BOREAL ENVIRONMENT RESEARCH 4: 307–317 ISSN 1239-6095
Helsinki 15 December 1999 

# Water exchange and mixing in a semi-enclosed coastal basin (Pohja Bay)

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Stipa, T. 1999. Water exchange and mixing in a semi-enclosed coastal basin (Pohja Bay). *Boreal Env. Res.* 4: 307–317. ISSN 1239-6095

The processes contributing to the deep water renewal and oxygen balance of Pohja Bay are reviewed here, with reference to the extensive experimental material from the years 1990–1993. The observations reveal previously unreported frequent minor inflows of sea water into the bay. A one-dimensional diffusion model is introduced and applied to computation of the diffusion-corrected biological oxygen sink during apparent deep water stagnation periods; the sink is evaluated as $0.06 \pm 0.01$ mg $l^{-1}$ $d^{-1}$. The results are summarized in a conceptual description of the physical processes underlying the estuarine circulation and oxygen balance of the bay.

## Introduction

Pohja Bay is a semi-enclosed coastal sill basin in the northern Baltic Sea. It is one of only a few of its kind along the coasts of Finland, and thus offers an interesting platform for the study of the physical processes of coastal sill basins in its relatively rare salinity regime.

Furthermore, Pohja Bay (also called Pojovik or Pojo Bay in older literature) has a long history of serving as a location for scientific studies, as the first physical studies there date from the beginning of the century (Witting 1914).

Currently, the level of oxygen in the deep water of the bay, which tends towards zero, is a major environmental concern. Anoxic conditions in the deep water could cause a release of nutrients from the sediments, leading to further eutrophication.

The researchers of the Tvärminne Zoological Station of the University of Helsinki conducted a three-year monitoring program in Pohja Bay. During this program, biological data as well as hydrography (CTD), currents and water level were measured either continuously or during field excursions at 1 or 2 week intervals. I was offered the chance to use this extensive data to study the physical features of the bay, the results of which this work presents.

The aim of this study is to clarify the process of deep water renewal in the bay, and to provide a link between the physical and biological processes. In particular, the problems of the oxygen dynamics in the deep water are discussed in the light of physical mixing mechanisms.

## Study area

Pohja Bay is situated near the southern edge of the Finnish mainland, between northern latitudes

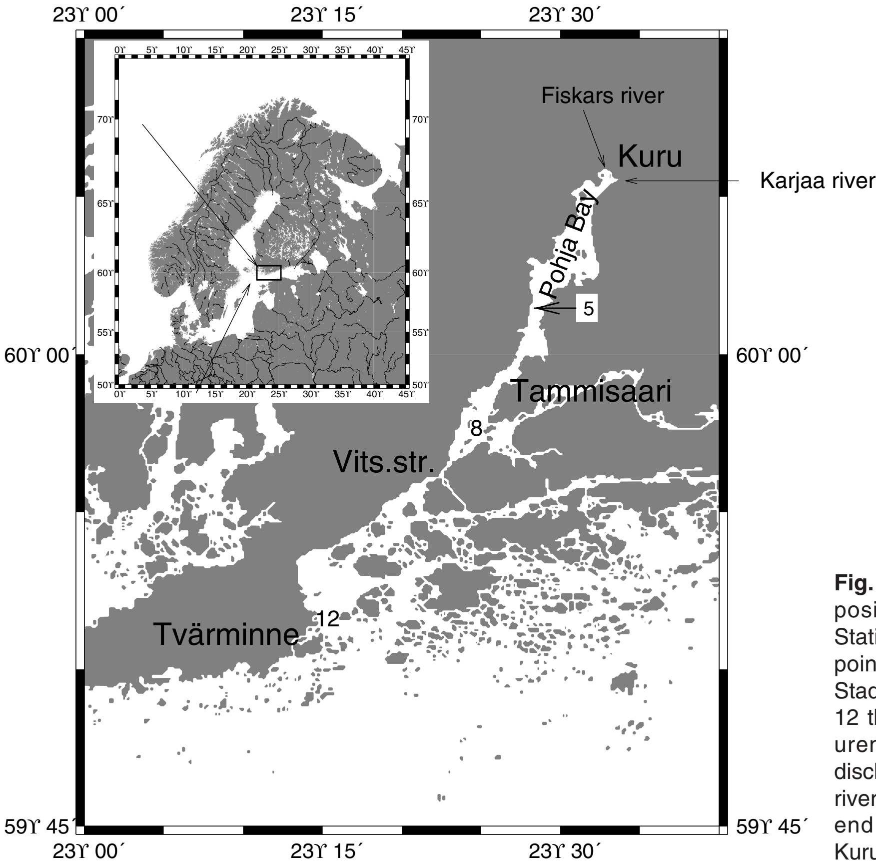


**Fig. 1**. The geographical position of Pohja Bay. Station 5 is the deepest point (Sällvik), station 8 Stadsfjärden, and station 12 the end of the measurement transect. The discharge points of major rivers are at the northern end of the Bay (near Kuru).

59°59´ and 60°06´ and eastern longitudes 23°26´ and 23°34´ (Fig. 1). The waters leaving the bay enter the Baltic Sea in a transition zone between the Gulf of Finland and the Baltic Proper.

The bay is narrow, with a width of 1–2 km compared to its length of 14 km. It is oriented approximately SW–NE, the direction of its main axis being 30°. Thus, lateral variations can be expected to be small.

The entrance from the sea to the bay consists of an irregular network of narrow passages between small islands. The topographic variability has its effects on the flow conditions via e.g. increased mixing. The most important sills are found at Vitsandsströmmen and the Tammisaari bridges, with maximum sill depths of 8 m. The relatively flat (mean depth around 3 m, deepest parts near the ship route at 5–6 m) area between these sills is called Stadsfjärden (indicated by station number 8 on Fig. 1) and is the immediate source of deep and saline water of Pohja Bay.

Pohja Bay gets its fresh water supply mostly from the Karjaanjoki drainage basin, which has a mean discharge of 17 $m^3 s^{-1}$ (Leppäjärvi 1992). The total drainage area of the bay, as computed from the registered drainage basins, is 2 280 $km^2$. The largest basins, Karjaa and Fiskars, discharge at the northernmost end of the bay (Kuru). Their drainage area is 96% of the total, i.e., to a very good approximation, all fresh water flowing into the bay enters it from the northern side. It seems reasonable to estimate the total fresh water discharge into the bay as proportional to the discharge from the river Karjaa, i.e. $Q_{\mathrm{total}} = 1.11 Q_{\mathrm{Karjaa}}$.

## Qualitative observations

A thorough qualitative description of the measurement program results can be found in (Stipa 1996); a summary of the most relevant and unexpected observations is given here.

The bay has a permanent density stratification, which is in summer mostly, and in winter

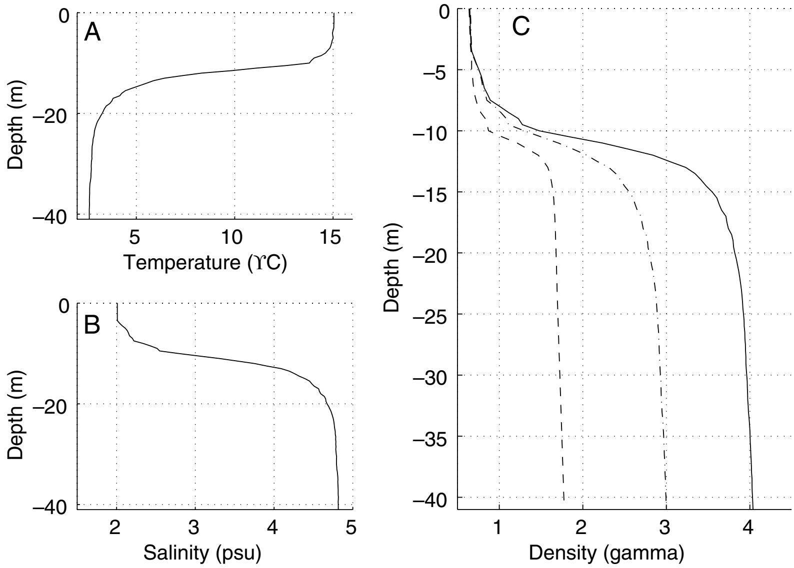


**Fig. 2**. Illustration of the dependence of density ($\gamma = \rho - 1\ 000$ kg m$^{-3}$) on (A) temperature (°C) and (B) salinity (psu). The solid line in C shows the density profile corresponding to A and B, the dashed line shows a hypothetical density profile if the salinity were a constant 2 psu, and the dash-dotted line shows a density profile with $s = s(z)$ and $T =$ 15 °C. The data for A and B were measured on 2 July 1991.

practically completely, caused by a salinity difference of 0.5–4 psu between the deep water and the mixed layer (Fig. 2; the unit psu is used here to stress that salinity values are given on the Practical Salinity Scale; cf. Fofonoff 1985; [psu] = 1).

The highest fresh water discharge into the bay during the observation period already occurred in spring 1990, when the discharge was higher than 50 m$^3$ s$^{-1}$ for several months (Fig. 3). This may be one reason for the alarmingly low oxygen concentrations in summer 1990 (Fig. 4). Mostly, however, the discharge remained below 40 m$^3$ s$^{-1}$ even during springtime, with the lowest values being generally around 8 m$^3$ s$^{-1}$.

The inflow events can best be identified by comparing successive profiles of salinity $s$ and temperature $T$ from the Sällvik deep (station 5), but also to some extent from isopleths in Fig. 5. In the early summer of 1990, the deep water was pushed up several times by inflows of denser water from the outside. Such events can be detected until mid-June (roughly day number 160; this study uses day numbers starting from 1 January 1990 as a unified time axis).

After a period of apparent stagnation, new inflow events could be observed at the end of October, followed by inflows of varying magnitude during the winter months, until the beginning of April 1991 (day ~460).

