Tellus (1993), 45A, 311–324
Printed in Belgium – all rights reserved

TELLUS
ISSN 0280–6495

Deep water exchange in the Baltic Proper

By TARMO KÕUTS[1] and ANDERS OMSTEDT[2]*, [1]*Estonian Meteorological and Hydrological Institute, EE-0106 Tallinn, Estonia*; [2]*Swedish Meteorological and Hydrological Institute, S-601 76 Norrköping, Sweden*

(Manuscript received 13 October 1992; in final form 20 April 1993)

ABSTRACT

Deep-water exchange and mixing properties in the Baltic Proper were analysed on the basis of temperature and salinity data measured during the period 1970–1990. The data were analysed applying basic model concepts as the conservation principles, the two-layer approach and the geostrophic flow assumption. The renewal of the deep water in the Baltic Proper consists of inflow from upstream basins. The inflowing dense water is diluted by surface water and on a 20-year average increased by a factor of 4, when entering from the Kattegat into the Landsort Deep. Three main mixing zones were localized. Firstly, the Belt Sea and the Sound, where the deep-water inflow increases by 79%; secondly, the Arkona Basin, where vertical mixing causes the increase of deep current volume flow by an average of 53%; and thirdly, in the Stolpe Channel, where the turbulent entrainment adds an average of 28% to the deep current. Applying the geostrophic flow model on salinity data, time series of deep current flow rates were calculated. The model was calibrated by 20-year mean flows calculated from conservation principles. The effective sill depths were introduced as calibration coefficients. It was found that the geostrophic flow model described deep-layer flows in the Bornholm Channel well, but the flow rate was underestimated in the Stolpe Channel and overestimated in the Fårö Channel. In the Stolpe Channel and in the Fårö Channel, the deep-layer flow showed seasonal variations with rapid increase during the autumn and winter seasons, respectively.

1. Introduction

The renewal of the Baltic Sea deep water takes place through inflow of more saline waters from the Kattegat (Fig. 1). The inflowing water moves as a dense bottom current mainly through the Great Belts (Jacobsen, 1980). In the Arkona Basin, the inflowing water forms a dense, about 5–15 m thick bottom layer with salinities up to 24 PSU, (Matthäus, 1985). Due to meteorological forcing, the water exchange between the Baltic Sea and the Skagerrak is, however, highly variable. Entering into the Arkona Basin, the salinity of the inflowing water is decreased due to mixing with less saline surface water (Sturm et al., 1986). As the Arkona Basin is quite shallow, strong winds can cause intensive mixing and dilution of the dense bottom water.

* Corresponding author.

From the Arkona Basin, the water leaks through the Bornholm Channel into the Bornholm Basin. Direct observations of inflowing water through the Bornholm Channel have been performed by Petrén and Walin (1976), see also Walin (1981). It was observed that the inflowing dense bottom water was distributed over salinities from 8 to 18.5 PSU, having a flow intensity of $1.55 \cdot 10^3$ m^3/s per PSU. After entering the Bornholm Basin, the dense bottom water continues through the Stolpe Channel into the Southern Gotland Basin and the Eastern Gotland Basin.

From the Eastern Gotland Basin, the water flows further north and finally enters into the Landsort Deep through quite complex bottom topography. The maximum depth in the Landsort Deep is 459 m, which is also the deepest part of the Baltic Sea. A typical salinity transect through the deepest parts of the Baltic Sea is given in Fig. 1b.

Several physical factors may influence the flow

Tellus A/45/4-6

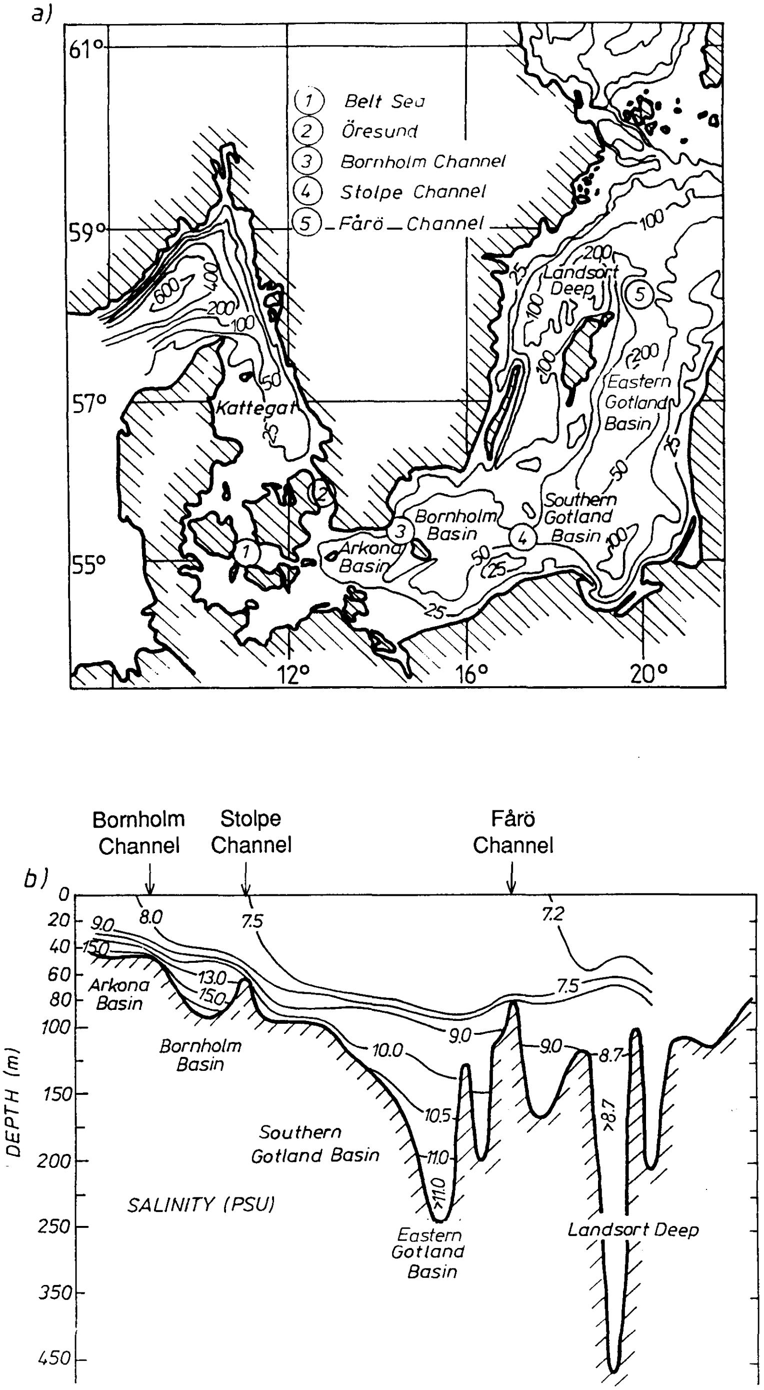

Fig. 1. Map of the Baltic Sea with locations of sub-basins (a) and a vertical transect giving bottom topography and salinity distribution in the estuary (b). The salinities were measured during the cruise of R/V ARGOS in January 1992.

