was set up early in spring melt and substantiated by the tritium, conductivity, temperature and dye profiles revealed three distinct layers beneath the ice: icemelt on top, followed by inflowing snowmelt from runoff, and subice lakewater on the bottom. Evidence from total phosphorus and silica profiles for P & N Lake (Welch 1980, unpubl. data) revealed that inflowing particulate matter and labile substances also conformed to this stratification. Figures 11 and 12 show that as icemelt and runoff continued, the meltwater layer became thicker, with a small amount of mixing with the subice water. This limited vertical mixing persisted until ice-off, which was also the time of spring turnover, and is typical of arctic lakes (Schindler *et al*. 1974, Minns 1977, LaPerriere 1981).

Flow from the outlet begins after the start of lake inflow. The initial P & N Lake outflow consisted of cold (<1^{o}C), low conductivity meltwater with no distinction between the icemelt (ice compartment) and runoff water (in-outflow compartment) as determined from the HTO data. A similar occurrence was first noted by Schindler *et al*. (1974) and has since been shown elsewhere (Henriksen and Wright 1977, Jeffries *et al*. 1979).

The vertical distribution of dye for P & N, Spring and Hawk lakes also showed that inflowing water (snowmelt) flowed beneath the ice with little mixing with deeper water (Fig. 15-17). Runoff water entered the lake via the inflows and, with the mixing of the ice-surface snowmelt and the upper layer of low density water, formed a leading edge at a depth between 0 and 100 cm below the bottom of the ice.

The spatial distribution of the runoff water was a function of its volume and density. The dye addition to the P & N Lake inflow took place during a period of early flow. Shortly after the dye addition a period of cold weather caused the inflow and outflows to begin freezing, and 24 hours after addition the flow stopped. The dye slowly moved throughout a large portion of the lake, the dye still just below the ice, and the energy required to distribute the dye can probably be attributed to the residual water movement set up by the inflows and density differences within the lake. When warmer weather caused flow to resume four days later, the dye had become widely dispersed throughout the upper one meter beneath the ice, and with the subsequent decrease in concentration below threshold, a further description of dye movement was impossible.

The Spring Lake and Hawk Lake inflow experiments took place during periods of constant flow which resulted in a faster movement of dye to the outflow. The pattern of dye distribution was somewhat different for the Hawk Lake experiment (12 to 14 June 1981) than that found for Spring Lake (4 to 7 June 1980), due to the greater Hawk Lake inflow volume. The dyed inflow water was initially found between 10 to 40 cm beneath the bottom of the ice, in a direct band to the outflow. The fastest moving water, the topmost layer, was characterized by the rapid dilution and then total replacement of the dye in the layer 20 to 30 cm below the ice while the slower water was found between 0 to 20 cm and 30 to 60 cm below the ice and showed an increased residence time (Fig. 16). This indicated that the inflowing water (in-outflow compartment) had the greatest velocity in the water column and therefore the shortest residence time.

These results allow better interpretation of the data from lakes at lower latitudes, where during melt the concentrations of various materials in a water layer immediately beneath the ice are often very different from concentrations in the lake water below. Prentki et al. (1979) report that during springmelt a layer forms beneath the ice at Lake Wingra, Wisconsin which was composed of dilute, cold icemelt and runoff from shore. Samples were collected from Lake Wingra at 0.5 m and 1.5 m (ice thickness approximately 0.3 m) and showed that suspended solids, particulate phosphorus, dissolved reactive phosphorus, total nitrogen, and filtered biochemical oxygen demand were higher in the runoff layer (0.5 m) than the lake water (1.5 m) thus indicating a lack of vertical mixing between compartments. More recently a double inflow spike (HTO and dye) was carried out at L240 (Experimental Lakes Area, Ontario) which showed the layering of inflowing water beneath the ice (Linsey and Hesslein, pers. comm.).

The layering of inflowing water beneath the ice has also been shown by Prepas (pers. comm.) for Bob Lake (Haliburton, Ontario), where she found that total phosphorus and total dissolved phosphorus measurements were similar at the inflow and directly beneath the ice, higher than the rest of the lake water. Acid precipitation studies on snowmelt in central Ontario (Jeffries et al. 1979), eastern United States (Hendrey *et al.* 1980), Sweden (Hultberg 1977) and Norway (Groterud 1972, Henriksen and Wright 1977), have shown that during ice-cover the inflowing runoff (which has a low pH) forms a thin layer beneath the ice and is associated with the litteral zone. Trace metal studies in Manitoba (Franzin and McFarlane 1981) also show this layer of snowmelt (direct or runoff) beneath the ice. None of these studies have quantified the volumes and flows of the meltwater layers.

Compartment renewal times.

The daily discharge from P & N outflow was very different for 1980 and 1981, with 1980 wetter and 1981 drier than the 5 year (1977-1981) mean. In 1980, with major precipitation events occurring on three occasions during the ice-off period, 30% more water was lost from the outflow than in 1981, and the outflow continued flowing until freeze-up on 26 September. In 1981, flow ceased on 11 August, and only 5% of the total outflow occurred after ice-off.

Although the two water years were very different it was necessary to combine the HTO data of both years to determine the role of each compartment in the water budget. In order to do this it was assumed that during any year and within a specific time interval (ice-on, ice-off, annual) the compartment proportions which exit via the outflow remain constant. It was possible to determine these proportions based on the HTO and H_2O datasets (model M1). Table 4 indicates that for both time intervals (ice-on and annual period), the estimate of volume lost differed between the HTO and water budget data with the greatest deviation during the ice-on period. The theoretical turnover time for each compartment (TT_c, as predicted by model M1), based on the water budget data (H_2O dataset), was higher for the subice and lower for the in-outflow than determined from the HTO data (Table 2).

Based on the TTc determined using both the HTO and H_2O datasets, for lakes with no ice-cover and assuming the lakewater is completely mixed and knowing the lake's water budget, the difference between actual and predicted losses (for all compartments) via the outflow should be zero. However, in the arctic the overriding factor which reduces whole-lake mixing is the duration of the ice-cover period. In a year when most precipitation occurs prior to and during melt (but before ice-off) the difference between measured and predicted TTc will be large, with a large proportion of in-outflow and a small proportion of subice loss, as for P & N Lake in 1981. When most precipitation falls during the ice-off period when the lakes are well mixed, the difference will be smaller (Fig. 3).

The greatest proportion of in-outflow loss from the lake to the outflow for any year should occur when ice leaves the lake late in the year, when inflows cease early and when summer precipitation is low. The smallest proportion of in-outflow should occur when the ice-cover period is short, winter snow accumulation is small and high inflow volumes occur during the open water season. For lakes where ice-cover and/or density differences prevent whole-lake mixing the actual TTc for any year will depend on the amount of precipitation, season of precipitation, duration of the ice-cover period, and the volume of the lake basin relative to the size of the tributary drainage area.

Flux to the sediments.

From the onset of ice-cover to first outflow loss, HTO losses to the sediments as well as the incorporation into the ice were the only losses of HTO from the lakewater (subice). However at all other times of the year losses to the sediments were <5% of the total HTO loss.

The flux of HTO to the sediments depends on the diffusion coefficient of HTO in the sediments and the concentration gradient of HTO at the sediment-water interface. For P & N Lake both the diffusion coefficient and concentration of HTO varied during the course of the experiment (May 1980 to August 1981) and as a result the flux varied continually. Estimates of flux were made for three time periods corresponding to the three peeper profiles: 16 May to 2 September 1980 (Period 1; 1.3 X $10^{-5} cm^2\ s^{-1}$) 3 September 1980 to 21 April 1981 (Period 2; 5.0 X $10^{-5} cm^2\ s^{-1}$) and 22 April to 22 August 1981 (Period 3; 1.3 X $10^{-5} cm^2\ s^{-1}$). In 1976 at Lake 227 (Experimental Lakes Area, Ontario) Hesslein (1980a) determined a diffusion coefficient for HTO in the sediments of 1-2 X $10^{-5} cm^2\ s^{-1}$ for the summer stratification period which is close to that found for Periods 1 and 3. The higher diffusion coefficient for Period 2 probably reflects fall turnover when a density inversion occurred in the sediments. The low density interstitial water, which had been heated to above $4^{o}C$, was replaced with cooled lake water, which at the time of freeze-up was below $1^{o}C$. This density inversion has been proposed elsewhere for temperate lakes (Hesslein 1976b) and was measured in both Lake 227 and an Alaskan pond (Musgrave and Reeburgh 198x).

Hesslein (1980a) found that based on the diffusion coefficient and the water content of the Lake 227 sediments the observed diffusion can be accounted for by molecular processes only. Although the water content of P & N Lake sediments is somewhat less than that found by Hesslein (Table 5), diffusion by molecular processes is also occurring in the Saqvaqjuac lakes.

A comparison of gross evaporation as calculated by HTO mass balance and measured relative humidity.

Throughout the ice-off period atmospheric losses of HTO from P & N Lake include net surface water loss in the form of net evaporation and loss due to vapor exchange. The daily loss via vapor exchange was always greater than that lost to the sediments (Table 2).

Gross evaporation (GE) is defined as:

$$GE = \text{net evaporation}/(1-\text{relative humidity})$$

(Hesslein, pers. comm.)

However since GE is difficult to measure, it is usually obtained by difference. P & N Lake offered an opportunity to obtain an independent estimate of GE using the HTO data. From the differences between the measured HTO lost from the outflow and that estimate of HTO lost determined from the water budget (sediment-loss corrected), daily estimate of lake surface GE were obtained. For 1981 the mean relative humidity at Saqvaqjuac (Allan, pers. comm.) was 77% which predicted a GE rate of 1388 $m^3\ d^{-1}$ over the open water period (to 31 August) whereas estimated from the HTO dataset the GE was 1270 $m^3\ d^{-1}$. For 1980 the estimated HTO predicted GE rate was lower, 764 $m^3\ d^{-1}$ compared to 1240 $m^3\ d^{-1}$.

Table 5. Water composition (% water) of cores taken from P & N lake (Aug. 1978) and L227 (Aug. 1975).

P & N lake*			L227**	
Depth Interval (cm)	% water		Depth Interval (cm)	% water
	6.5m core	10.0m core		1.5m core
0 - 2	95.5	95.6	0 - 2.2	95.9
2 - 4	93.2	94.6	2.2 - 5.2	96.2
4 - 6	94.0	94.3	5.2 - 7.6	96.5
6 - 8	93.8	93.6	7.6 - 9.9	96.9
8 - 10	93.6	93.2	9.9 - 11.9	96.8
10 - 12	93.3	93.0	11.9 - 14.1	96.8
12 - 14	93.1	92.9	14.1 - 16.5	96.7
14 - 16	92.9	92.3	16.5 - 19.0	96.5
16 - 18	92.2	91.0	19.0 - 21.4	96.4
18 - 20	92.6	91.3	21.4 - 23.8	96.4
20 - 22	91.5	90.4	23.0 - 26.2	96.3
22 - 24	91.5	90.4		

* Brunskill, pers. comm.

** Hesslein 1980a

The GE equation used for P & N Lake required that the lake water and air temperatures be equal for any unit of time. Thus, the reason for the difference between the daily GE rate as determined from the hydrometeorological data and the GE rate determined from the HTO data is perhaps due to the direction and magnitude of temperature differences between the air and lake water observed for 1980 and 1981, which were not considered when determining the GE rates shown above. Turbulent, rather than molecular diffusion is the process which is occurring at the air/water interface, and the diffusion rates were probably different during 1980 and 1981.

Implications for lake loading models.

The results of this study show that existing lake loading models, which assume that on an annual basis 100% mixing occurs between the subice, ice and in-outflow compartments prior to loss from the outflow, need re-evaluation. For all lakes which have ice-cover for part of their annual cycle complete mixing of inflowing water probably does not occur.

There are at least two main consequences of the layering of meltwater (snow, ice and runoff) beneath the ice and its subsequent loss to the outflow. First, Schindler et al. (1974) and Welch (pers. comm.) have postulated that the cryoconcentration which occurs in arctic lakes is in part due to the freeze-out of solutes from the ice which remain in the subice and are only partly lost to the outflow. Second, snow, which often contains large amounts of acids and heavy metals, may accumulate on lakes and drainage basins, to be released as meltwater in the spring (Elgmork et al. 1973, Hagen and Langeland 1973, Johannessen and

Henriksen 1978, Franzin and McFarlane 1981, Zajac and Grodzinska 1982). During snowmelt, the meltwater which layers beneath the ice is not diluted with the lakewater, often resulting in highly polluted waters in the upper lake strata, which may have deleterious effects on the biota (Groterud 1972; Leivestad and Muniz 1976). Prentki et al. (1979) found that such surface layering of spring meltwater prevented the weedbeds found at a depth of 1-2m from intercepting the nutrient-rich inflowing water, therefore creating a nutrient-limited environment.

Conclusion.

This study has shown that in arctic lakes with thick ice cover (1.5 - 2.0 m annually), actual lake turnover times may be significantly longer (perhaps 2 times) than the traditional theoretical model, which assumes complete mixing of inflowing water with the lake volume. This lack of complete mixing is due to spring ice-cover which maintains the lake stratification. The implications are: (1) Toxic contaminants coming in with spring meltwater will be concentrated in a thin layer beneath the melting ice, and in the moat extending around the shore, (2) The loading of phosphorus, nitrogen and other elements to lakes is actually less than would be predicted using models developed for temperate lakes, particularly for those that are relatively high in spring meltwater, because inflowing meltwater remains near the ice surface, mixes only slightly with deeper lakewater of higher density, and may leave the lake directly via the outflow. Refinements to lake loading models (eg. Vollenweider 1969; Dillon 1975) will have to take this effect into account when applying these models which require theoretical lake turnover times for subarctic and arctic lakes.

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