As perhaps the most unexpected observation of this study, a major inflow in July 1991 (day

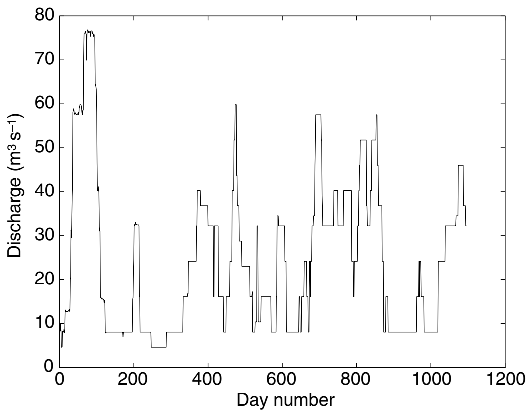


**Fig. 3**. Estimated fresh water discharge to Pohja Bay for the years 1990–1992.

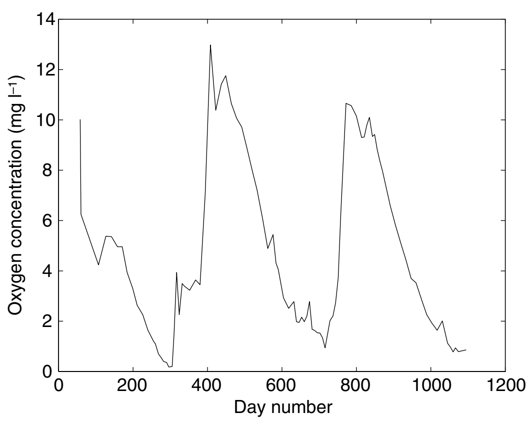


**Fig. 4**. Measured oxygen concentration at a depth of 35 m at station 5 (Sällvik).

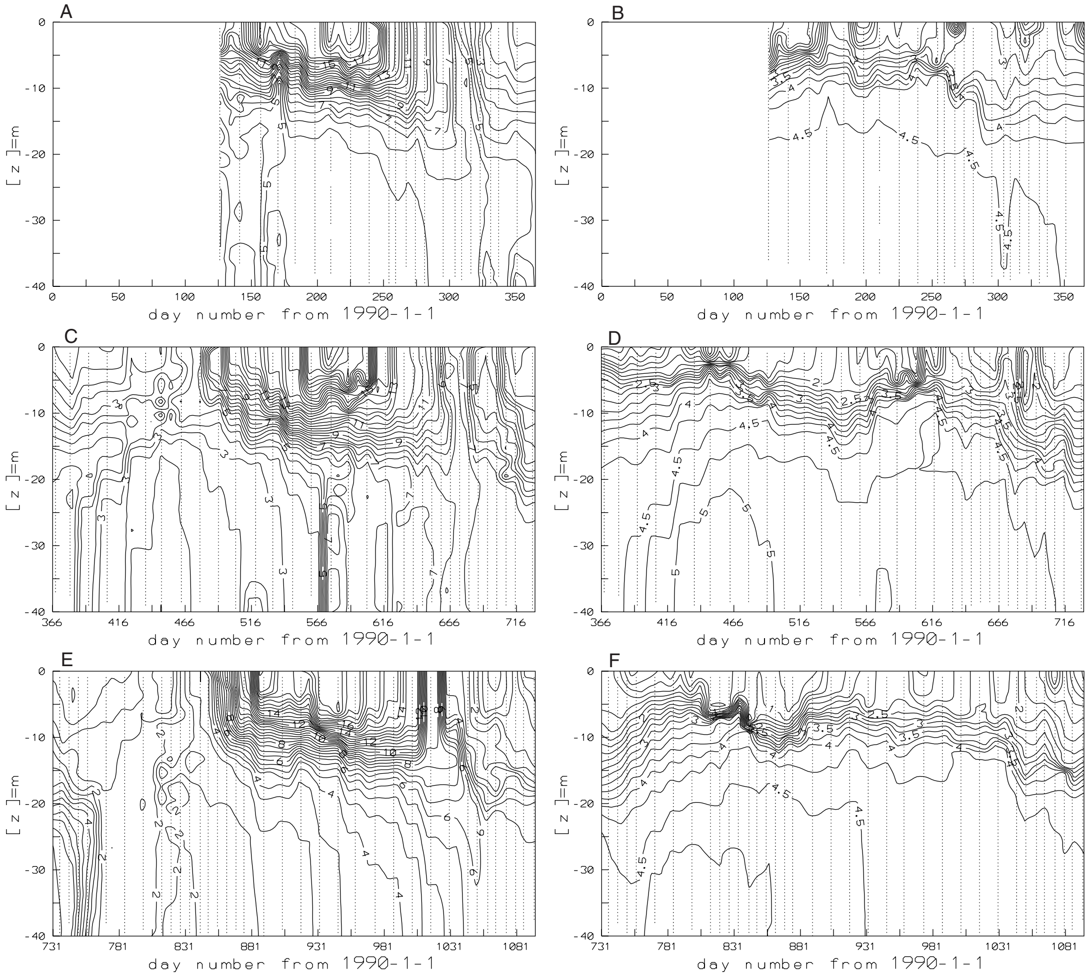

**Fig. 5**. Isopleths from station 5 (Sällvik) for temperature (A: 1990, C: 1991, E: 1992) and salinity (B: 1990, D: 1991, F: 1992). Individual measurement positions are shown as dots.

~570) renewed the bottom water of the bay below a depth of 20 m and pushed the older water up and towards the northern end of the bay (Figs. 5c and 6). The old water later mixed laterally in the bay. The sparsity of the data during this period does not allow for a thorough investigation of the reason for the inflow, but the external stratification seems to have played a role (Stipa 1996).

This event can hardly be detected in the oxygen concentration curve of Fig. 4. The reason is probably that the oxygen concentration in the new water was not essentially higher than in the old deep water, i.e. all inflows of sea water cannot be expected to increase the oxygen content of the deep waters in the bay.

After mid-September 1991 (day ~620), several minor inflows can again be detected in the data, but a major deep water renewal first took place in mid-January 1992 (day ~745). The deep water was flushed again several times during the winter. The inflow activity ceased by the end of March 1992 (day ~820). The first reliable signs of deep water renewal episodes can be seen at the end of October 1992 (day ~1 030), but a major renewal took place at the beginning of January 1993 (day ~1 100).

