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Water cooling in the entrance of the Baltic Sea

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ABSTRACT

Water cooling in the entrance of the Baltic Sea is analyzed using measured and numerically simulated data. The measurements cover two cooling periods from the prelude of the severe sea ice winters 1984/85 and 1985/86, during which the surface water was cooled from 10°C to freezing. The calculations are based upon a mathematical model which divides the entrance of the Baltic Sea into four sub-basins: the Kattegat, the Belt Sea, Öresund and the Arkona Basin. In each sub-basin, the conservation equations for momentum, heat and salt in the vertical dimension are solved with vertical mean velocities based upon parameterized in- and outflows. The turbulent exchange coefficients are calculated using a two-equation model of turbulence.

The measured data demonstrate the importance of considering the exchange through the entrance of the Baltic Sea when modelling autumn cooling. The mathematical model describes the exchange and the general development of the cooling in a satisfactory way.

1. Introduction

Severe sea ice conditions along the Swedish west coast occur roughly on a 4- to 5-year cycle. During the last 50 years, severe conditions occurred in the following years: 1939/40, 1940/41, 1941/42, 1946/47, 1954/55, 1955/56, 1962/63, 1965/66, 1969/70, 1978/79, 1984/85 and 1985/86.

In the entrance of the Baltic Sea, the frequency of ice in the middle of February is between 25 and 50% (Swedish Administration for Shipping and Navigation, 1982). The sea ice causes serious transportation problems, and accurate forecasts of the time for ice formation is thus of great importance for shipping and icebreaking services. The purpose of the present study is to formulate and explore a mathematical model for water cooling in the entrance of the Baltic Sea. The study extends the model by Omstedt et al. (1983) by considering the effects of in- and outflows between the Skagerrak and the Baltic Sea. The entrance is modelled as 4 coupled sub-basins, the Kattegat, the Belt Sea, Öresund and the Arkona Basin (Fig. 1). The modelling of the coupling between the sub-basins follows that by Stigebrandt (1983) and Jacobsen (1980).

The dynamics in each sub-basin are treated as transient turbulent flows which are assumed to be horizontally homogeneous. The turbulent mixing is modelled by considering one equation for the turbulent kinetic energy and one for the dissipation rate of turbulent kinetic energy. In the Kattegat and the Arkona Basin, a transient Ekman boundary layer theory is assumed, while a transient channel boundary layer theory is assumed in the Belt Sea and in Öresund. The entrainment in the Kattegat and the Arkona Basin is thus induced by winds. In the Belt Sea and in Öresund, the entrainment is induced both by currents and winds.

In the entrance of the Baltic Sea, the density structure may vary from a thin brackish surface layer during calm outflow events to a deeper and more saline surface layer during stormy inflow events. These aspects of the dynamics are modelled by considering the conservation equations for heat and salt in each basin and the exchange of properties between the basins. The net heat loss to the atmosphere is calculated for each sub-basin using bulk formulas for sensible heat, latent heat and radiation fluxes.

The model is explored by analyzing measured

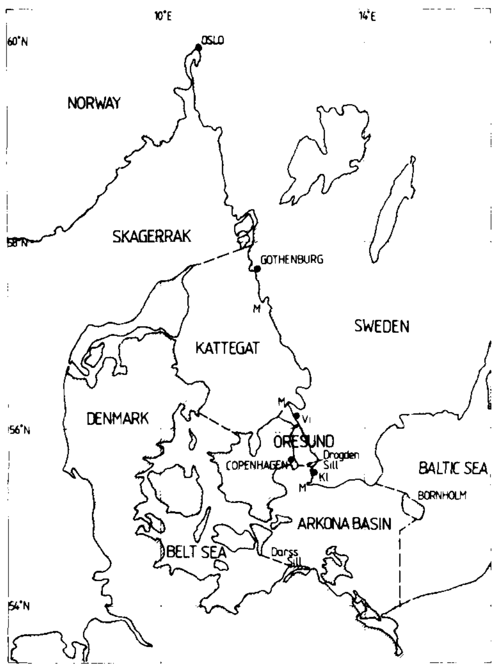

Fig. 1. Map of the entrance to the Baltic Sea. The dot indicates the site of the measurements. Vi = Viken, Kl = Klagshamn, M = Meteorological station.

data from 2 events. The first one is from December 1984 to January 1985, and the second one from November 1985 to February 1986. The comparison with field data will serve as a guide to the applicability of the model.