characteristics of a dense bottom current: topography, bottom friction, mixing with surrounding water, rotation, density difference between inflowing and surrounding water masses. At present, there exists no consensus as to the most proper mathematical formulation, but different aspects of the dynamics have been focused on in the modelling efforts. The concept of dynamical control sections in the Baltic Sea was first discussed by Rydberg (1980). Pedersen (1977) focused on bottom friction and entrainment, and Lundberg (1983) added a discussion of the effects due to pressure. In the models by Stigebrandt (1983, 1987a) and Omstedt (1987, 1990), the inflow was assumed to be in geostrophic balance at some different control sections associated with sills. Gidhagen and Håkansson (1992), applied a numerical two-layer model to the Bornholm Channel and argued from some idealized calculations that both friction and rotation are limiting the transport through the channel. The wind effects on the deep water flow through the Stolpe Channel were discussed by Krauss and Brügge (1991) on the basis of numerical model calculations. They observed that northerly and easterly winds increase the transport of dense water from the Bornholm Basin through the Stolpe Channel into the Eastern Gotland Basin.

The purpose with the present work is to examine the deep water properties and exchange in the Baltic Sea on the basis of temperature and salinity data from a 20-year period (1970–1990). Some basic model concepts are introduced, and the main outcomes from the study are estimates of the long-time mean flow and entrainment rates, as well as time variations of the deep layer flows between different subbasins within the Baltic Sea.

In Section 2, the basic concepts are introduced in the derivation of the models. The data and our method of interpolating data are presented in Section 3. The results together with a discussion are given in Section 4. Finally, some conclusions are given in Section 5.

2. Theoretical considerations

2.1. The two-layer approach

The vertical stratification in an estuary like the Baltic Sea is mainly dependent on salinity, with brackish water on top of more saline water. Even though the vertical stratification is multi-layered, the two-layer approach captures the basic features (Fig. 1b).

The two-layer approach together with a division of the Baltic Sea into subbasins provides us with a theoretical framework, from which the dense bottom water flow within the deeper parts of the Baltic Sea can be calculated. In each sub-basin, we also assume that horizontal gradients are small compared to vertical.

The sub-basins considered are: the Arkona Basin, the Bornholm Basin, the Eastern Gotland Basin and the Landsort Deep. We define the two layers in each sub-basin by the volume-weighted salinities in the surface layer (S_1) and the bottom layer (S_2) and by the halocline depth (H_1). The salinities then read:

$$S_1 = \frac{1}{V_1} \int_{A(z)} \int_0^{H_1} S(z)\, \mathrm{d}z\, \mathrm{d}A, \tag{1}$$

$$S_2 = \frac{1}{V_2} \int_{A(z)} \int_{H_1}^{H} S(z)\, \mathrm{d}z\, \mathrm{d}A, \tag{2}$$

where V_1 and V_2 denote volumes above and below the halocline respectively, $A(z)$ is the area at some depth z in the sub-basin under consideration, and H is the total depth. The sum of the two volume-weighted salinities multiplied with the surface and deep water volumes respectively gives the total salt content. The halocline depth (H_1) is defined as the depth where the salinity gradient has its maximum value (Fig. 2).

2.2. Conservation of salinity and flow

As a long-term average, the inflow intensity into the deep layer of the Baltic Sea was first calculated by Knudsen (1900). Together with the conservation of salinity for a steady two-layer flow the hydrographic theorem reads:

$$Q_{\text{out}} - Q_{\text{in}} = Q_{\text{riv}} + Q_{\Delta}, \tag{3}$$

$$Q_{\text{out}} \cdot S^{\text{out}} - Q_{\text{in}} \cdot S^{\text{in}} = 0, \tag{4}$$

where Q_{Δ} denotes the fresh water supply from the difference between precipitation and evaporation and S^{in} and S^{out} represent the salinities of inflowing and outflowing water to and from the Baltic Sea, respectively.

The inflowing dense deep water will be diluted

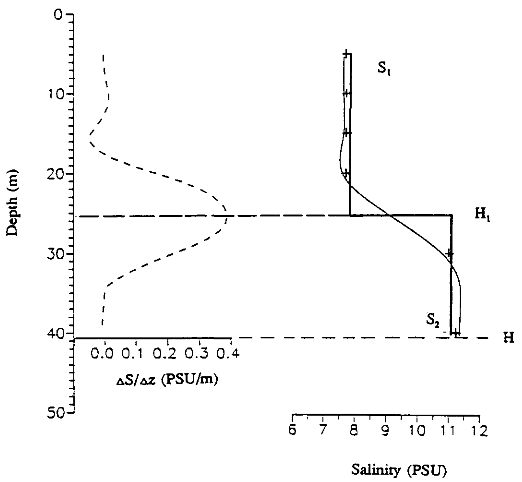

Fig. 2. An example of the splined profile of salinity (thinner solid line) in the Arkona Basin together with calculated surface (S_1) and bottom (S_2) layer salinities and halocline depth (H_1). Crosses indicate measured standard depth values. The dashed line shows the vertical gradient of salinity ($\Delta S/\Delta z$) with the maximum at the halocline depth.

by surface water (Stigebrandt, 1987a) in the entrance area and in the Arkona Basin, which is a rather shallow basin (maximum depth 55 m). The wind forcing increases the vertical mixing, see further discussions by Sturm (1986) and Fennel et al. (1987). The conservation of volume flow and salinity of a deep current diluted by a fresher surface layer can be expressed as:

$$Q^{i-1} + Q^{i}_{\text{ent}} = Q^{i}, \tag{5}$$

$$Q^{i-1} \cdot S_2^{i-1} + Q^{i}_{\text{ent}} \cdot S_1^{i} - Q^{i}_{\text{diff}} \cdot S_2^{i} = Q^{i} \cdot S_2^{i}, \tag{6}$$

where Q^{i}_{ent} denotes the entrainment of the surface water into the deep current, Q^{i}_{diff} the turbulent salt diffusion into the surface layer, Q^{i-1} the inflow into the sub-basin with the salinity S_2^{i-1}, and Q^{i} the outflow from the sub-basin. S_1^{i} and S_2^{i} are the salinity of the surface and the deep layers in the sub-basin, respectively. From different estimations, the turbulent salt diffusion volume flows are found to be in the order of 10 to 100 m^3/s (Matthäus, 1990; Rahm and Wulff, 1992). They are some order of magnitudes less than the inflows (see Subsection 4.3) and will therefore not be further dealt with in the present analysis. The same model was recently applied by Marmefelt and Omstedt (1992) for a similar study of the deep water exchange in the Gulf of Bothnia.

We now assume that as a long-time mean, the average deep layer inflow is equal to the sum of river runoff and net precipitation $Q_{\text{in}} = Q_{\text{riv}} + Q_{\Delta}$ and that all inflow into the Baltic Sea will pass through the deepest connection between the Arkona Basin and the Bornholm Basin, the Bornholm Channel. The sill depth in the Bornholm Channel is about 48 m, and the depth south of the Bornholm Island is less than 30 m. The salinity of outflowing surface layer water is assumed to be equal to the surface layer salinity in the Arkona Basin (S_1^{AR}), thus $S_1^{\text{out}} = S_1^{\text{AR}}$. The long time average river runoff was put equal to 15,000 m^3/s (Mikulski, 1986), and the net supply from precipitation (difference between precipitation and evaporation) was put equal to 2000 m^3/s according to Henning (1988).

The eqs. (3) to (6) together with calculated volume-weighted long term mean salinities of the surface and deep layers form now a closed system for calculation of the mean deep layer inflow and entrainment rates into the deep current. This model will be applied to salinity data presented in Section 3, and the results are given in Subsection 4.1. In the next section we should first, however, introduce another simple model for the calculation of the flow rates, that is the geostrophic flow model.

2.3. *Geostrophic flow model*

Starting from thermal wind equation in a right-handed coordinate system where the x-axis is directed across the channel, the vertical gradients of the along channel component of the geostrophic velocity (v_g) together with a simplified equation of state read:

$$\frac{\partial v_g}{dz} = -\frac{g\beta}{f}\frac{\partial S(x, z)}{\partial x}, \tag{7}$$

$$\rho(x, z) = \rho_0(1 + \beta S(x, z)), \tag{8}$$

where g is gravity acceleration, f Coriolis parameter, $\rho(x, z)$ density of seawater and constant ρ_0 density of freshwater. The coefficient β in the equation of state is put equal to $8 \cdot 10^{-4}$ PSU^{-1}. The underlying assumption is that friction and entrainment effects are small, which is not always the case. This will be further discussed in Subsection 4.3.

If eq. (8) is integrated both vertically and horizontally and the water in the surface layer is assumed to be at rest, the following equation could be derived,

$$Q_g = \frac{g\beta(H_{\text{sill}} - H_1)^2}{2f}(S_2 - S_1), \tag{9}$$

where Q_g denotes the geostrophic volume transport through the channel and H_{sill} the sill depth. This equation now defines our geostrophic flow model. The model has also been used by Stigebrandt (1983, 1987a, b) and by Omstedt (1987, 1990) and will be further discussed in Subsection 4.3.

3. Material and methods

Data of measured sea water salinity and temperature for the period 1970 to 1990 have been taken from the database of the International Council for the Exploration of the Seas (ICES). One representative station for every sub-basin was chosen. The representative stations are: BY2 for the Arkona Basin, BY5 for the Bornholm Basin, BY15 for the Eastern Gotland Basin and BY31 for the Landsort Deep (Fig. 1). The stations are situated at the deepest parts of the different sub-basins and represent mainly the open basin properties. They therefore do not give a good representation of the shallow coastal areas, Rahm (1988). Added to these stations are: Fladen, representing the Kattegat, BY8, representing the Southern Gotland Basin and BY28, representing the Northern Gotland Basin. The positions of the oceanographic stations are given in Table 1.

Table 1. *The oceanographic station*

Stations	Latitude	Longitude
Fladen	57°11.5′	11°40′
BY2	55°00′	14°05′
BY5	55°15′	15°59′
BY8	55°38′	18°36′
BY15	57°20′	20°03′
BY28	59°02′	21°05′
BY31	58°35′	18°14′

The monthly distribution of the number of observations during the studied period at each station is given in Table 2. The water masses were analysed on the basis of the temperature and salinity data from these stations. The density was calculated from the salinity and temperature data according to the equation of state defined by UNESCO, see Gill (1982). Since the Baltic Sea is a rather shallow and salinity-stratified sea (maximum depth 459 m but mostly not more than 250 m), the pressure effect upon the density is neglected. Density in this paper is presented in σ_t units, which is equivalent to $(\rho - 1000) + 0.025$, where ρ is the density of seawater calculated from in situ measurements of temperature and salinity.

All data were measured at standard depths. To interpolate these discrete values to continuous vertical profiles, the spline method was applied. The coefficients of the cubic spline polynomials were calculated using the algorithm developed by Reinsch (1967). The method is called the smoothing spline method. An example of a splined profile together with defined salinities of surface and

Table 2. *Monthly distribution of number of the measurements for each station during the period 1970–1990*

Station	Jan.	Feb.	Mar.	Apr.	May	Jun.	Jul.	Aug.	Sep.	Oct.	Nov.	Dec.
Fladen	21	21	25	25	38	26	20	32	33	25	29	17
BY2	21	20	27	20	32	29	13	36	22	16	30	12
BY5	14	23	35	22	37	46	12	46	35	19	39	8
BY8	12	18	24	12	33	27	5	32	16	11	30	7
BY15	16	14	19	21	33	30	16	40	16	15	31	6
BY28	10	5	12	7	38	21	10	27	18	18	28	4
BY31	20	10	22	10	30	35	19	29	23	16	30	17

bottom layers and halocline depth is given in Fig. 2. The salinity structure of the different sub-basins, including the Kattegat, was analysed using the smoothing spline method. From every splined salinity profile, the halocline depth, the volume-weighted surface and bottom layer salinities were calculated according to eqs. (1) and (2). The calculated time series of the salinities and the halocline depths were smoothed by averaging over a time period of one month. Variations shorter than one month are thus not dealt with. Based upon the two-layer model approach, presented in Subsections 2.1 and 2.2, the mean flow and entrainment rates were calculated for the studied period and for the different sub-basins. Using the two-layer approach and the assumption of geostrophically balanced flows over the sills as well as conservation principles, the time series of deep-water flow rates were also calculated according to eq. (9) together with an analysis of variability. The sill areas considered were the Bornholm Channel, the Stolpe Channel and the Fårö Channel (see Fig. 1).

4. Results and discussions

4.1. Mean properties

4.1.1. Mean stratification. In this section, 20-year mean properties are presented and discussed. The average volume-weighted salinities of upper and lower layers and the mean halocline depths for the different sub-basins are given in Table 3. It can be noticed that the mean surface salinities in the Baltic Sea decrease by about 1 PSU when entering from the Arkona Basin into the Landsort Deep, while the mean bottom salinities decrease by about 4 PSU.

Although the Bornholm Basin is downstream compared to the Arkona Basin, the salinities are slightly higher there. The reason for this is that the bottom water in the Bornholm Basin creates a more stable dense bottom pool compared to the Arkona Basin, as the Bornholm Basin has a sill, which the Arkona Basin does not have. The mean halocline depth increases downstream, but the mean halocline depth in the Landsort Deep is slightly less than in the Eastern Gotland Basin. This feature is also indicated in Fig. 1b. The reason is the definition of the halocline depth. Other definitions can be applied. For example, Stigebrandt (1987b) introduced a definition, where observed stratification was transformed to a dynamically "equivalent" two-layer stratification.

4.1.2. Mean flows. Based upon the parameters of the mean stratification in different sub-basins, presented in Table 3 and the two-layer model presented in Subsection 2.1, the 20-year mean flow and entrainment rates were calculated using eqs. (3)–(6).