The annual cycle of temperature and salinity in Pohja Bay is only quasiperiodic, as interannual differences are appreciable. There was a shift towards lower salinity during the study period (Fig. 7). This

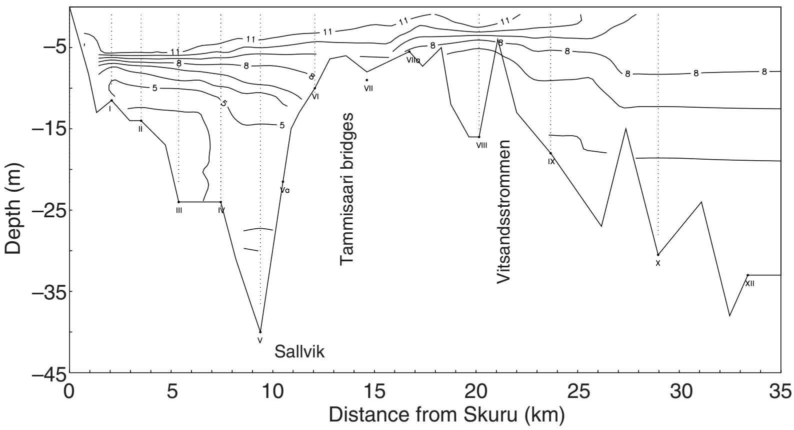


**Fig. 6**. Temperature section on 22 May 1990, starting at Kuru (Swedish name Skuru in the label). Roman numbers indicate stations of the transect (Fig. 1). The incomplete horizontal mixing after the inflow in the deep water is clearly visible.

will probably be a reflection of more general long term trends in the northern Baltic Sea.

Stagnation periods should be seen as periods of slowly increasing temperature and decreasing salinity (lines with a downward slope in Fig. 7). Such periods do not seem abundant in the bay (cf. Fig. 7), which is in disagreement with the hypothesis of a long period (100–200 d) of complete stagnation in the summer.

A comparison with the values measured by r/v *Aranda* in 1954 in Sällvik (Granqvist 1955), $T$ = 3.0 °C, $s$ = 4.94 psu and $O_2$ = 2.58 mg $l^{-1}$ at 36 m, does not seem to indicate any significant long-term change having taken place in the overall hydrography.

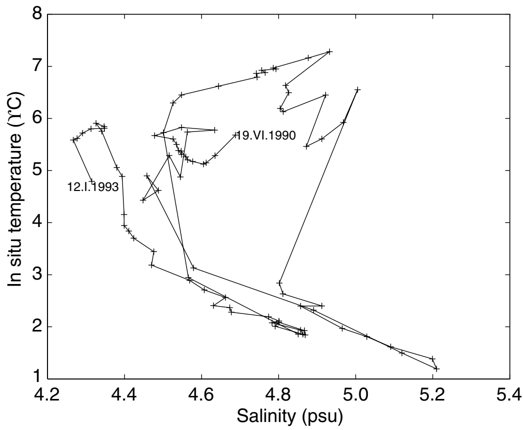


**Fig. 7**. A *T-s* plot for a depth of 35 m in Pohja Bay for the study period. The first and last days are indicated, as well as individual measurement points.

## Water exchange

The water exchange of a coastal bay can be caused by "external" or "internal" forcing. "External" forcing, or more exactly *barotropic* forcing, is caused by water level differences between the bay and the sea, by air pressure or by similar factors that are experienced in the same way by the whole water column. "Internal", or *baroclinic*, forcing is caused by differences in the internal structure of the water column in the bay and in the sea.

### Barotropic flow

The barotropic forcing discussed here is caused by sea level variations and volume flow from fresh water.

Various attempts were made to determine the barotropic water exchange from differences in water levels between the Hanko mareograph (59°49.4´N, 22°58.5´E, some 50 km south-west of Tammisaari) and the Tammisaari water level. The results, however, were very discouraging; apparently the channel from the sea to the final bay entrance is so long that its own dynamics make a direct comparison of water levels in the bay and in the sea unproductive.

The barotropic water exchange $Q_b$ must, for physical reasons, be proportional to the water level derivative $\partial h_i/\partial t$, i.e.,

$$Q_b = A\frac{\partial h_i}{\partial t} \tag{1}$$

Here the bay surface area $A$ is constant for small changes of water level.

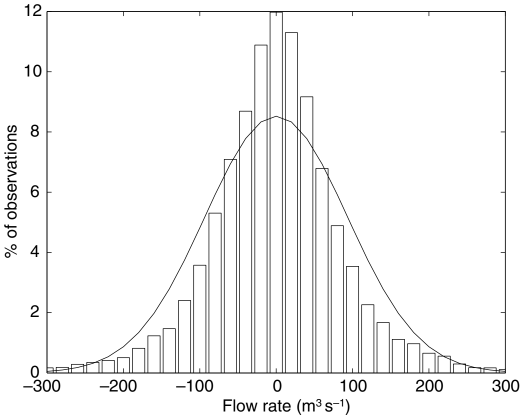


**Fig. 8**. The distribution of flow rates into Pohja Bay as determined from water level changes at Tammisaari. Positive values are towards the bay. The corresponding Gaussian distribution is also shown, with $\sigma = 93.6$ m$^3$ s$^{-1}$

The water level derivative at the Tammisaari bridges multiplied by the bay area is taken as the true barotropic water exchange. For the whole period of water level data available from Tammisaari (June 1990–December 1992), the mean absolute flow rate caused by sea level fluctuations, $\overline{|Q|}$, is 65 m$^3$ s$^{-1}$. With a representative cross-sectional area of 1 000 m$^2$, the corresponding mean current in the contraction would be 6.5 cm s$^{-1}$.