In the following section, a general description of the area studied is given. Then, in Section 3, the mathematical formulation will be presented together with some considerations of applicability. In Section 4 details about the calculations are given, and in Section 5 the results from measured and calculated data are presented. Finally, a summary and some conclusions are given in Section 6.

2. General description of the area studied

The Skagerrak–Baltic Sea system can be regarded as a large estuary, consisting of a sequence of sub-basins. The connection between the Skagerrak and the Baltic Sea is through rather narrow and shallow sill areas, the Kattegat, the Belt Sea and Öresund. On the annual average, the fresh water supply to the Baltic Sea and a corresponding amount of saline water from the Skagerrak are balanced by a brackish surface water outflow from the Baltic Sea. However, superimposed on the estuarine circulation, barotropic flows are also present, caused by the meteorological forcing.

Review papers on the topographical and hydrographical conditions may be found in Svansson (1975, 1984), Jacobsen (1980) and Kullenberg (1981). The Kattegat is a shallow and highly stratified sea. The stratification is caused by brackish water of Baltic Sea origin, on top of deeper saline layers of Skagerrak origin. The connection to the Baltic Sea goes through the Belt Sea and Öresund. In the Belt Sea, two waterways exist: the Great Belt and the Little Belt. The former is the most important one for the deep water inflow into the Baltic Sea. The main sill is in the southern part of the Belt Sea, at the Darss Sill, with a sill depth of 18 m. In Öresund, the sill depth is 8 m at the Drogden Sill.

The Arkona Basin is the westernmost part of the Baltic Sea and belongs hydrographically to the Baltic Sea. The basin is influenced by the inflowing dense water, which creates a rather shallow halocline compared to the rest of the Baltic Sea (Matthäus, 1985). The deep water flow from the Arkona Basin to the next Baltic Sea sub-basin, the Bornholm Basin, goes through a channel north of Bornholm, the Bornholm Channel.

During severe winters, ice formation starts in Öresund and the Belt Sea. In the middle of January, the first ice from the Kattegat is reported. Later, ice forms in the Arkona Basin and the Skagerrak. The maximum ice extent is generally reached in the middle of February, with level ice thicknesses of about 0.1–0.3 m. The sea ice is easily moved by winds and currents, causing rafted and ridged ice. The break-up of the sea ice cover normally starts with ice drift towards the north and with Kattegat ice drifting out into the Skagerrak. By the beginning of April most of the sea ice has melted.

Fig. 2 shows a chart with different ice types as well as the sea surface water temperature. The chart illustrates the maximum sea ice extent during the winter 1984/85. Two particular features can be noticed. First, the ice zone between the warm surface layer of Skagerrak and

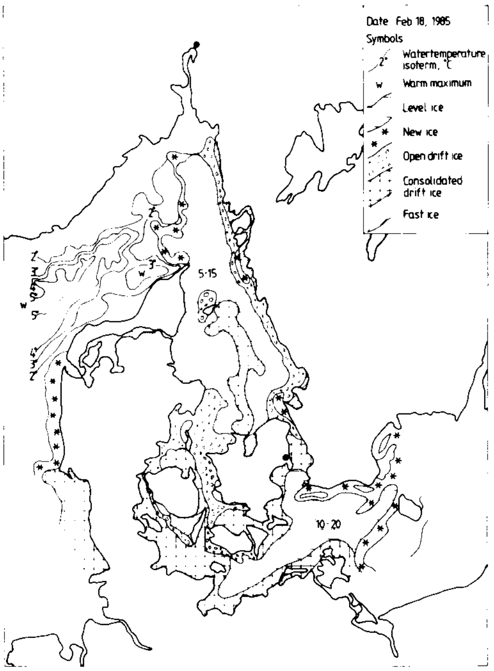

Fig. 2. The sea ice distribution in February 18, 1985, with sea water temperatures analyzed from the NOAA-6 weather satellite.

the ice-covered Kattegat; second, the ice zone between the ice-covered Arkona Basin and the open Baltic Sea east of Bornholm (the Bornholm Basin). The position of the ice zones can be taken as good evidence of the importance of the hydrographic state in the different sub-basins for cooling and sea ice formation. These features also indicate how a proper division should be made when modelling the thermal regime of the Baltic Sea.