Applying eqs. (3) and (4) in the entrance of the Arkona Basin and assuming $Q_{\text{in}} = Q_{\text{riv}} + Q_{\Delta}$ we estimate that inflowing water has an average salinity that is two times stronger than that of outflowing surface layer water $S_{\text{in}} = 2S_{\text{out}}$. Using the assumption that $S_{\text{out}} = S_1^{\text{Ar}}$ (see Subsection 2.2), we get that $S_{\text{in}} = 16.9$ PSU. On the way into the Baltic Sea, the deep dense current will entrain the surface layer water until it has a salinity of $S_2^{\text{Ar}} =$ 14.0 PSU (Table 3). When applying now eqs. (5) and (6) on these salinities, we estimate that the mean deep outflow from the Arkona Basin into the downstream sub-basin is $Q^{\text{Ar}} = 25{,}900$ m^3/s, and that the entrainment of surface layer water added an average of 53% to the deep volume flow.

Similar estimations were made between the Kattegat and the Belt Sea/the Sound. Thus applying (5) and (6), we estimate that the entrainment into the Belt Sea/the Sound adds 79% to the mean flow.

From the Arkona Basin dense water leaks into the Bornholm Basin. The inflow with higher densities will enter into the deeper layers and the inflow with intermediate densities enters into intermediate layers. While the halocline in the Bornholm Basin lies at an average of depth 59 m, the direct wind mixing can not reach the deep layer during most of the year. Assuming the conserva-

Table 3. *20-year mean properties of surface salinity (S_1) and deep water salinity (S_2) and halocline depth (H_1) based upon data from 1970–1990*

Basin	S_1 (PSU)	S_2 (PSU)	H_1 (m)
Kattegat	23.6	32.4	15.0
Arkona Basin	8.45	14.0	35.0
Bornholm Basin	8.6	14.4	59.0
E. Gotland Basin	7.9	11.9	76.0
Landsort Deep	7.4	10.5	65.0

tion of volume below the mean halocline depth it is reasonable to assume that as much as enters into the Bornholm Basin deep layer will also leave it, thus $Q^{Bh} = 25{,}900\ m^3/s$. The outflow from the Bornholm Basin occurs through the Stolpe Channel. In the recent paper by Krauss and Brügge (1991), the authors argue that northerly and easterly winds may increase the volume transport through the Channel considerably. The idea is that the deep-layer flow increases in contra-direction to wind because of the inclination of the sea surface caused by the wind-driven currents. As the sill depth in the Stolpe Channel is about 60 m and the mean halocline depth is 59 m, the entrainment of surface layer water into the gravity current is probably driven by turbulence generated by bottom boundary mixing. Applying eqs. (5) and (6) in the Stolpe Channel where the mean deep layer salinity is $S_2^{ST} = 13$ PSU, we estimate the entrainment rate in the Stolpe Channel to 28%. The mean deep layer flow through the Channel is thus estimated to $33{,}200\ m^3/s$.

Further downstream the top of the deep layer is at the depths of 76 m and 65 m in the Eastern Gotland Basin and in the Landsort Deep respectively, which is shielded by the seasonal thermocline during most of the year. Applying the eqs. (5) and (6) in the sill area between the Eastern Gotland Basin and the Landsort Deep (the Fårö Channel) we estimate that the mean entrainment rates of the surface water into the deep current is 13.5%, and the mean deep layer salinity there is $S_2^{FÅ} = 10.9$ PSU. Deep-layer volume flow then increases up to $Q^{FÅ} = 37{,}500\ m^3/s$. The calculated mean deep flows and entrainment rates in the different sub-basins are all given in Table 4.

Using the geostrophically balanced flow assumption we will discuss the time dependent water exchange between the sub-basins in Subsection 4.3, but first the water masses which participate in the deep water renewal will be discussed.

4.2. *Water masses*

The oceanographic conditions in each sub-basin can be analysed from TS-plot diagrams (Figs. 3a–f), where the thermohaline structure of the water masses in each sub-basin are presented.

When entering the Baltic Sea, one can basically identify 3 different water masses, see for example the plotted data from the Eastern Gotland Basin. The 3 water masses are: the surface water with variable temperatures but small variations in salinity, the halocline water with small variations in temperature and salinity, and the deep water with small variations in temperature but some variations in salinity. Exceptions to these typical water masses can be noticed in the Arkona Basin and the Bornholm Basin where large temperature variations also are present in the deep layer.

On the TS diagrams the deep water of each sub-basin is marked with a solid line box, and the deep water transformation cascade can be followed from the Kattegat to the Landsort Deep. Main mixing zones are situated between the Kattegat

Table 4. *20-year mean flow and entrainment rates based upon data from 1970–1990*

Basin	Upstream deep water inflow (m^3/s)	Downstream deep water outflow (m^3/s)	Entrainment (%)
Kattegat	—	9,500	—
Belt Sea	9,500	17,000	79
Arkona Basin	17,000	25,900	53
Bornholm Channel	25,900	25,900	0
Bornholm Basin	25,900	25,900	0
Stolpe Channel	25,900	33,200	28
E. Gotland Basin	33,200	33,200	0
Fårö Channel	33,200	37,500	14
Landsort Deep	37,500	—	—

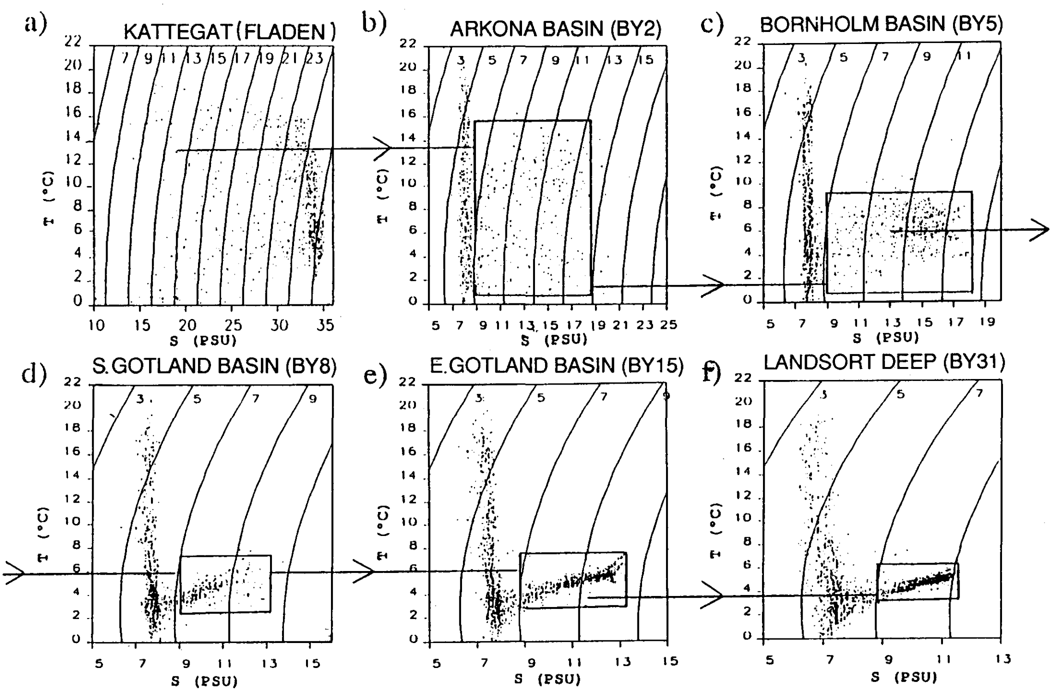

Fig. 3. TS diagrams of water masses from different sub-basins of the Baltic Sea. On the diagrams are plotted isolines of density (σ_t units). Temperatures and salinities were measured at the stations given in the figure. The squares plotted by a solid line indicate the deep water of the sub-basin. The arrows between squares indicate the deep water transformation cascade in the Baltic Sea.

and the Arkona Basin and between the Bornholm and Southern Gotland Basins.

Mixing between the Kattegat surface layer water with salinities from 18 to 26 PSU and the Arkona Basin surface layer water (Figs. 3a, b) in the Belt Sea and the Sound form the deep water of the Arkona Basin and the Bornholm Basin. The bottom water in the Arkona Basin has a salinity variation of 9 to 19 PSU and a high temperature variation from 0.5 to 15°C. Further leakage into the deep layer of the Bornholm Basin occurs without mixing with surrounding water (Fig. 3b). Due to a sill depth of 60 m in the Stolpe Channel (the maximum depth of the Bornholm Basin is about 105 m) a stable dense pool with salinities up to 18 PSU is formed. From Fig. 3b, we also notice that the deep water of the Arkona Basin (having a temperature of more than 9°C) does not reach the deep water of the Bornholm Basin. This indicates that the main inflow of saline water into the Baltic Sea deep layers takes place during the period from October to May. The characteristic thermohaline structure of the main Baltic Proper deep water (salinities 9–13 PSU and temperatures 2–7°C, Figs. 3c, d) forms between the Bornholm Basin and the Southern Gotland Basin. The sill depth in the Stolpe Channel is a controlling factor of the deep water inflow further into the downstream sub-basins. Deep water with a salinity of more than 13 PSU is arrested in the Bornholm Basin until more dense inflow raises it above the sill. However, such a mechanism works only during major inflows, which are rather unusual (Fonselius, 1969; Matthäus and Franck, 1992). Turbulent vertical diffusion slowly decreases the deep-water salinities during stagnation periods (Walin, 1977; Rahm, 1985). The characteristic TS structure of the deep-water of the main Baltic Proper does not change any more downstream from the Southern Gotland Basin (Figs. 3e–f), except for the fact that the sill area between the Eastern Gotland Basin and the Landsort Deep cuts off salinities larger than 11.5 PSU.

On the basis of the calculations of flow and entrainment rates in Subsection 4.1, we can estimate the salinity and temperature range needed to form the downstream bottom water masses. Any water mass at the Arkona Basin (BY2), having a σ_t higher than 7.0 will reach the bottom water in the downstream basins. There are two main ways for deep water to flow downstream. Firstly, water being dense enough ($\sigma_t > 10$) will leak into the stable dense pool of the Bornholm Basin and will be arrested until the vertical advection raises it into σ_t layer between (7–8). Secondly, water with a density range σ_t between (7–8) can directly flow over the sill and continue into the intermediate layer (the halocline layer) of the Eastern Gotland Basin. This way of water renewal can be called "the ventilated halocline of the Eastern Gotland Basin". In this way the water in the halocline layer is renewed directly and more rapidly than by advection from deeper layers. The model of the vertical circulation of the Baltic Sea developed by Stigebrandt (1987a) showed that the inflow into the Baltic Sea occurs mostly in the intermediate salinity interval and thus causes a much shorter residence time in the intermediate layer compared to the deeper layer.

4.3. Deep water flows

The density-driven deep water flow rates were calculated for each sub-basin. As the density is mostly a function of salinity in the Baltic Sea, the flows were calculated only on the basis of monthly mean salinity data. The equation of state, eq. (8), and the geostrophic inflow model presented in Subsection 2.3 were used in the calculations. For the calibration of the model, the 20-year mean geostrophic flows were calibrated to become equal to the flows based upon conservation laws presented in Subsection 2.2. The variations of the halocline depth in the different sub-basins are first presented in Fig. 4. From the figure we observe

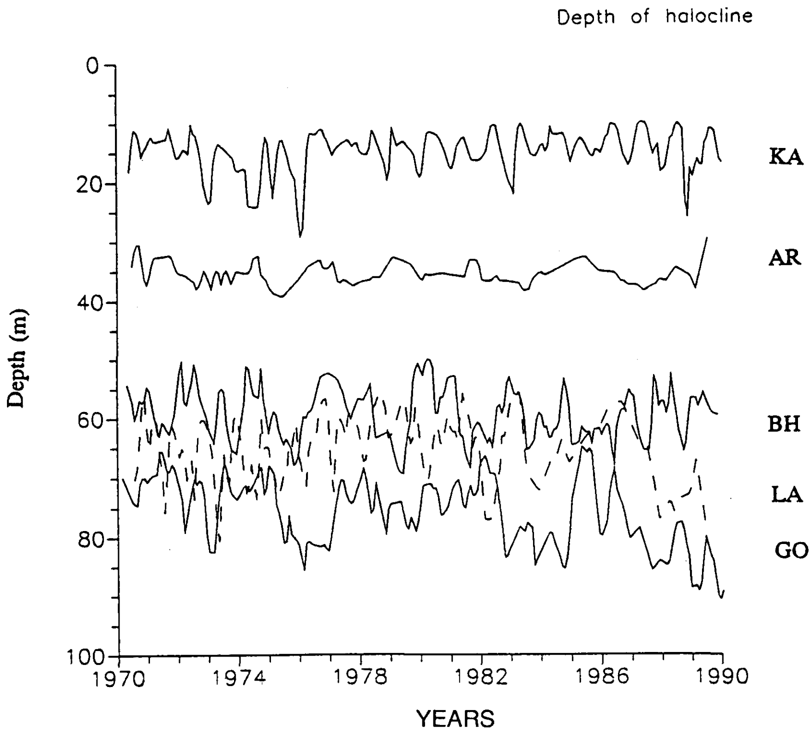

Fig. 4. Time series of calculated halocline depths for the Kattegat (KA), the Arkona Basin (AR), the Bornholm Basin (BH), the Eastern Gotland Basin (GO) (solid lines) and the Landsort Deep (LA) (dashed line).

that the halocline depths in the downstream sub-basins are deeper than those in the upstream sub-basins. Thus the assumption of unidirectional deep layer flow from Kattegat through the Arkona, Bornholm and Southern Gotland Basins to the Eastern Gotland Basin is valid. In the Landsort Deep the halocline (dashed line in Fig. 4) is less deep compared to the Eastern Gotland Basin almost during the whole studied period, owing to the definition of the halocline depth.

The geostrophic flow model, eq. (9), is sensitive to the layer thickness between halocline and sill. A variation of this difference with a factor of 10 changes the flow by a factor of 100. To calibrate the geostrophic inflow model against the conservation model, we used the sill depths as the calibrating factor. This implies that we introduce an effective sill depth instead of a topographical one in a similar manner as Stigebrandt (1987b). For the Bornholm Channel, Stigebrandt (1987b) calculated the effective sill depth to 41 m, instead of the topographic sill depth, which is 48 m. The effective and topographic sill depths for the different sub-basins calculated in the present study are given in Table 5. The calibrations of the effective sill depth were through iterations comparing mean conservation flows with mean calculated geostrophic flows over the 20-year time period. The results of the calculations gave that the effective sill depth is close to the topographical one in the Bornholm Channel and differs about 15–25% in the Stolpe Channel and in the sill area between the Eastern Gotland Basin and the Landsort Deep (the Fårö Channel). The effective sill depth that is greater than the topographic one in the Stolpe Channel indicates the existence of an additional mechanism which increases the deep-water transport through the Channel. As argued by Krauss and Brügge (1991) for favorable wind forcing the flow through the Stolpe Channel can be increased.

Table 5. *Calculated effective sill depths and corresponding topographical values*

Sill	Sill depth (m)	
	effective	topographical
Bornholm Channel	48	48
Stolpe Channel	71	60
Fårö Channel	92	115

The difference between topographic and effective sill depths in the Fårö Channel can be caused by several factors, as, for example, bottom friction, which will reduce the volume transport.

The geostrophic inflow rates in different sub-basins during the period 1970–1990 are given in Fig. 5. The coupling between the different sub-basins are shown in some more detail in Fig. 6. During the period 1975 to 1979 several inflows were observed in groups (Matthäus and Franck, 1992). In Fig. 6, one observes intensification of the deep flow with some delay in the different sub-basins. The typical time delay for an inflow from the Bornholm Channel to be observed in the Stolpe Channel was about half a year, and to be observed in the Fårö Channel about a year. From Fig. 6 we also notice that the duration of inflow in the different sub-basins is approximately the same, one year and a half.

The distribution of the inflowing water masses among the different salinities is shown in Fig. 7. Two main deep water inflow classes can be defined.

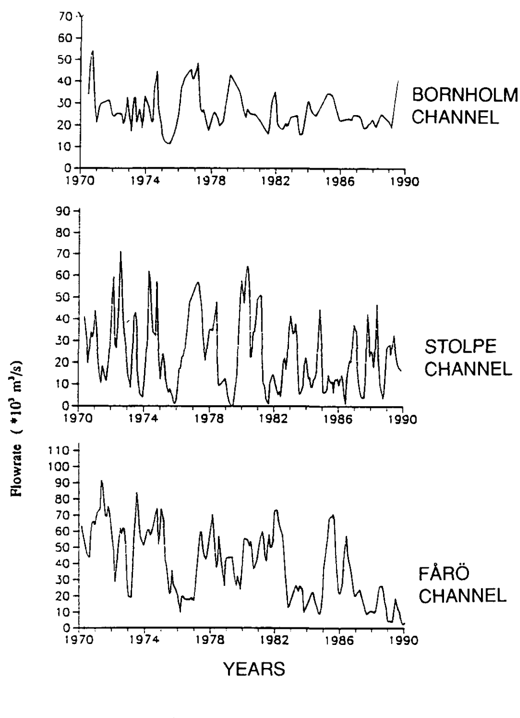

Fig. 5. Geostrophic outflow calculations for the period 1970–1990. The flows reflect the vertical salinity structure in the upstream basins and the sill depths in the channels.

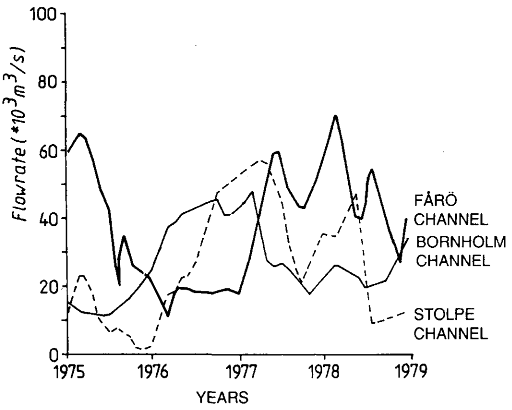

Fig. 6. Calculated time series of the geostrophic flow rate in the different channels during a group of major inflow events 1975–1979.

Firstly, the deep water inflow with salinities between 13 and 15 PSU, occurring in the Bornholm and the Stolpe Channels. Secondly, the main Baltic Proper deep-water flow with characteristic salinities between 10 and 12 PSU, occurring from the Southern Gotland Basin to the Landsort Deep. Stigebrandt (1987b) calculated the mean geostrophic flow rate for the Bornholm Channel to be 23,650 m^3/s with a mean salinity of 13.59 PSU and a salinity range between 12 and 16 PSU. Our calculations in the Bornholm Channel are in good accordance with these values (mean inflow 25,900 m^3/s and mean salinity 14.0 PSU), despite that we used time-averaged data instead of directly measured values which were used in Stigebrandt's calculations.

About 1/3 of the deep-layer flow out from the Southern Gotland Basin was estimated to occur at the intermediate salinity interval of 9–10 PSU. It illustrates that direct ventilation of the intermediate layer (the halocline) is important to the deep water renewal of the Baltic Sea. Deep thermohaline anomalies in the intermediate layer of the Eastern Gotland Basin during autumn and winter seasons in 1986–1991 were experimentally observed by Kõuts et al. (1991). The water was

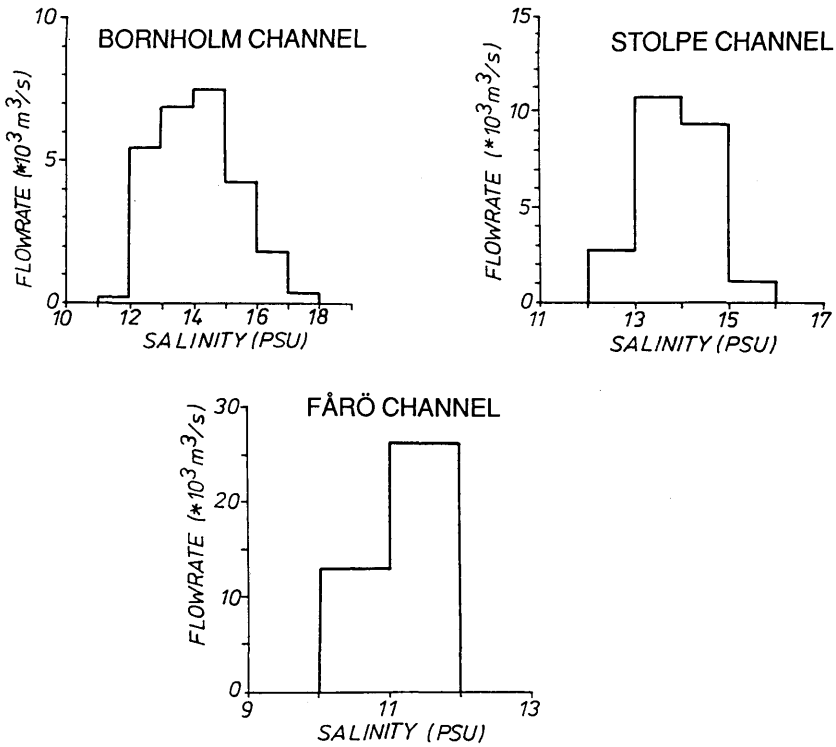

Fig. 7. Calculated salinity distribution for the geostrophic outflow rates in the different channels.

intruded into the halocline layer between 90 and 140 m in the Eastern Gotland Basin, and the authors argued that the water was coming from the Arkona or the Bornholm Basin.

4.4. Seasonal variability

In this section, the seasonal variability of salinities, halocline depths and geostrophic flow rates are examined.

The mean values for each month of the year were calculated on the basis of the data from the period 1970–1990. The results are presented in Fig. 8. From the figure one observes that the seasonal variability of surface layer salinities was 5 to 7 times weaker than in lower layer salinities and also that the seasonal variations were higher in the southern Baltic compared to those in the Central Baltic. For mean seasonal variability see also Matthäus (1984). In the Arkona Basin lower deep water salinities were observed from May to August. This is probably due to a decrease in the saline water transport from the Kattegat because of weaker western winds during summer time, also pointed out by Jacobsen (1980). In the Bornholm Basin the deep water salinity showed a weaker but clear seasonal cycle with higher values from March to September and with its minimum in November,

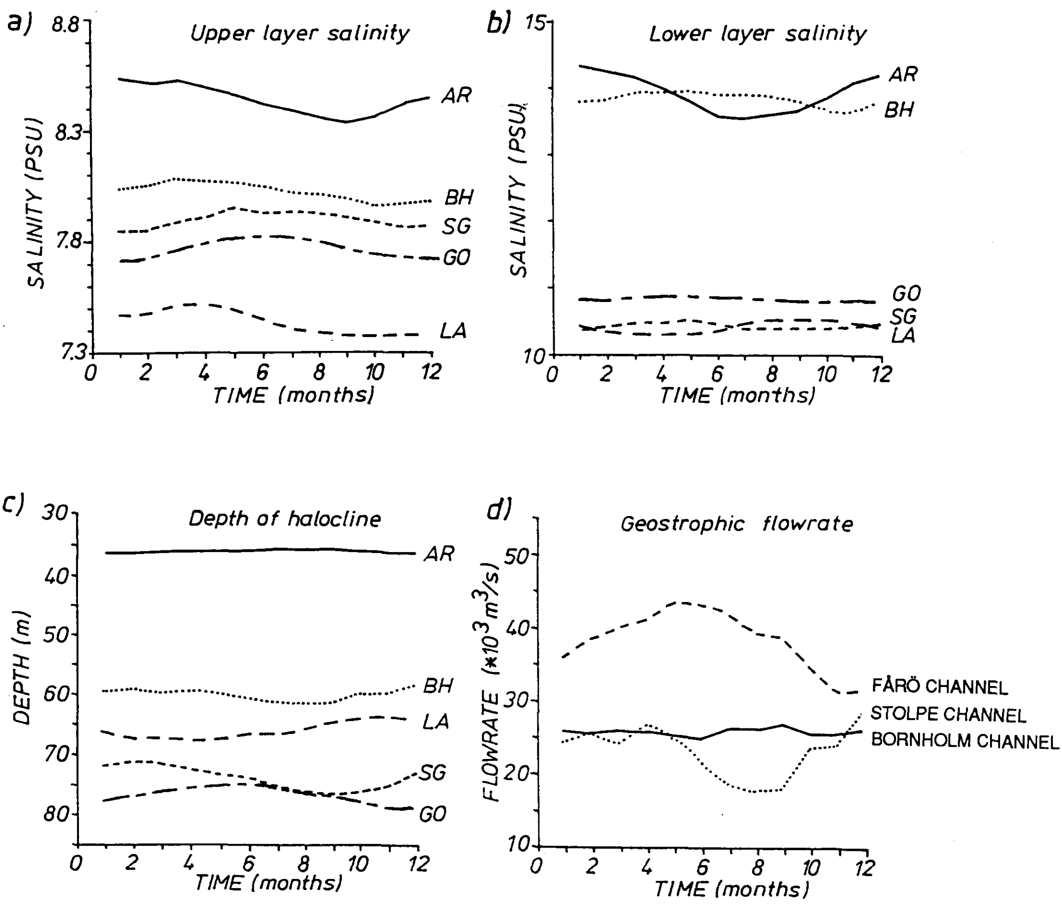

Fig. 8. Time series of monthly mean properties showing the seasonal variability in the different sub-basins. Salinity of upper layer (a) and lower layer (b), halocline depths (c) and deep water flow rate (d). Caption of sub-basins is the same as in Fig. 4. Data from the Southern Gotland Basin (SG) are also added.

which probably is caused by low summer salinities in the Arkona Basin. An increase of the salinity of the deep layer from February to June in the Southern Gotland Basin can also be noticed.

The halocline depth in the Arkona Basin does not show any seasonal cycle. The reason is probably due to the fact that the vertical observational interval was too large for such variability to be detected. Seasonal variations were, however, clearly present in the other sub-basins. In the Bornholm Basin and the Southern Gotland Basin, the halocline depth increases from June to September and decreases after October. This indicates increased inflow events starting in the autumn.

The geostrophic calculations show that the flow out from the Arkona Basin through the Bornholm Channel occurs with a typical mean flow rate of about 25,900 m^3/s and without larger seasonal variability. The seasonal variability in the geostrophic flow rates are more pronounced in the Stolpe Channel. The decrease of deep water outflow from the Bornholm Basin during the summer months is also confirmed in the TS plots, discussed in Subsection 4.2. The seasonal cycle of the geostrophic outflow from the Eastern Gotland Basin shows increasing values during the first half of the year and decreasing values from June to December.

5. Summary and conclusions

In the present work, the deep-water exchange and the mixing properties in the Baltic Proper were analysed on the basis of temperature and salinity data from a 20-year period, 1970–1990. Data from representative oceanographic stations from the sub-basins (the Kattegat, the Arkona Basin, the Bornholm Basin, the Eastern Gotland Basin and the Landsort Deep) were used. All data were taken from measured standard depths and the spline method was applied to create continuous profiles.

The data were analysed on the basis of the conservation principles, the two-layer approach and the geostrophic flow assumptions.

The mean flow rates and entrainment into a deep current were first calculated using conservation principles and the two-layer approach. Based upon these calculations the main mixing zones for the deep water inflow were localized. The temperature, salinity and density structure were then examined on the basis of a water mass analysis. The transformation of the deep water properties downstream in the sub-basins illustrated the importance of considering entrainment processes.

The variability of the deep water flow rates between different sub-basins was calculated on the basis of a geostrophic flow model. Salinity profiles from the period 1970–1990 were used as inputs. The model was calibrated using the 20-year mean flows and introducing effective sill depths instead of geometric ones. On the basis of the geostrophic model a group major inflow events between 1975 and 1979 were also analysed. The distribution of the inflowing water masses among different salinities was calculated, and the seasonal variations of flow rates, salinities and halocline depth were examined.

The conclusions may be summarized as follows.

(1) The renewal of the deep water in the Baltic Proper is through inflows from upstream basins. The inflowing dense water is diluted by surface water and on a 20-year average increased by a factor of four, when going from the Kattegat to the Landsort Deep.

(2) Three main mixing zones were localized. The first was the Belt Sea and the Sound, where the deep water inflow increased by 79 %. The second was the Arkona Basin, where the entrainment added 53 % to the deep water inflow. The third was the Stolpe Channel, where the entrainment added 28 % to the deep volume flow. The flow through the Bornholm Channel was found to pass almost without mixing with surrounding water, and in the Fårö Channel the entrainment added about 14 % to the deep water flow.

(3) The effective sill depths in the Bornholm Channel, the Stolpe Channel and the Fårö Channel were calculated to 48, 71 and 92 m, respectively. For the Bornholm Channel this is equal to the geometrical sill depth, which shows that a geostrophic model gives good representation of the deep flow. The differences between the effective and the geometrical sill depths were 15–25 % at the other sill areas. This indicates that the Stolpe Channel volume flow is more intensive and that the Fårö Channel volume flow is weaker than described by the geostrophic flow model.

(4) The typical time delay for an inflow from the Bornholm Channel to be observed in the Stolpe and the Fårö Channel was estimated from a group of major inflow events during 1975–1979 to be about 6 months and 1 year, respectively.

(5) The salinity of the inflowing deep water ranged from 11 to 18 PSU in the Bornholm Channel with a mean inflow rate of 25,900 m^3/s to 10 to 12 PSU in the Fårö Channel with a mean flow rate of 37,500 m^3/s.

6. Acknowledgements

This work was made during Tarmo Kõuts' scholarship from the Swedish Institute. The authors would like to thank Anders Stigebrandt, Bertil Håkansson, Lars Rahm and Eleonor Marmefelt for valuable discussions. Per Sandén and Jan Szaron who made data available and Vera Kuylenstierna who typed the manuscript are also gratefully acknowledged.

REFERENCES

Fennel, W., Sturm, M. and Helm, R. 1987. Response patterns of the Western Baltic. *Beitr. Meereskd.* (Berlin) *57*, 3–10.

Fonselius, S. H. 1969. *Hydrography of the Baltic deep basins.* Part III. Fishery Board of Sweden, Ser. Hydrogr., 23, 97 pp.

Gidhagen, L. and Håkansson, B. 1992. A model of the deep water flow into the Baltic Sea. *Tellus 44A*, 414–424.

Gill, A. E. 1982. *Atmosphere-ocean dynamics.* Academic Press, Inc., International Geophysics series, Volume 30, Appendix A3.

Henning, D. 1988. Evaporation, water and heat balance of the Baltic Sea: Estimates of short and long-term monthly totals. *Meteor. Rundsch. 41*, 33–53.

Jacobsen, T. S. 1980. Sea water exchange of the Baltic: Measurements and methods. The National Agency of Environmental Protection, Denmark, 106 pp.

Knudsen, M. 1900. Ein hydrographischer Lehrsatz. *Ann. der Hydrographie und Maritimen Met.*, 316–320.

Krauss, W. and Brügge, B. 1991. Wind-produced water exchange between the deep basins of the Baltic Sea. *J. Phys. Oceanogr. 21*, 373–384.

Kõuts, T., Elken, J. and Lips, U. 1991. Late autumn intensification of deep thermohaline anomalies and formation of lenses in the Gotland Deep. *Proceedings of 17th CBO*, Norrköping 1990, SMHI, S-601 76 Norrköping, Sweden.

Lundberg, P. 1983. On the mechanics of the deep-water flow in the Bornholm Channel. *Tellus 35A*, 149–158.

Marmefelt, E. and Omstedt, A. 1992. Deep water properties in the Gulf of Bothnia. *Continental Shelf Research 13*, 169–187.

Matthäus, W. 1984. Climatic and seasonal variability of oceanological parameters in the Baltic Sea. *Beitr. Meereskd.* (Berlin) *51*, 29–49.

Matthäus, W. 1985. Mittlere jahreszeitliche und langzeitliche Veränderlichkeit des Salzgehaltes im Arkonabecken. *Beitr. Meereskd.* (Berlin) *53*, 17–26.

Matthäus, W. 1990. Mixing across the primary Baltic halocline. *Beitr. Meereskd.* (Berlin) *61*, 21–31.

Matthäus, W. and Franck, H. 1992. Characteristics of major Baltic inflows, a statistical analysis. *Continental Shelf Research 12*, 1375–1400.

Mikulski, Z. 1986. Inflow from drainage basin. In: *Water Balance of the Baltic Sea*, Ch. 4. Baltic Marine Environmental Protection Commission, Helsinki Commission, Baltic Sea Environmental Proceedings, No. 16, 174 pp.

Omstedt, A. 1987. Water cooling in the entrance of the Baltic Sea. *Tellus 39A*, 254–265.

Omstedt, A. 1990. Modelling the Baltic Sea as 13 subbasins with vertical resolution. *Tellus 42A*, 286–301.

Pedersen, F. B. 1977. On dense bottom currents in the Baltic deep water. *Nordic Hydrology 8*, 297–316.

Petrén, O. and Walin, G. 1976. Some observations of the deep flow in the Bornholm strait during the period June 1973–December 1974. *Tellus 28*, 74–87.

Rahm, L. 1985. On the diffusive salt flux of the Baltic Proper. *Tellus 37A*, 87–96.

Rahm, L. 1988. A note on some hydrographic properties of the Eastern Gotland Basin. *Beitr. Meereskd.* (Berlin) *58*, 47–58.

Rahm, L. and Wulff, F. 1992. Using parallel salinity and temperature profiles for calculations of estuarine fluxes with reference to the Baltic Proper. *Netherlands Journal of Sea Research 29*, 281–289.

Reinsch, Ch. 1967. Smoothing by spline functions. *Num. Math. 10*, 177–183.

Rydberg, L. 1980. Rotating hydraulics in deep-water channel flow. *Tellus 32*, 77–89.

Stigebrandt, A. 1983. A model for the exchange of water and salt between the Baltic and the Skagerrak. *J. Phys. Oceanogr. 13*, 411–427.

Stigebrandt, A. 1987a. A model for the vertical circulation of the Baltic deep water. *J. Phys. Oceanogr. 17*, 1772–1785.

Stigebrandt, A. 1987b. Computations of the flow of dense water into the Baltic Sea from hydrographical measurements in the Arkona Basin. *Tellus 39A*, 170–177.

Sturm, M. 1986. On the formation of intermediate water mass structures in the nearshore zone of the western Baltic Sea. *Beitr. Meereskd.* (Berlin) *55*, 91–93.

Walin, G. 1977. A theoretical frame-work for the description of estuaries. *Tellus 29*, 128–136.

Walin, G. 1981. On the deep water flow into the Baltic. *Geophysica 17*, 75–93.