Launiainen (1972) determined the contribution of the tidal $M_2$ component as 27 m$^3$ s$^{-1}$. Since then a deeper ship route has been dredged under the Tammisaari bridges. This can be expected to decrease the choking effect at the control and thus increase water exchange. Hence, the tidal influence is at least 40% of the total barotropic water exchange of the bay. The Baltic Sea tides are not negligible in this case.

The distribution of flow rates in different classes is almost symmetric, and deviates from the Gaussian distribution by the excess of small and high flow rates (Fig. 8).

The flow rates (Fig. 8) do not take the fresh water discharge into the bay into account. In order to get the true flow rates, one must subtract the fresh water discharge (and evaporation) from the flow rates presented, i.e. the true barotropic volume flow is $Q_\mathrm{b} = A[(\partial h_i)/(\partial t)] + Q_\mathrm{f}$. This will, however, merely shift the mean of the distribution to about –20 m$^3$ s$^{-1}$.

## Baroclinic water exchange

The complexity of two-layered flow phenomena in non-ideal conditions and the lack of measurement data prevent us from determining a time series for the baroclinic water exchange into Pohja Bay. Therefore only some limiting values for extreme hydrodynamic situations are presented (*see* Fig. 9).

The transition from a two-layered, bidirectional flow (Fig. 9) to a single-layered, unidirectional flow takes place when (Armi and Farmer 1986, Stipa 1996)

$$\left|\frac{Q_\mathrm{b}}{B}\right| \geq 0.544 h_s \sqrt{g' h_s} \qquad (2)$$

where $h_s$ is the depth at the contraction, $g' = g(\rho_2 - \rho_1)/\rho$ the reduced gravity, $B$ the channel width and $Q_b$ the barotropic volume flow. Choosing $h_s = 8$ m, $g' = 0.02$ m s$^{-2}$ and $B = 100$ m, we obtain $Q_\mathrm{b} = \pm 174$ m$^3$ s$^{-1}$ as the barotropic forcing which corresponds to the blocking of one layer.

The corresponding flow rate for the case in which the interface retreats completely from the control section (“box flow”), is $Q_\mathrm{b} = \pm 320$ m$^3$ s$^{-1}$.

The fresh water discharge produces a net barotropic current of up to 60 m$^3$ s$^{-1}$, thus pushing the upper layer towards the sea and thereby increasing the barotropic flow towards the bay that is needed to stop the outflow completely. A water level rise of 3.4 cm h$^{-1}$ will cause the blocking of the out-flowing upper layer in this case. About 5% of hourly observations were unidirectional flows out of the bay and 2% unidirectional flows into the bay.

The flooding of the contraction was indeed observed in January 1993, with continuous CTD recordings from an 8 m depth at the Tammisaari bridges (Fig. 10). The rapid change in salinity seems to indicate that the layers were fairly distinct; the two-layer approximation can thus be expected to be satisfactory.

These calculations are exactly valid only in the extreme case of critical flow in the control

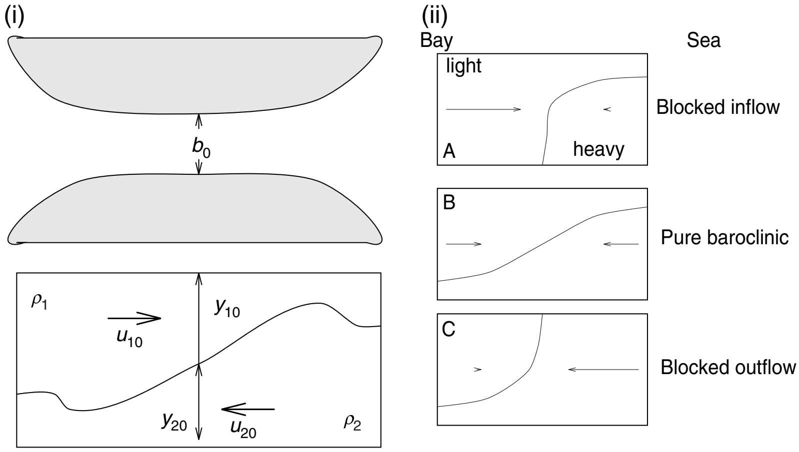


**Fig. 9**. (i): The schematic contraction of width $b_0$, top (upper plot) and side view (lower plot; *y*s are layer depths, *u*s layer velocities and $\rho$s layer densities). (ii): Interface positions for maximal two-way exchange as a function of external forcing. A and C illustrate the "box flow" situation, whereas in B there is no barotropic forcing. Arrows indicate the strength of the flow in each layer. Simplified and redrawn according to Armi and Farmer (1986).

section with no interfacial friction; in non-ideal conditions the limiting flows will be lower. Therefore, the values presented above should be considered as maximal values, the actual limits being in many cases essentially less and unidirectional flows therefore more frequent than estimated here.

## Vertical circulation

### A one-dimensional model for the basin

For many purposes, it is sufficient to consider the properties of the deep water in a basin in the vertical direction only. The approximation is good as long as the horizontal variations in the basin are negligible, and the effects of horizontal movements on the vertical stratification can be properly parameterized.

The most important horizontal processes acting in the bottom water are internal waves and diffusion. The internal waves contribute to the mixing by breaking either in the water column itself because of shear instability, or on sloping bottoms, and can thus be taken into account in a one-dimensional model with a turbulent diffusion coefficient (Gargett 1984, Stigebrandt and Aure 1989).

For vertical one-dimensional fluxes and a volume element with horizontal planar faces, the diffusion equation reads (Stigebrandt 1987)

$$\frac{\partial(\rho s)}{\partial t} = -\frac{1}{A}\frac{\partial}{\partial z}\left[A\omega\rho s + AF_{\mathrm{d}}\right] + \frac{Q_i}{Adz}\rho_i(s_i - s). \quad (3)$$

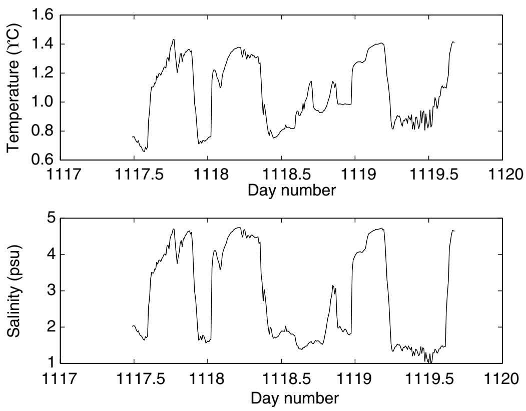


**Fig. 10**. Salinity and temperature measured with a moored CTD at a depth of 8 m at the Tammisaari bridges from January 21 1993 09:42 UTC to January 23 14:12 UTC

Here $Q_i\rho_i(s_i - s)$ represents the effect of an inflow (density $\rho_i$, salinity $s_i$) that has settled to a certain depth, $\omega$ the vertical advection caused by the inflow, $A$ the cross-sectional area perpendicular to the depth coordinate $z$, and $F_{\mathrm{d}}$ the diffusive flux.

The equation above deals only with salt, but below the mixed layer it is applicable to heat and oxygen concentrations as well, although a biological source/sink term should be added for the latter.

The molecular diffusive flux is generally described by Fick's law, $F_{\mathrm{d}} = -\rho k_{\mathrm{mol}}\ [(\partial s)/(\partial z)]$, where $k_{\mathrm{mol}}$ is the molecular diffusivity of salt, in this case. Analogously, the turbulent diffusion flux is written as $F_d^{turb} = -k_{turb}\,\rho[(\partial s)/(\partial z)]$. Unless otherwise stated, $k$ will hereafter stand for $k_{turb}$.

During periods of no inflow, Eq. (3) reduces further to

$$\frac{\partial(\rho s)}{\partial t} = \frac{1}{A}\frac{\partial}{\partial z}\left(A\rho k_s \frac{\partial s}{\partial z}\right). \quad (4)$$

## Oxygen dynamics

During stagnation periods, the oxygen concentration in the basin water is affected only by diffusive (turbulent) fluxes (as in Eq. 4) and the source/sink term $S_{O_2}$, and can be represented as

$$\frac{\partial \rho c_{O_2}}{\partial t} = \frac{1}{A}\frac{\partial}{\partial z}\left(A\rho k \frac{\partial c_{O_2}}{\partial z}\right) + \rho S_{O_2} \quad (5)$$

Below the euphotic layer, which approximately coincides with the mixed layer, $S_{O_2}$ can be considered a pure sink.

Since the molecular diffusivity of oxygen is of the same order of magnitude as the molecular diffusivity of salinity, the molecular diffusion of oxygen can be confidently neglected in the treatment of effective diffusivity. Therefore, the diffusive flux is purely turbulent and can be determined if $k$, $A$, $\rho$ and $c_{O_2}$ are known as a function of $z$. The total oxygen consumption rate $S_{O_2}(z)$ is then equal to sum of the change in oxygen concentration and the contribution from the diffusive flux.

To determine the profile of oxygen consumption, Eq. 5 must be integrated between measurement times $t_1$ and $t_2$. This yields for the sink term

$$S_b = \frac{c(t_2) - c(t_1) - \int_{t_1}^{t_2} \frac{1}{A}\frac{\partial}{\partial z}\left(A\rho k \frac{\partial c}{\partial z}\right)dt}{t_2 - t_1} \quad (6)$$

The remaining integral must be integrated numerically together with the corresponding equations for $s$ and $T$, as done by Stipa (1996), or approximated with a mean flux.

The oxygen concentration at station 5 at a depth of 30 m increases in winter and decreases in summer, when the oxygen consumption rate exceeds the combined advective and diffusive supply (Fig. 4).

For the purpose of determining an oxygen budget for the bay, it is an essential point that the deep water renewal of July 1991 can hardly be seen on the oxygen curve (Fig. 4). This would suggest that the oxygen content of the advected water was not much higher than that of the deep water. As the probable temperature of the inflowing water was roughly 10 °C, and that of the deep water 3 °C, a mixing ratio of 50% would give a final temperature of 6.5 °C, which is quite close to the observed 6.8 °C.

The oxygen concentration of the advected water, originating from the intermittently-present saline lower layer of the Stadsfjärden water (under number 8 in Fig. 1), cannot therefore have been higher than some 6 mg $l^{-1}$, which means that the assumption of a 100% saturation is not justified for the advected water. Without a knowledge of the oxygen concentration in the advected water, it is not possible to construct a closed oxygen budget even if the inflowing water volumes were known, except, of course for a stagnation period.

As was noticed earlier, there are no stagnation periods in a strict sense in Pohja Bay. The oxygen concentration indicates the same result: there are very few short periods, during which the oxygen depletion rate is the small negative number one would expect for a stagnation period (Fig. 4).

The diffusion model presented by Stipa (1996) was used to estimate the turbulent diffusion of oxygen, which was then subtracted from the oxygen depletion rate estimated from the measurements to obtain $S_b$ in Eq. 6. The analysis was done for days 856–989 (5 May–15 September, 1992). The result indicates an approximately constant oxygen sink of 0.06 mg $l^{-1}$ $d^{-1}$ for days 860–940, whereas the peaks are most likely attributable to advection (Fig. 11). Note the constancy of the oxygen sink with respect to both depth and time, when advection is absent. Based on the variation of the sink estimate, an error estimate of ± 0.01 mg $l^{-1}$ $d^{-1}$ would seem appropriate.

Therefore it is possible to conclude that 0.06 ± 0.01 mg $l^{-1}$ $d^{-1}$ is the oxygen sink which should be compared to biological measurements of the oxygen consumption. The respiration values of the water column are very low (H. Kuosa pers. comm.), therefore most of the oxygen consumption is expected to take place in the sediment surface.

The bottom oxygen consumption has been

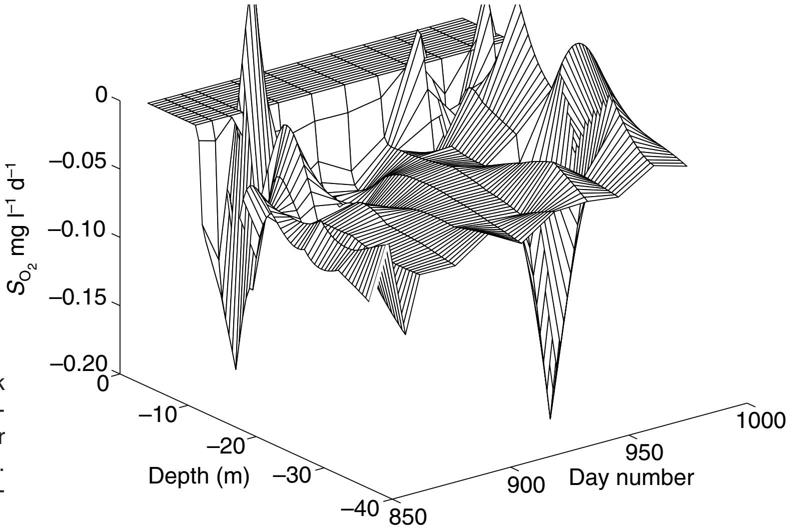


**Fig. 11**. Oxygen source/sink term during the whole computation period. The surface layer (0–10 m) was not computed. Note the flatness of the mid-depths.

measured as 230 ± 40 ml m$^{-2}$ d$^{-1}$. Assuming the oxygen is consumed below 15 m depth, the corresponding water volume is approximately $45 \times 10^6$ m$^3$. Then a total oxygen consumption of $2.7 \times 10^6$ g d$^{-1}$ is obtained from the oxygen budget presented in this paper. A bottom area of 8.2 km$^2$ would be needed to cause alone this oxygen consumption. Since the bottom area below 15 m is approximately 2.2 km$^2$, it seems likely that the bottom oxygen consumption cannot be held responsible for the whole oxygen consumption. The fate of oxygen in Pohja Bay needs further studies.

The role of the oxygen consumption in Stadsfjärden in the oxygen concentration in Pohja Bay would deserve a closer look; if the oxygen concentration of the inflowing water is already low before it enters the bay, it cannot increase the oxygen concentration in the bay, as was seen in July 1991. An increase in oxygen consumption outside the bay would therefore cause a decrease of oxygen concentration inside the bay.

Municipal waste waters, which are discharged into Stadsfjärden, increase the oxygen consumption there in summer, decreasing the oxygen concentration in the dense bottom layer, which has only two ways out of Stadsfjärden: by mixing into the surface water or by flowing downstream into Pohja Bay. Therefore, a considerable amount of the nutrients that are discharged into Stadsfjärden in the summertime may end up in Pohja Bay instead of flowing out to the sea.

## A conceptual model of Pohja Bay

A conclusive summary of the physical processes acting in the bay is given below; the arguments that lead to the conclusions can be found in the text above or in Stipa (1996).

### Stratification

The waters in the bay are stably stratified throughout the year. The stratification is almost completely caused by the salinity difference between the nearly-fresh surface layer and the deeper brackish waters.

The waters in the bay can be divided into three layers according to their renewal time:

1. The mixed layer, which extends from the surface to the pycnocline at a depth of 5–10 m. This layer is constantly renewed by the fresh water discharge from the rivers and the entrained brackish water from below the pycnocline. Its depth depends somewhat on the season, but the strength of the stratification, the buoyancy supply from fresh water and the protection provided against wind mixing by the proximity of the shores limit the mixing to the uppermost 10 m. The mixed layer comprises roughly 70% of the water volume in the bay. This is also where the primary production and energy exchange takes

place. It also acts as a sink for the salinity in the deeper water.

2. The intermediate layer is a somewhat more vague concept, extending from the pycnocline to a depth of about 20 meters. This layer experiences frequent inflows from outside the bay, and is renewed semi-continuously. Advection cannot therefore be neglected when considering the various budgets in this layer. The water volume in this layer makes up approximately 23% of the total volume.
3. Deep water can be defined as the layer in which advective transport plays a negligible role during certain periods of the year. This layer extends from 20 m to the bottom.

The advection, however, cannot be completely overlooked without the possibility of serious errors. Advective transport can take place at any time of the year, although statistically winter is the favorable season.

In the absence of advection, the vertical fluxes of salinity, temperature, oxygen and other relevant dissolved substances in the bay are driven by turbulent diffusion, which in turn is driven below the pycnocline presumably by breaking internal waves. In a first approximation, the turbulent diffusive flux depends linearly on the vertical gradient of a diffusing quantity. The oxygen budget involves an additional source/sink term, which in this study was determined as $-0.06$ mg $l^{-1}$ $d^{-1}$. A similar term is needed for the description of other non-conservative substances.

An inflow entering the bay descends to the depth where the ambient density equals that of the inflow. At the settling depth, the inflow starts spreading horizontally and finally mixes with the surrounding water. From the settling depth it lifts the older, lighter water up. As a counter-action, the surface mixing tends to increase the pycnocline depth by entrainment. Since the pycnocline shows no large vertical movements between inflow events, the processes cancel each other, and the inflows in this way drive the upward flux of substances through the pycnocline.

## Water exchange

Pohja Bay is a semi-enclosed, stably stratified, deep estuary, whose deep water is renewed by intermittent inflows. The hydraulic control limiting the inflow is stronger during the summer and weaker during the winter.

The barotropic (externally forced) water exchange of the bay is driven by the water level changes caused by astronomical and meteorological tides, and by the fresh water discharge. The primary driving force for the barotropic water exchange is the water level gradient between the bay and the sea. Unfortunately, the use of the water level at Hanko was found to give unsatisfactory results.

The water level changes would cause a mean flow of $\pm$ 65 $m^3$ $s^{-1}$, but the river runoff causes a mean net outflow of 20 $m^3$ $s^{-1}$, which skews the flow rate distribution by an equal amount towards net outflow. The astronomical tide causes at least 40% of the barotropic water exchange.

Qualitatively the baroclinic water exchange is essentially hydraulically controlled by two distinct layers, namely the fresher mixed layer of Pohja Bay and the saltier, intermittently-present Baltic water layer in the area between Vitsandsströmmen and the bridges.

The flow in the lower layer (towards the bay) is controlled by the layer thicknesses and their densities. The thicker the upper layer, i.e. the mixed layer in the bay, the larger the control on the lower layer. The denser the lower layer on the outside, the easier it flows towards the bay.

Barotropic flow, superimposed on the baroclinic flow, forces the layer interface to move in the horizontal direction in the contraction. By doing so, it changes the velocities in both layers so that the sum of the volumes transported by the layers equals to the total barotropic transport.

In an extreme case, barotropic forcing can stop the flow in either of the layers (depending on the flow direction), and even cause the retreat of one layer from the contraction area, i.e. "box flow". In the present case, box flows can only be expected for barotropic flows out of the bay, since the area outside the contraction almost always has a surface layer.

Varying barotropic forcing (astronomical and meteorological tides) has been shown to increase the transport capacity of a contraction (Stigebrandt 1980).

The factors affecting the rate of Baltic water inflow into Pohja Bay would then be the same as those affecting the layer thicknesses and their densities, i.e.:

1. Wind mixing thickens the mixed layer in the bay and thereby decreases the inflow rate. It also entrains water from the lower layer in the bay entrance area and thus decreases the lower layer thickness
2. External stratification affects the lower layer density and, to some extent, its thickness. Higher density increases the inflow rate.
3. Fresh water discharge decreases the surface layer density, but also increases the net outflow rate.
4. Ice cover effectively reduces surface mixing, which should in turn yield a thinner surface layer and larger inflow rate in the lower layer. The ice cover is, however, closely tied up with fresh water discharge changes and stratification, which makes the assessment of its own role difficult or even fruitless.

For a more detailed study of the dynamics of baroclinic water exchange (the inflows), continuous measurements of the stratification profile at Stadsfjärden, at the bridges as well as in Sällvik deep will be necessary. In addition, currents in the narrowest section should be measured at at least two depths.

*Acknowledgements*: The measurements used in this work were planned by prof. Jouko Launiainen and Dr. Jouko Pokki. The field and laboratory work, without which this study would not exist, was done by Torsten Sjölund, Svante Degerholm, Marita Pokki, Anne-Marie Åström and Ulla Sjölund. In addition the Finnish Institute of Marine Research and the Finnish Environmental Agency are acknowledged for sea level and river discharge data. Leena Karjala (Finnish Institute of Marine Research) provided the bottom oxygen consumption estimate. I thank prof. Jouko Launiainen for his supervision, and Timo Vihma, Matti Leppäranta as well as many others at the Department of Geophysics and Tvärminne Zoological Station for helpful comments and general support. Financial support from Maj and Tor Nessling foundation has made this study possible.

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Received 20 October 1995, accepted 26 June 1999