3. Theory

3.1. Basic assumptions

The approach adopted is to divide the entrance of the Baltic Sea into sub-basins, which are assumed to be horizontally homogeneous. The dynamics in each sub-basin is treated as transient turbulent flows, with one space coordinate in the vertical direction and with one time coordinate. The sub-basins are coupled to each other by the exchange of volume, salt and heat. Properties from the Skagerrak and the Baltic Sea east of Bornholm are assumed to be deep Kattegat water or surface Arkona water properties, respectively. The inflowing water levels are calculated from the assumption that the inflow is spread out at its own density surface without mixing. The outflowing water levels are assumed to depend on the geometry and the layer from which water is exchanged. Fronts, selected withdrawal, and dense bottom water mixing are not dealt with.

In the Kattegat and the Arkona Basin, a transient Ekman boundary layer is assumed, while a transient channel boundary layer theory is assumed in the Belt Sea and in Öresund. The turbulent mixing is therefore assumed to be increased in the Belt Sea and in Öresund because of current shear associated with the barotropic exchange. The heat loss to the atmosphere is calculated using bulk formulas for sensible heat, latent heat and radiation fluxes. The meteorological input parameters are air temperature, relative humidity, wind velocity and mean cloudiness.

The barotropic flows, associated with the sea level changes, are calculated using observed sea level data from the Kattegat and the fresh water supply to the Baltic Sea. The baroclinic flows, associated with the estuarine circulation, are calculated assuming steady, geostrophic and hydrostatic flows at the Skagerrak-Kattegat boundary, at the Darss Sill and in the Bornholm Channel. The vertical velocities and the sea level variations are calculated from the prescribed in- and outflows, using the continuity equation. A sketch of the model is given in Fig. 3.

The model employed extends the work by Omstedt et al. (1983) by introducing water exchange between sub-basins with associated mean vertical velocities and moving free surfaces. The reader is referred to this paper for further details about the basic assumptions, etc. Further applications of the model may also be found in Omstedt (1983, 1984).

3.2. Mean flow equations

Within the assumptions made, the heat and salinity equations are:

$$\frac{\partial(\rho C_p T)}{\partial t} + \frac{W \partial(\rho C_p T)}{\partial z} = \frac{\partial}{\partial z}\left(\frac{\nu_T}{\sigma_T}\frac{\partial(\rho C_p T)}{\partial z}\right), \quad (1)$$

$$\frac{\partial S}{\partial t} + \frac{W \partial S}{\partial z} = \frac{\partial}{\partial z}\left(\frac{\nu_T}{\sigma_S}\frac{\partial S}{\partial z}\right), \quad (2)$$

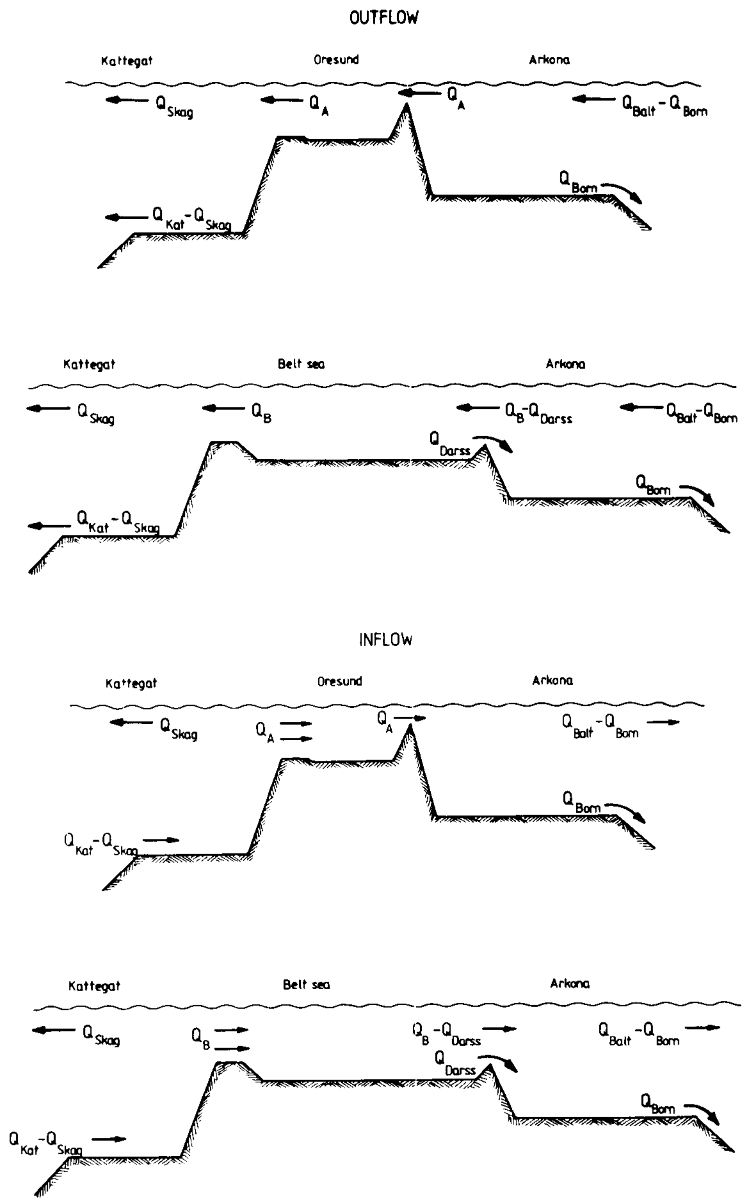

Fig. 3. Sketch showing the basic flows.

where z is the vertical space coordinate, t time coordinate, T mean temperature, S mean salinity, W mean vertical velocity, ρ water density, C_p specific heat of water, ν_T kinematic eddy viscosity, σ_T and σ_S turbulent Prandtl and Schmidt numbers. The basic boundary layer equations in the different basins considered are:

In the Kattegat and the Arkona Basin

$$\frac{\partial \rho U}{\partial t} + \frac{W \partial \rho U}{\partial z} = \frac{\partial}{\partial z}\left(\nu_T \frac{\partial \rho U}{\partial z}\right) + f\rho V \qquad (3)$$

$$\frac{\partial \rho V}{\partial t} + \frac{W \partial \rho V}{\partial z} = \frac{\partial}{\partial z}\left(\nu_T \frac{\partial \rho V}{\partial z}\right) - f\rho U \qquad (4)$$

In the Belt Sea and Öresund

$$\frac{\partial \rho V}{\partial t} + \frac{W \partial \rho V}{\partial z} = \frac{\partial}{\partial z}\left(\nu_{\mathrm{T}} \frac{\partial \rho V}{\partial z}\right) - \rho g \frac{\partial \zeta}{\partial y}, \tag{5}$$

where U and V are mean velocities in horizontal directions, f Coriolis' parameter, g constant of gravity and $\partial\zeta/\partial y$ the sea surface tilt between the Arkona Basin and the Kattegat.

An approximate form of the equation of state is used:

$$\rho = \rho_0(1 - \alpha(T - T_{\mathrm{M}})^2 + \beta S), \tag{6}$$

where α and β are constants and T_{M} is the temperature of maximum density. In the present study α and β are put to $7 \cdot 10^{-6}$ and $8 \cdot 10^{-4}$, respectively. The temperature of maximum density is treated as a function of salinity according to Caldwell (1978), pressure effects are not dealt with.

The boundary conditions for the mean flow equations at the sea surface are:

$$\frac{\nu_{\mathrm{T}}}{\sigma_{\mathrm{T}}} \frac{\partial T}{\partial z} = F_{\mathrm{N}}(t)(\rho_0 C_{\mathrm{p}})^{-1},$$

$$\frac{\nu_{\mathrm{T}}}{\sigma_{\mathrm{S}}} \frac{\partial S}{\partial z} = 0,$$

$$\nu_{\mathrm{T}} \frac{\partial U}{\partial z} = -\tau_x(t)\rho_0^{-1},$$

$$\nu_{\mathrm{T}} \frac{\partial V}{\partial z} = -\tau_y(t)\rho_0^{-1},$$

where ρ_0 is a water reference density equal to 10^3 (kg m^{-3}), $\tau_x(t)$ and $\tau_y(t)$ wind stresses in the x- and y-directions, and $F_{\mathrm{N}}(t)$ the net heat loss.

At the lower boundary, a zero flux condition is used for all variables, except for the velocities, where a zero velocity condition is given.

3.3. *Turbulence model*

The turbulence model is based upon exchange coefficients calculated with a two-equation model of turbulence, one equation for the turbulent kinetic energy, k, and another for the dissipation rate of turbulent kinetic energy, ε. The equations allow for the conversion of turbulent kinetic energy into potential energy, a key feature of the mixing in the upper layers of the ocean. The generation of turbulence is calculated from shear production. Buoyancy production generated by cooling is included, but of minor importance. In the Arkona Basin and the Kattegat, the shear production is calculated assuming an Ekman boundary layer flow, eq. (3) and (4). The mixing is therefore associated with the wind stress and the earth rotation. In the Belt Sea and Öresund, the shear is calculated assuming a channel boundary layer flow, eq. (5). The mixing is therefore associated with the barotropic currents and the bottom friction. Local wind-mixing is also included in the Belt Sea and Öresund. The destruction of turbulence is calculated from the dissipation rate and the buoyancy.

The turbulence equations read:

$$\frac{\partial k}{\partial t} + \frac{W \partial k}{\partial z} = \frac{\partial}{\partial z}\left(\frac{\nu_{\mathrm{T}}}{\sigma_{\mathrm{k}}} \frac{\partial k}{\partial z}\right) + P_{\mathrm{s}} + P_{\mathrm{b}} - \varepsilon, \tag{7}$$

$$\frac{\partial \varepsilon}{\partial t} + \frac{W \partial \varepsilon}{\partial z} = \frac{\partial}{\partial z}\left(\frac{\nu_{\mathrm{T}}}{\sigma_{\varepsilon}} \frac{\partial \varepsilon}{\partial z}\right) + \frac{\varepsilon}{k}(C_{1_\varepsilon} P_{\mathrm{s}} + C_{3_\varepsilon} P_{\mathrm{b}} - C_{2_\varepsilon} \varepsilon), \tag{8}$$

$$P_{\mathrm{s}} = \nu_{\mathrm{T}}\left[\left(\frac{\partial U}{\partial z}\right)^2 + \left(\frac{\partial V}{\partial z}\right)^2\right], \tag{9}$$

$$P_{\mathrm{b}} = \nu_{\mathrm{T}} g \left[-\frac{2\alpha(T - T_{\mathrm{M}})}{\sigma_{\mathrm{T}}} \frac{\partial T}{\partial z} + \frac{\beta}{\sigma_{\mathrm{s}}} \frac{\partial S}{\partial z}\right], \tag{10}$$

$$\nu_{\mathrm{T}} = C_\mu \frac{k^2}{\varepsilon}, \tag{11}$$

where P_{s} is the production due to shear, P_{b} is production/destruction due to buoyancy, σ_{k} and σ_ε turbulent Prandtl/Schmidt numbers, C_μ, C_{1_ε}, C_{2_ε} and C_{3_ε} are constants. The surface boundary conditions for turbulent kinetic energy, k, and its dissipation rate, ε, are related to the friction velocity and the net heat flux at the air/sea interface. At the lower boundary, a zero flux condition is used.

For a general description of the turbulence model, the reader is referred to Svensson (1979) or Rodi (1980).

3.4. *In- and outflow parameterization*

The sub-basins are coupled through the exchange between the basins (see Fig. 3). In the entrance of the Baltic Sea, the meteorologically forced barotropic flow often dominates the exchange. A simple model for this flow was presented by Stigebrandt (1980), and is used in the present paper. The barotropic flow is:

$$Q_{\mathrm{BAR}} = C \frac{z_{\mathrm{KAT}} - z_{\mathrm{BALT}}}{|z_{\mathrm{KAT}} - z_{\mathrm{BALT}}|^{1/2}} \quad [\mathrm{m^3\,s^{-1}}], \tag{12}$$

where $z_{KAT} - z_{BALT}$ is the sea water level difference between the Kattegat and the Baltic Sea and C is a geometrical factor. From 90 days integration, Stigebrandt (1980) calibrated the model giving a geometrical factor of $2.95 \cdot 10^5$.

To test eq. (12) further, a 4-year integration was performed by the present author (see Fig. 4). The modelling of the barotropic forcing from the sea was based upon observed sea level variations in the Kattegat (Viken, North 56°09′, East 12°34′). The river runoff to the Baltic Sea was modelled by a constant value of 10^4 m^3 s^{-1}, and the mean sea level of the Baltic Sea was represented by observations from Stockholm. The efficiency of eq. (12), given by the explained variance (R2) becomes 0.73, where R2 is defined according to:

$$\mathrm{R2} = \frac{\sum_{i=1}^{N} (z_{ST} - \bar{z}_{ST})_i^2 - \sum_{i=1}^{N} (z_{CAL} - z_{ST})_i^2}{\sum_{i=1}^{N} (z_{ST} - \bar{z}_{ST})_i^2},$$

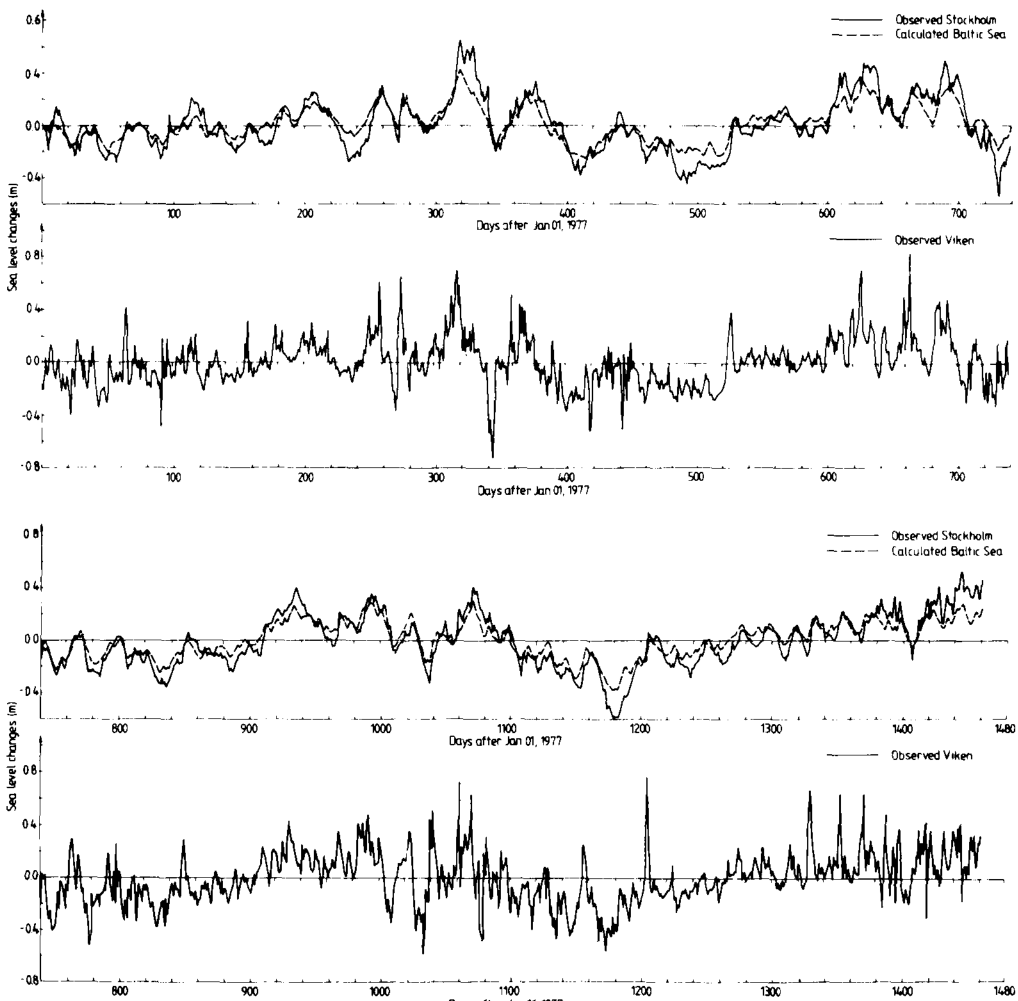

Fig. 4. A 4-year comparison between observed sea level changes in the Baltic Sea represented by Stockholm (fully drawn line) and calculated values (broken line). The forcing from the Kattegat is represented here by measurements from the water level station at Viken.

where z_{ST} represents the observed sea level at Stockholm, $\bar{z}_{ST}$ observed mean, z_{CAL} calculated mean level of the Baltic Sea and N the numbers of observation studied.

The agreement between observed and calculated mean sea levels in Fig. 4 illustrates the strength of eq. (12). The equation, however, only considers the net volume flow, and a division between the barotropic exchange through the Belt Sea and Öresund is thus needed. The division made in the present paper reads:

$$Q_A = 10^5 \frac{z_{KAT} - z_{BALT}}{|z_{KAT} - z_{BALT}|^{1/2}} \quad [m^3\ s^{-1}], \qquad (13)$$

$$Q_B = 1.95 \cdot 10^5 \frac{z_{KAT} - z_{BALT}}{|z_{KAT} - z_{BALT}|^{1/2}} \quad [m^3\ s^{-1}], \qquad (14)$$

where Q_A is the barotropic flow through Öresund and Q_B the barotropic flow through the Belt Sea. The geometrical factor, equal 10^5, in eq. (13) is in accordance with Jacobsen (1980), who gives the values of $9 \cdot 10^4$. Thus, about 1/3 of the exchange is assumed to pass Öresund, while 2/3 pass through the Belt Sea.

In the Kattegat, the water level is represented by observed data from Viken and the corresponding volume flux is calculated according to:

$$Q_{KAT} = A_{KAT} \frac{\Delta z_{VI}}{\Delta t} \quad [m^3\ s^{-1}], \qquad (15)$$

where A_{KAT} is the Kattegat surface area and $\Delta z_{VI}/\Delta t$ the water level change at Viken.

The calculations of the water-level variations in the Arkona Basin are based upon the barotropic flows and the corresponding volume flux reads:

$$Q_{BALT} = (Q_A + Q_B + Q_F)\left(1 - \frac{A_{ARK}}{A_{BALT}}\right) \quad [m^3\ s^{-1}], \qquad (16)$$

where Q_F is the fresh water supply to the Baltic Sea and A_{ARK}/A_{BALT} is the ratio between the surface areas of the Arkona Basin and the Baltic Sea. The fresh water supply, Q_F, is put at a constant value of 10^4 $m^3\ s^{-1}$ and is always treated as an inflow to the Arkona Basin.

Eq. (16) indicates that the water level variations in the Arkona Basin are treated as the water level variations in the Baltic Sea. For short time variations—less than one day—this is, of course, not realistic. On longer time scales—days to weeks—this is, however, a more reasonable assumption. Recalling Fig. 4 one can observe that the main volume changes are associated with a longer time scale—from weeks to a month. This fact supports the present simplification.

Superimposed on the barotropic exchange, the estuarine circulation causes baroclinic flows. In fjords with wide sounds, wider than the internal Rossby radius, steady baroclinic flows form concentrated coastal or bottom currents. From the assumption of a steady, geostrophic, hydrostatic flow and with the surrounding fluid at rest, Stigebrandt (1983) proposed that the surface water flowing out to the Skagerrak, Q_{SKAG}, and the deep water flowing in over Darss Sill, