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Estimating the Water and Salt Budgets of a Stratified Estuary

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Small tidal range, low salinity, moderate temperature and salinity stratification during summer which becomes weaker during winter, abundant fresh water inflow and connection through a shallow sill with the adjacent sea are the characteristic features of some estuaries along the coast of the Baltic Sea. In this study the flow and transport of salt in the upper and lower layers of the outlet of an estuary of this type were determined. First the magnitude of the terms of the transport equations was estimated using the analogy between the transport of salt and heat. Then the important terms of the transport equation were computed with a model which is based on the measured slope of the water level and on the water budget of the estuary. Testing was carried out by comparing the measured and computed salt contents of the estuary. The operation of the model on Pohjanpitäjänlahti bay in southern Finland was found to be satisfactory.

Introduction

There are some estuaries along the coast of the Baltic Sea which are connected to the adjacent sea by a shallow sill. The difference between these and the true estuary of the fjord type stems from the fact that the tidal range of the Baltic Sea is small, and its salinity low. For this reason the effect of wind on transport processes may be higher and, due to temperature variation, stratification changes during different seasons perhaps more obvious.

Pohjanpitäjänlahti estuary was selected for this study. The area is particularly suitable for this purpose, because its physiographical features are rather simple. In addition a great deal of biological and chemical research has been carried out in this area; there are also plans about the future use of the area which require more information about the exchange properties of the estuary.

Earlier studies (Witting 1911; Granqvist 1911; Halme 1944; Niemi 1973) show that the salinity of the surface water of the estuary is low and that of the deep water almost as high as the salinity of the adjacent sea, or about 5‰. At the beginning of summer the temperature of the deep water is about 1 C°. During the summer, autumn and winter the temperature increases and the salinity decreases until March or maybe even earlier, when stratification is such that the cold sea water is able to enter the hollow of the estuary. The increase in temperature and decrease in salinity of the deep water are due to the mixing of the water masses in the estuary.

Since we have no quantitative understanding of the water exchange during summer and autumn, this study only considers the situation during this time. The exchange during winter has been illustrated by Granqvist (1911).

The study area

The whole estuary may be divided into three parts (Fig. 1), Pohjanpitäjänlahti bay (A), Dragsvikinlahti bay (B) and Kaupunginselkä bay (C). Table 1 gives some important information concerning the estuary. This information was obtained from the dredging plan made by the National Board of Public Roads and Waterways, depth soundings of the Board of Navigation, from maps (scale 1:20 000) and from (Sirén 1955) and from (Mälkki, Launiainen, Voipio 1974).

Table 1 - Physiographical features of the study area

Catchment area at point CV		2,500 km²	
Lake percentage		12	
	Part (A)	Part (B)	Part (C)
Area, km²	22	5	12
Volume, m³	$2.3 \cdot 10^8$	$0.1 \cdot 10^8$	$0.4 \cdot 10^8$
Max. depth, m	40	5	15
Mean depth, m	10	2	3
Greatest length, km	14	7	6
Greatest breadth, km	3	1	5
Greatest depth at the outlet, m	5	5	15

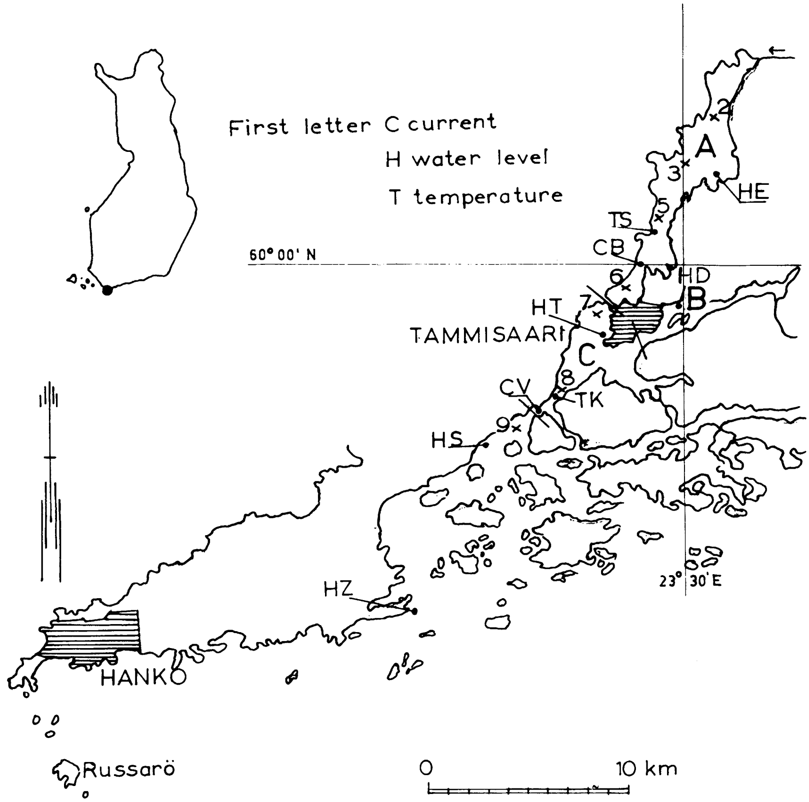

Fig. 1. Location map of the study area. (A) Pohjanpitäjänlahti, (B) Dragsvikinlahti, (C) Kaupunginselkä. (x) marks points of temperature and conductivity soundings.

Fig. 2 presents the longitudinal profile of Pohjanpitäjänlahti. The lines of equal salinity and density corresponding to the mean conditions between May 22 and October 10, 1973 have also been drawn into this figure.

In the following the values for the tidal constituents for Hanko computed by Lisitzin (1944) and Magaard and Krauss (1966) are given.

	Lisitzin 1932-1935	Magaard and Krauss 1958
O_1	0.93 cm	1.18 cm
K_1	0.69	0.79
M_2	0.88	1.21
S_2	0.44	0.69

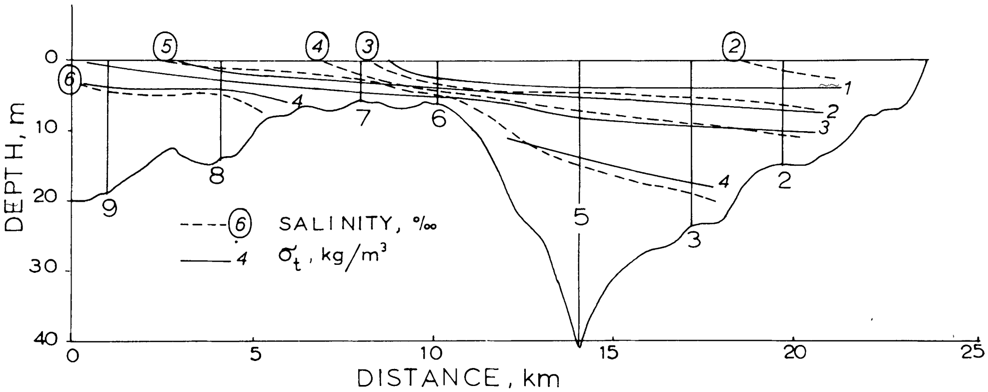

Fig. 2. The longitudinal section of Pohjanpitäjänlahti and Kaupunginselkä. Lines of equal salinity and density (σ_t) correspond to mean conditions between May 22 and October 10 1973. The sounding points are numbered as in Fig. 1. Published by the permission of the Board of Navigation.

The constituents computed from the present data for Pohjanpitäjänlahti did not deviate significantly from the values presented above.

The variation of inflow of fresh water is rather small (see Table 4). The reason for this is that the lake percentage of the catchment area of the estuary is rather high, and that the water course is regulated.

According to the classification by Hansen and Rattray (1966), an estuary may be considered as highly stratified if the densimetric Froude number F_m is greater than 0.01 and P the ratio of fresh water velocity to the root mean square tidal current speed, is greater than 0.1. Using the mean values from summer 1973, the values of these numbers were computed to be respectively 0.018 and 0.24 at the outlet of the estuary (point CV in Fig. 1), thus at this point the estuary may be defined as highly stratified.

Observations

In Fig. 1 the observation net during the years 1972-1974 is given. The water level was measured at points HZ (Tvärminne), HS (Skogby). HT (Tammisaari) and HE (Ekerö) using punched tape recorders, model OTT. The interval of recordings was 5 minutes and the degree of resolution 1 mm. The measurements at point HD were made by an analog recorder.

Measurements of the current velocity were made by magnetic tape recorders, models PLESSEY and AANDERAA. In addition to the current velocity, the current direction and temperature of the water were also recorded with these instruments. In

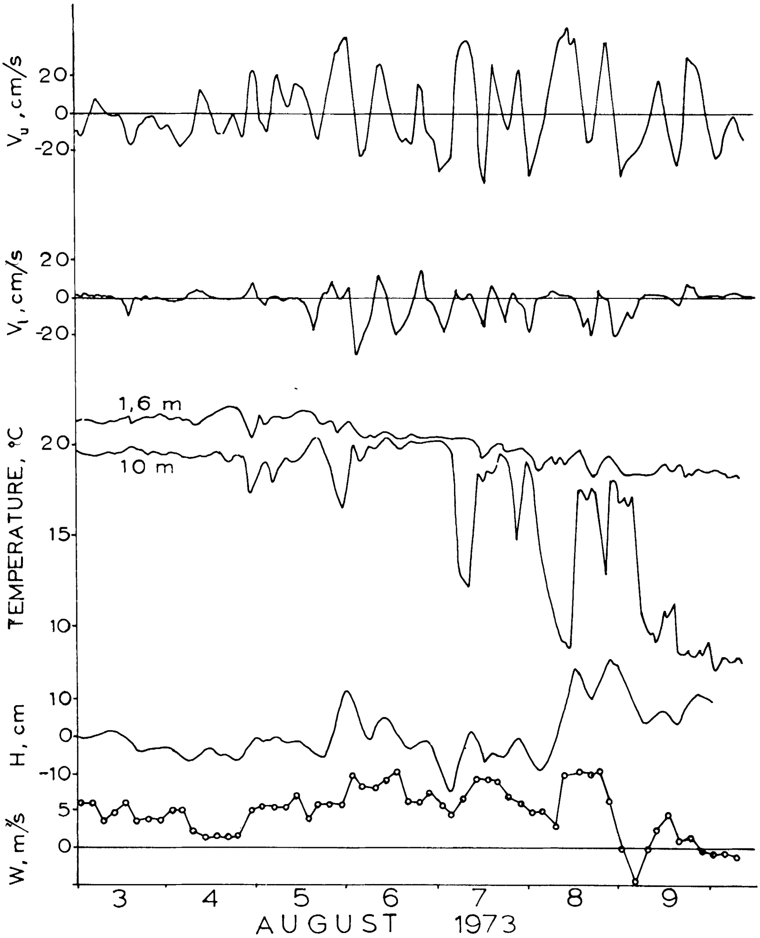

Fig. 3. An example of the hourly mean values of flow velocities at point CV, and water level *H* at HZ. The component of the wind velocity *W* along the direction of Pohjanpitäjänlahti measured at Russarö island is also given. The positive direction of the wind velocity is to the NE.

1973 measurements were carried out at point CV at depths of 1.6 m and 10 m. Measurements in 1974 were carried out at point CB. The interval between measurements was 5 or 10 minutes.

To measure the fluctuations of the internal surface, records of the water temperature at 10 depths with the interval of 40 cm were taken at point TK during summer 1973 and later also at point TS. The time interval between recordings was 5 or 10 minutes, and the instrument was the datalogger model AANDERAA.

In addition to the observations mentioned above, the salinity and temperature of the estuary were measured from two to four times a month. During the summers of 1973 and 1974 the instrument used was the temperature-conductivity-salinity probe model BECKMAN. During summer 1972 the salinity was determined from water samples by titration. The points of measurement are numbered by I-9 in Fig. 1.

The wind velocity used in this study was measured in the island of Russarö by the Finnish Meteorological Institute.

Examples of the hourly values of some quantities can be seen in Fig. 3. The variations in water level and velocity in this case are mainly due to wind and tidal effects.

Transport Equations

When fresh river water flows into an estuary, it occupies a position above the denser sea water. Due to disturbances such as wind stress, mixing takes place at the interface of the two water masses and salt water is entrained into the more turbulent fresh water. From this it follows that salt water must flow from the sea to the estuary to replace the entrained salt water. The importance of wind in this process has been illustrated recently by Gade (1970), Farmer (1974) and Göransson and Svensson (1975), although the mechanism has been clear for at long time (see Gade 1970).

Since the outlet of the estuary may be considered as highly stratified, the flows in the upper and lower layers may be handled separatly. The measurement section was a narrow sound (point CV in Fig. 1), 150 m wide. Because the sound is narrow the dependence of flow variables in the lateral direction is not taken into account; only the vertical variation is considered. The effect of lateral variation comes into consideration when calibrating the model.

Using the simplification and modification of Fischer's (1972) notation the variation of salinity s and velocity v may be divided into three parts, and the cross-sectional area into two parts. The equations for the upper layer become

$$s_u = s_{0u} + s_{1u} + s'_u \ , \qquad v_u = v_{0u} + v_{1u} + v'_u \ , \qquad A_u = A_{0u} + A'_u$$

Similar equations may also be written for the lower layer, in which case the subscript u must be replaced by ℓ.

The symbols have the following meaning:

s_{0u}, $s_{0\ell}$, v_{0u}, $v_{0\ell}$ time and cross-sectional average of salinity s and velocity v in the upper (u) and lower (ℓ) layer.

s_{1u}, $s_{1\ell}$, v_{1u}, $v_{1\ell}$ cross-sectional average of salinity and velocity fluctuations.

s'_u, s'_ℓ, v'_u, v'_ℓ depth dependent part of salinity and velocity. Also random fluctuations are included into this term.

A_{0u}, $A_{0\ell}$ time averages of the cross-sections of the upper and lower layer

A'_u, A'_ℓ fluctations of cross-sections

The time average of the volume flow Q is obtained from

$$<Q> = <Q>_u + <Q>_\ell = (v_0 A_0)_u + <v_1 A'>_u + (v_0 A_0)_\ell + <v_1 A'>_\ell \qquad (1)$$

Brackets (< >) mean averaging in time.
The time average of salt transport Q_S is

$$\begin{aligned} <Q_s> = <Q_s>_u + <Q_s>_\ell = & \underset{1}{(s_0 v_0 A_0)_u} + \underset{2}{(A_0 <s_1 v_1>)_u} + \underset{3}{(v_0 <s_1 A'>)_u} + \\ & + \underset{4}{(s_0 <v_1 A'>)_u} + \underset{5}{<s_1 v_1 A'>_u} + \underset{6}{(A_0 <\overline{s'v'}>)_u} + \underset{7}{<A'\overline{s'v'}>_u} + \\ & + \underset{11}{(s_0 v_0 A_0)_\ell} + \underset{12}{(A_0 <s_1 v_1>)_\ell} + \underset{13}{(v_0 <s_1 A'>)_\ell} + \underset{14}{(s_0 <v_1 A'>)_\ell} + \\ & + \underset{15}{<s_1 v_1 A'>_\ell} + \underset{16}{(A_0 <\overline{s'v'}>)_\ell} + \underset{17}{<A'\overline{s'v'}>_\ell} \end{aligned} \qquad (2)$$

A bar (—) over a symbol means cross-sectional average.
The computation of all terms in Eqs. (1) and (2) demands a long unbroken series of observations concerning current velocity, salinity and internal surface. In this case it is not possible, and in any case it is unnecessary, because some of the terms are so small that their estimated magnitude renders them negligible.

The following assumptions are made for further computations:

1. The average velocities in the upper layer and lower layer are proportional to the velocities V_u and V_ℓ measured at one depth in each layer, or

$$v_{0u} = k_u V_{0u}, \quad v_{1u} = k_u V'_u, \quad v_{0\ell} = k_\ell V_{0\ell}, \quad v_{1\ell} = k_\ell V'_\ell$$

   V_{0u} and $V_{0\ell}$ are the means and V'_u and V'_ℓ are the fluctuations of the measured current velocities.
2. Fluctuation of the depth of a proper isotherm presents the fluctuation of the depth of the internal surface between upper and lower layer. It is easy to compute this from temperature recordings.

A two-day averaging period will be used for Eqs. (1) and (2). The two-day period is chosen, firstly because the important variable wind often has a daily variation and secondly because this period is sufficiently long to yield data for statistical analysis.

This period is short enough to mean that variations in meteorological conditions during the period are slight.

The interval of recordings was 5 minutes. Thus the effect of shorter period fluctuations on the transport of salt cannot be taken into account. Their effect is, however, known to be small compared to other mechanisms (see Fischer 1972).

Terms (6), (7), (16) and (17)

Terms (6) and (16) represent the effect of the momentary linear correlation between salinity and velocity. From Fig. 4 it may be assumed that this correlation is effective in the upper layer, but poor in the lower layer. For this reason term (16) will be disregarded. To obtain the magnitude of term (6) a linear dependence is assumed for the fluctuations of salinity and velocity in the form presented in Fig. 4, or

$$s'_u = \frac{1}{h}\,(s_{0u}+s_{1u}-s_{\ell})\left(\frac{h}{2}-z\right)+r_s$$

$$v'_u = \frac{2}{h}\,(v_{0u}+v_{1u}-v_{\ell})\left(\frac{h}{2}-z\right)+r_v$$

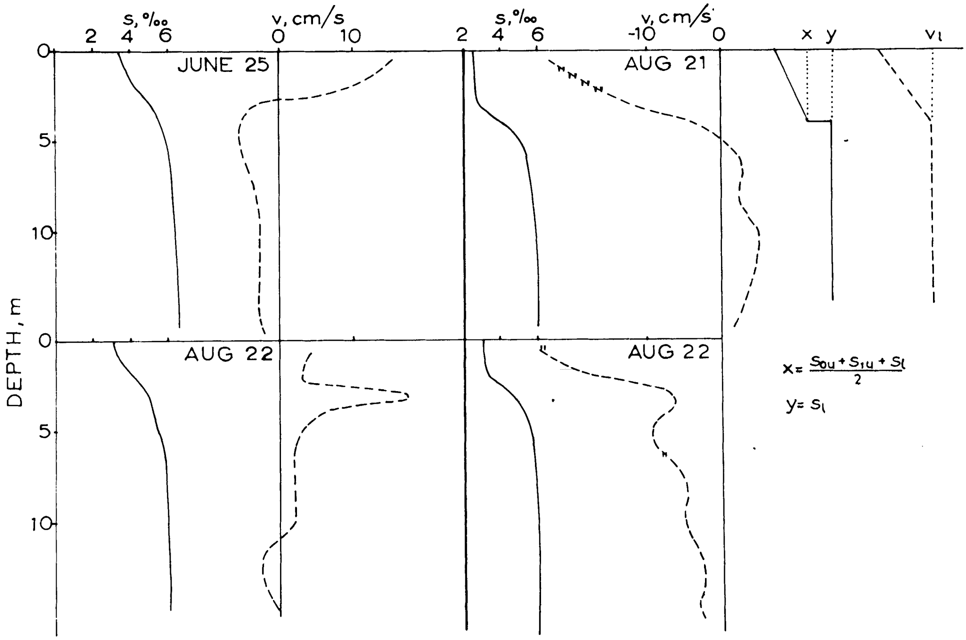

Fig. 4. Examples of distributions of salinity and velocity at point TK (1973). The diagram on the right represents the assumed velocity and salinity profiles. Positive velocity means inflow.

z is the depth, positive downwards and h is the thickness of the upper layer. r means the uncorrelated random part. When forming the time and depth averages, and taking only the greatest term, the following formula is obtained

$$\overline{\langle s'v'\rangle}_u = -s_{0u}\,v_{0u}\,\frac{s_{0\ell}-s_{0u}}{6\,s_{0u}} \tag{3}$$

As a conclusion term (6) is included when term (1) is multiplied by $1-(s_{0\ell}-s_{0u})/6s_{0u}$ The value of this coefficient was observed to vary between 0.90 and 0.99.

In the following paragraph it will be shown that the terms due to the variation of the cross-sectional area are small compared with the important terms. For this reason terms (7) and (17) may be disregarded.

A Comparison of Terms (1)-(5) and (11)-(15)

Terms (1) and (11) may be estimated directly from the measurements. Other terms will be either estimated indirectly or may be disregarded as small. The estimation of the order of magnitude of these terms is based on the assumption that the longitudinal transport of salt and heat takes place analogically. This kind of assumption may be made, because during the period of analysis there is no strong warming or cooling of the water, and the daily variation of temperature due to meteorological factors is small. The longitudinal temperature change is made greater by the fact that shallow Kaupunginselkä bay warms more quickly than the adjacent sea. Discussion about this assumption will be given at the end of this article. Based on the above assumption the following transformation between temperature and salinity fluctuations may be stated

$$s_{1u} = T_{1u}\,\frac{\Delta S_L}{\Delta T_L}\ ,\qquad s_{1\ell} = T_{1\ell}\,\frac{\Delta S_L}{\Delta T_L} \tag{4}$$

T_{1u} and $T_{1\ell}$ are the fluctuations of temperature at upper and lower layers. ΔS_L and ΔT_L are the longitudinal changes in salinity and temperature respectively between points TKand CV.

The transport of heat may be computed using the measured temperature and then returning to the transport of salt using the above transformation.

In the following computation the average of a layer has been replaced by the velocity measured at one level in both layers (assumption 1). Because the coefficients k_u and k_ℓ do not deviate much from unity, this does not carry a great error in the study of magnitude.

Fluctuations of cross-section will be estimated from equations

$$A'_u = B\,H'_i\ ,\qquad A'_\ell = -B\,H'_i \tag{5}$$

where B is the channel breadth at the interface, assumed to be constant, and H'_i is the momentary fluctuation of a proper isotherm (assumption 2). Only a few two-day periods could be used for the computation, because there were difficulties in obtaining simultaneous results with every instrument.

Table 2 shows terms (1)-(5) and (11)-(15) of Eq. (2) Although there are only a few periods, some conclusions can be drawn about the magnitude of the terms. Terms (1) and (11) are greatest and terms (2) and (4) smaller but still significant. Other terms are negligible. Table 2 gives also the component of the wind velocity along the direction of Pohjanpitäjänlahti Bay. As it may be seen, the change of wind direction causes a change both on term (1) and term (2) (see 25 April).

Table 2 - Transport of salt (kg/s) at cross-section CV in 1973. Terms (2)-(5) and (12)-(15) have been estimated from fluctuations in temperature. Positive sign means inflow. Also the component of wind velocity W (m/s) along the direction of Pohjanpitäjänlahti Bay positive to NE is given.

	$s_0v_0A_0$	$A_0\langle s_1v_1\rangle$	$v_0\langle s_1A'\rangle$	$s_0\langle v_1A'\rangle$	$\langle s_1v_1A'\rangle$	W
Upper	(1)	(2)	(3)	(4)	(5)	
20-21 Apr	-197	6.9	1.0	-1.7	-0.3	-3.8
22-23	-218	2.3	0.4	-1.2	0.0	-3.4
24-25	-206	2.9	0.1	-4.6	0.0	-4.5
26-27	22	17.4	0.6	-13.3	-2.0	3.5
10-11 May	-79	12,7	7.1	-5.2	-0.1	1.6
12-13	10	12.7	0.1	-6.9	3.5	3.9
14-15	-17	16.2	1.4	-2.3	-6.9	6.5
Mean	-99	10.1	1.5	-5.0	-0.8	
Lower	(11)	(12)	(13)	(14)	(15)	
10-11 May	90	0.4	0.0	2.0	0.0	
12-13	77	0.6	0.0	2.3	0.0	
14-15	93	-0.4	-0.1	-1.9	0.1	
Mean	87	0.2	0.0	0.8	0.0	

Term (2)

Fluxes of salt and heat in the upper layer with the fluctuating part of flow are:

$$F_{su} = \rho\langle s_1 v_1\rangle_u = -\rho D_s \frac{\partial s_{0u}}{\partial x} \qquad (6a)$$

$$F_{hu} = \rho c\langle T_1 v_1\rangle_u = -\rho c D_h \frac{\partial T_{0u}}{\partial x} \qquad (6b)$$

Here ρ and c are the density and specific heat of water. D_s and D_h are the dispersion coefficient of salt and heat. If the transport of salt and heat takes place analogically, then $D_s = D_h = D$. Discussion about the effect of this assumption will be given at the end of this article.

The dispersion coefficient will be computed from Eq. (6b). The temperature difference between the outgoing and incoming water will be used to compute a value for the temperature change rather than measuring it over a small distance. This method has several advantages: small differences in temperature can be measured with the same meter; all measurements of current can be used even if other measuring instruments were not functioning; any possible lateral currents have no effect on the coefficient of eddy diffusivity, which seems to afffect the results measured at point TK.

The length ΔX needed in Eq. (6b) is obtained from the equation

$$\Delta X = \Delta X_L \frac{\Delta T_0}{\Delta T_L}$$

where ΔX_L is the distance between points CV and TK, ΔT_0 is the average difference between outgoing and incoming water and ΔT_L is the corresponding temperature difference between points TK and CV. For the distance ΔX a value of 1300 m was computed. This value is slightly less than the length of the outlet, which is 1600 m.

There were 25 such two-day periods during which the temperature difference ΔT was greater than 0.2 C°. These periods were distributed in 1973 as follows: March-April 13 periods, May-June 4 and August 8. For the dispersion coefficient a mean value of $D =$ 62 m^2/s was obtained. The standard deviation of D was 36 m^2/s, and the standard deviation of the mean 7 m^2/s. The correlation coefficient between dispersion coefficient and the standard deviation of current velocity was 0.83. This dependence will not be used, however, because the effect of term (2) is not very important.

The dispersion coefficient obtained above was corrected by comparing the measured and computed salt contents of Pohjanpitäjänlahti. The corrected value for the coefficient is $D = 74$ m^2/s. This value naturally includes some of the systematic errors of the model.

Estimating the Cross-Sectional Areas A_{0u} and $A_{0\ell}$

In Eqs. (1) and (2) the cross-sectional areas of the upper and lower layers A_{0u} and $A_{0\ell}$ are needed. It is possible to determine these areas from the cross-sectional profile of the measuring section and salinity soundings. However, it is better to determine these areas from the water budget, because at the same time it is possible to take into account the fact that the current velocity measured at one level is not equal to the mean velocity in the cross-section. The cross-sections determined from the water budget may be considered as »effective«; they include the coefficients k_u and k_ℓ mentioned in assumption 1.

The water budget equation may be written in the form

$$\langle Q_f \rangle + \langle Q_u \rangle + \langle Q_\ell \rangle = A_p \frac{\Delta H}{\Delta t} \qquad (7)$$

where Q_f is the fresh water inflow, Q_u and Q_ℓ the inflow at the upper and lower layers, A_p is the total surface area of the bay sytem and H is the water level. With the aid of velocities this equation may be written in the form

$$\frac{\Delta H}{\Delta t} = \frac{A_{0u}}{A_p} V_{0u} + \frac{A_{0\ell}}{A_p} V_{0\ell} + \frac{f}{A_p} + \frac{\langle Q_f \rangle}{A_p} \qquad (8)$$

where V_{0u} and $V_{0\ell}$ are the measured mean velocities in the upper and lower layers. f is a term derived from the fluctuations of the cross-sectional areas and corresponds to the second and fourth terms on the right-hand side of Eq. (1). This term may be considered as independent of velocities V_{0u} and $V_{0\ell}$ and thus included in the random error of Eq. (8). Because the fluctuation of fresh water inflow is small over a short period, the last term in Eq. (8) may be considered as constant.

Coefficients A_{0u}/A_p and $A_{0\ell}/A_p$ may be determined by the method of least squares from the measurement data. In this analysis, ΔH, the change in water level over one hour at point HT or HE was the dependent variable, an V_{0u} and $V_{0\ell}$, the measured hourly velocities at point CV were independent variables.

The computations were carried out over two periods of time, 31 March - 17 April and 1 August - 21 August 1973. The first period consisted of 426 and the second of 489 hourly observations. Table 3 presents the results.

Table 3 - The determination of the relative cross-sectional areas A_{0u}/A_p and $A_{0\ell}/A_p$ from hourly data. R is the correlation coefficient and s the standard deviation of residuals.

Period	A_{0u}/A_p	$A_{0\ell}/A_p$	R	s cm/h
31 March - 17 April	$0.141 \cdot 10^{-4}$	$0.341 \cdot 10^{-4}$	0.978	0.52
1 - 21 Aug	0.211	0.372	0.898	0.76

It can be seen that the area $A_{0\ell}$ does not change much, but that the area A_{0u} is much greater during the second period. One possible reason for this change is that the current meter may have been at a slightly different depth and the results of current velocities could have been lower than during the first period. The results of measurements with the lower current meter are not so sensitive to changes of depth, as may be concluded from Fig. 4. Strong winds which occurred during four days in the second period may also have given rise to different results. In computing the budgets the cross-sectional area A_{0u} is not so important as the cross-sectional area $A_{0\ell}$, because the water budget calculation gives directly the quantity $A_{0u}V_{0\ell}$.

The following constant values for the cross-sectional areas based on the values in Table 3 have been accepted:

$$A_{0u} = 510 \text{ m}^2 \quad , \quad A_{0\ell} = 1\ 280 \text{ m}^2 \ ;$$

At the point where the current velocity is measured the area of the cross-section corresponding to the mean water level is 1,600 m^2. The area of the cross-section of the upper layer corresponding to a thickness of 3 m is 400 m^2. These values agree well with those given above.

Since the areas of the cross-sections may vary seasonally and with different wind situations the above computations were carried out for 16 two-day periods. It was observed that the areas varied quite a lot, and there was some evidence that A_{0u} may be greater during heavy southerly winds. A test using a wind-dependent coefficient A_{0u} did not, however, improve the computation model of the salt content of Pohjanpitäjänlahti. For this reason the latter computations were made using the constant values of the cross-sectional areas given above. Obviously the depth of the internal surface at the point of measurement is mainly determined by the depth of the Pohjanpitäjänlahti sill, which is about 5 m.

Determining the Velocities V_{0u} and $V_{0\ell}$ from Water Level Observations

In order to compute velocities V_{0u} and $V_{0\ell}$ two equations can be used. One is the water budget Eq. (8) and the other is the dynamic equation. At the moment all other quantities are known except the velocities V_{0u} and $V_{0\ell}$ and the term f. The inflow of fresh water Q_f is obtained from the equation

$$Q_f = (P-E)\,A_p + Q_r \tag{9}$$

where P is the precipitation of the estuary, E evaporation, A_p the surface area of the estuary and Q_r the river runoff. P was computed as a weighted mean taken from two precipitation stations of the Meteorological Institute (Tvärminne, Fiskars). For the estimation of evaporation values, those of Kuusisto (1975) for a lake situated in the south-western part of Finland were used. This method was considered to be accurate enough, because the absolute value of the difference of precipitation and evaporation is at most 10% of $Q_r/A_p \cdot Q_r$ was computed from the values of the Hydrological Office obtained at the Åkerfors power plant. The catchment area of this plant is 1,925 km^2 (Sirén 1955). The catchment area of the estuary at the measurement section is 2,460 km^2, when the estuary itself is excluded. Thus 28% was added to the runoff at Åkerfors to obtain Q_r.

The term f is the sum of the second and fourth terms of Eq. (1). The last, however, may be neglected as a small quantity, because term (14) in Table 2 is small. Thus knowing the velocity $V_{0\ell}$, the sum of terms (1) and (4) may be calculated using the salinity s_{0u} and the water budget equation.

The dynamic equation may be formed for the flow in both the upper and lower layers. However, the dynamic equation of the upper layer is more complicated because of the effect of wind, and this was not used. Since it is now a question of the mean values of the current velocities, which are rather small, the steady state equation was applied. This equation for the lower layer is

$$\frac{\partial H}{\partial x} - \varepsilon \frac{\partial h}{\partial x} = g\ (V_{0\ell}, V_{0u}) \qquad (10)$$

where H is the water level height, h the thickness of the upper layer, $\varepsilon = (\rho_\ell - \rho_u)/\rho_\ell$ where ρ_ℓ and ρ_u are the density of water in the lower and upper layers, respectively. When this equation is used for unsteady flow, the right-hand side also includes variances and covariances of velocities. The right-hand side of equation 10 may be determined from Bernoulli's equation (Kullenberg, B, 1954). It is believed, however, that in the present case it is better to determine function g empirically, because the variation of velocity is high during the averaging period, and the friction properties of the outlet are not known.

Results from 38 two-day periods were available; it was possible to use these to obtain the coefficient of a regression equation formed from Eq. (10). These periods were in April, August and September 1973. The dependent variable was the two-day mean $V_{0\ell}$; independent variables were water level difference ΔH between points HT and HS, two-day mean V_{0u}, $\varepsilon \Delta h$, which was computed from salinity soundings, the numeral I of the two-day period in question, and the component of the wind velocity along the estuary. ΔH was normalized so, that its value zero corresponds to the mean water level difference between points HT and HS. The value $I = 1$ corresponds to the period 22-23 March.

The most significant variables were ΔH and I. With these the total correlation coefficient 0.81 and the standard deviation of residuals 1.2 cm/s were computed.

The reason why the term $\varepsilon \Delta h$ is not a significant variable is that its variation was rather small. Its standard deviation was only 5% about the standard deviation of ΔH. Over such long periods, during which the wind blows from the same direction, the omission of this quantity may lead to errors. According to the salinity soundings the mean value of $\varepsilon \Delta h$ during the calibration period was not the same as its mean value over the whole period. It was estimated, that the difference corresponded to a value 0.23 cm/s in the velocity $V_{0\ell}$. This correction has been added into the final equation.

The effect of the numeral I was small. It was neglected in the final equation because it was considered that this quantity indicated the effect of different locations of the current meter during different periods ofmeasurement, and not the real variation of the equation with time.

The final equation for computing the velocity $V_{0\ell}$ is

$$V_{0\ell} = a\Delta H + b \qquad (11)$$

where a = - 3.24 $1/s$ and b = 1.09 cm/s.
The value of b is affected by the mean inclination of the surface of equal density as well as the selection of the zero point of ΔH.

Fig. 5 presents the dependence of velocity $V_{0\ell}$ on the water level difference ΔH.

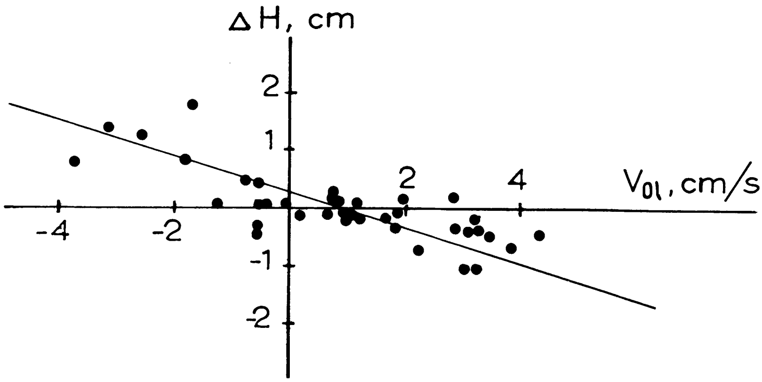

Fig. 5. The velocity $V_{0\ell}$ plotted against the water level difference ΔH at measuring points HT and HS. Each point marks the mean value of two days.

The Components of Salt and Water Budgets

The method for correcting the computations will now be presented, followed by the corrected terms of the water budget.

There are several assumptions in the method presented here for computing the salt and water budgets. For this reason it is important to compare the measured and computed salt contents of the estuary. The following equation between the transport of salt and salt content in the estuary can be stated:

$$M_{sn} = \sum_{i}^{n} (Q_{s0n} + Q_{s1u} + Q_{s0\ell})_i \cdot \Delta t + M_{s0} \tag{12}$$

where M_{sn} is the amount of salt at the end of time period $n.Q_{s0u}$ is the transport of salt at the upper layer with the mean flow (or terms (1) and (4) in Eq. (2)), $Q_{s0\ell}$ is the transport of the mean flow in the lower layer (term (11)) and Q_{s1u} is the transport of the fluctuating flow in the upper layer (term (2)).

The salt content of the estuary can also be determined from salinity soundings. Because it is not possible to compute the absolute values of the salt content from Eq.

(12) since M_{S0} cannot be computed, the comparison between measured and computed salt contents was made using the deviations from the mean value of each measuring period.

It appeared from this comparison that there was some systematic source of error which made measured and computed salt contents deviate more from each other during the summer. The source of error was believed to be in the determination of the fluctuating part of the transport equation, because term (2) was perhaps estimated

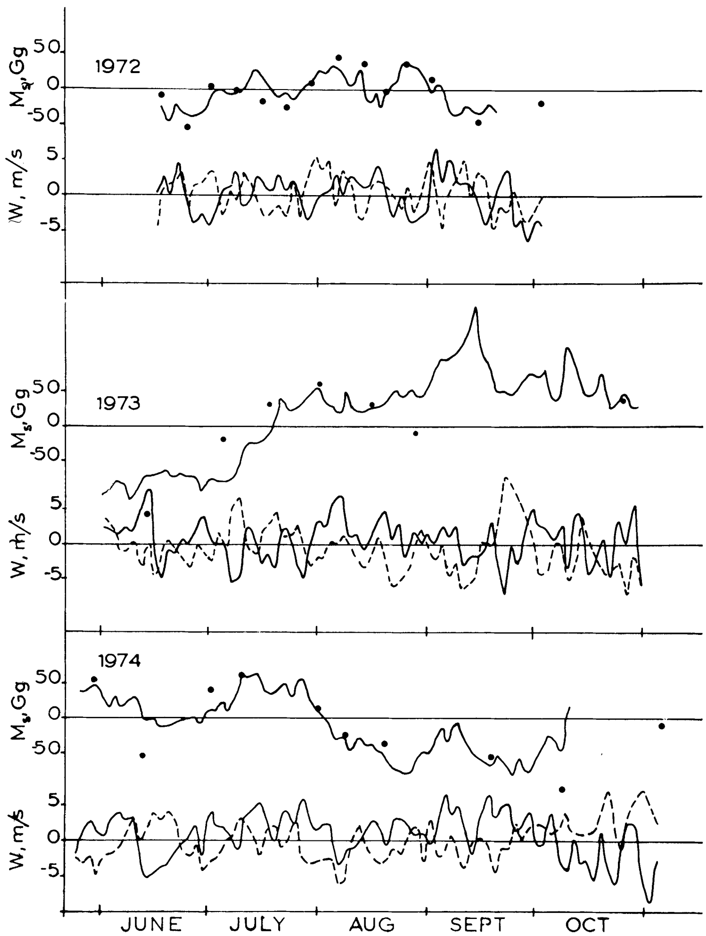

Fig. 6. The computed (heavy line) and measured (points) deviation of the salt content M_S from the mean value. The wind velocity W in the direction of Pohjanpitäjänlahti (positive to NE, heavy line) and perpendicular to this direction (positive to NW, broken line) is also presented.

with least accuracy. Trials showed that the best correlation between measured and computed salt contents was obtained with a value of 74 m^2/s for the dispersion coefficient instead of 63 m^2/s., the latter having been determined from temperature fluctuations. The reason for this change in dispersion coefficient may come from its approximate determination or from the fact that some terms in the transport equation were neglected.

Fig. 6 presents the two-day means of the computed and measured salt contents of Pohjanpitäjänlahti. It can be observed that the measured values have the same general variation as the computed values. A correlation coefficient of 0.78 was calculated between measured and computed salt contents. The standard deviation of the measured salt content was $46 \cdot 10^6$ kg and that of the computed salt content was $39 \cdot 10^6$ kg. The agreement between measured and computed values may be considered fair, because the measured values may also have measurement errors. The standard deviation of salt content is only about 10% of the total salt content. Some errors also derive from the lack of knowledge concerning daily salinity variations at the outlet of the estuary.

Fig. 6 also gives the wind velocity in the direction of and perpendicular to the direction of Pohjanpitäjänlahti. It can be seen that in general, an increase in salt content occurs when the wind blows from the NE and a decrease in salt content occurs when the wind blows from the SWS.

Table 4 - Components of the water budget (m^3/s) and salt budget (kg/s) during different months. A positive sign denotes inflow to Pohjanpitäjänlahti.

	Q_f	Q_{0u}	$Q_{0\ell}$	Q_{s0u}	$Q_{s0\ell}$	Q_{s1u}
1972						
July	14	-37	22	-140	120	40
Aug	17	-38	22	-160	120	30
Sept[1]	27	-49	21	-170	120	30
1973						
May	22	-34	10	-140	60	20
June	12	-11	2	-30	20	10
July	11	-41	33	-170	200	20
Aug	12	-25	10	-110	70	40
Sept	10	-45	34	-240	220	30
Oct	10	-31	23	-170	150	10
1974						
June	18	-48	33	-180	150	30
July	20	-34	16	-120	90	20
Aug	20	-38	13	-110	80	10
Sept	14	-19	7	-60	40	20
Mean	16	-35	19	-140	110	20

[1] 20 days

The great effect of wind on transport processes also becomes clear from the fact that the correlation coefficient between wind velocity and the slope of the water level is quite high. Using two-day means, values of 0.78-0.88 were computed for this coefficient for different years.

Table 4 gives monthly means of the component of the water and salt budgets. The following symbols are used:

Q_f	inflow of fresh water
Q_{0u}	inflow at upper layer
$Q_{0\ell}$	inflow at lower layer
Q_{s0u}	inflow of salt at upper layer with the mean flow
$Q_{s0\ell}$	inflow of salt at lower layer with the mean flow
Q_{s1u}	inflow of salt at upper layer with the fluctuating flow

Summary and Conclusions

The basic problem was the determination of the water and salt budgets of Pohjanpitäjänlahti. For this purpose a model was formulated, the parameters of which were determined from observation data from 1973. The flow at the outlet was approximated as being two layered. The two-day means of the velocity at the lower layer were determined from the slope of the water level, and the flow at the upper layer was computed from the water budget of the estuary. The inflow of salt with the mean flow was determined from the measured salinity and computed velocity. The transport of salt with the fluctuating flow was computed with the dispersion coefficient and the longitudinal change of salinity. The dispersion coefficient was determined from temperature fluctuations and was corrected by comparing the measured and computed salt contents of Pohjanpitäjänlahti. The model is applicable only during the summer and autumn when stratification is stronger. The following conclusions can be drawn from the results of this study:

1. The method used seems to be applicable to an estuary of the Pohjanpitäjänlahti type. Difficulties arise due to the low inclination of the water level and the small mean velocities of the flow.
2. The role of the mean flow in the transport processes of salt is greater than that of the fluctuating flow. The inflow of salt in the latter case accounts on average, for 20% of the inflow of salt with the mean flow in the lower layer.
3. During the observation period the mean inflow of salt water in the lower layer was 19 m^3/s. This corresponds in a month to about 20% of the volume of Pohjanpitäjänlahti. The outflow in the upper layer is, on average, about twice the fresh water inflow.

Discussion on the Results and the Method of Computation

If it were possible it would be quite convenient to use temperature for determining mixing parameters especially in fresh and low salinity waters where recording electric conductivity is difficult. Earlier temperature was used at least to determine the vertical dispersion coefficient; now it is used to determine the horizontal dispersion coefficient.

The distribution of salinity is a result of longitudinal and vertical mixing and advection in the water body. The distribution of temperature is also influenced by heat flow through the water surface. The transport processes of heat and salt may also deviate in that dispersion coefficients may be different. For example Gade (1970) computed that the ratio of the vertical dispersion coefficient of heat to that of salt ranges from 1.3 to 5.5 for Oslofjord.

In this study the temperature fluctuations for determining transport processes were used twice. Some of the values presented in Table 2 and the longitudinal dispersion coefficient were computed with temperature. The values in Table 2 were used only to study the magnitude of some of the terms.

The computed average value of the dispersion coefficient is 63 m^2/s. On comparing the measured and computed salt contents of Pohjanpitäjänlahti, the corrected coefficient was 74 m^2/s. The difference is not great, because the latter value is derived from a rest term, which is influenced by both systematic measurement errors and also neglecting some of the term of the transport equation. These factors justify the use of temperature for determining the parameters of the transport processes in the case in hand.

Another fact which also justifies the use of temperature is that the computations were carried out in quite different conditions. In March and April, when the vertical variation of temperature is small results were computed which were similar to those for August, when the vertical variation of temperature is higher. The following presents average values of the dispersion coefficient D, the standard deviation of the current velocity σ_v, the difference in temperature ΔT and difference in salinity Δs between the depths of 0.5 m and 10 m computed from the soundings and also the amount of two-day periods n, from which D and σ_v were computed.

Period in 1973	D m^2/s	σ_v m/s	ΔT °C	Δs ‰	n
March 26 - April 16	68	.15	1.0	-3.0	7
April 18 - 29	43	.10	2.5	-3.1	6
May 10 - June 6	75	.13	6.8	-2.8	4
August 1 - 20	67	.14	6.7	-2.5	8

As can be seen, the mean of coefficient D is rather stable in spite of the great variation in both weather conditions and vertical temperature distribution. The variations in the value of D are mainly explained by the variation in the standard deviation of velocity.

The reason why using temperature in this case seems to be successful is that the conditions at Pohjanpitäjänlahti are favourable. Because the water body is highly stratified, the vertical mixing of each layer is hindered, the shallow Kaupunginselkä bay warms up in spring more quickly than the adjacent sea, and temperature differences naturally result. Another positive factor is that during the computation periods the vertical heat flow through the surface was small. No strong heating had occurred by March or April and August presents a period during which the water temperature is at its maximum.

There are several assumptions which must be discussed. Because the velocity was measured only at one level at each layer, an error may result. However, this error is minimized when we consider the mean flow because the cross-sectional areas of the layers include this error. When we consider the fluctuating part of transport, the error is included in the corrected dispersion coefficient D.

The method of computation is not valid in every situation. During heavy southwesterly winds the cross-sectional area of the upper layer may rise, and correspondingly the crosssectional area of the lower layer may decrease. This is clear in the example in Fig. 3 where the warm upper layer temporarily reaches a depth of 10 m. As a result, inflow at the upper layer is estimated to be too low during this period. When the salt content of Pohjanpitäjänlahti changes a great deal during long periods erroneus values are computed for flows in the upper and lower layers because the term in Eq. (10), due to density stratification, was not taken into account. The computation is not valid in winter either, when friction due to ice affects the flows, or when the density stratification deviates greatly from that during the calibration period.

In the following some systematic sources of error will be taken into consideration. The most important part of the computation is the determination of velocity $v_{0\ell}$ According to the results of regression analysis carried out with 38 observations, the standard deviation of the mean of $v_{0\ell}$ is 0.19 cm/s. This corresponds to the value 2.5 m^3/s of the standard deviation of the mean of $Q_{0\ell}$. Because Q_{0u} was determined with the water budget, its error must be the same, but with the opposite sign, as the error of the term $Q_{0\ell}$. The corresponding error Δ_{s1} of Q_{s1u} is then

$$\Delta_{s1} = \Delta_0 (s_{0u} - s_{0\ell})$$

Using mean values for salinities, the standard deviation of the mean of Q_{s1u} becomes 4.5 kg/s. Of the other systematic errors, that of Q_f may be significant. The error of this term may be 5%, and causes an error of 2% in Q_{0u} which corresponds to error of 3 kg/s in Q_{s0u} and Q_{s1u}. From above the conclusion can be drawn that the total error of Q_{s1u}, when determined from water and salt budgets, may be almost 50%.

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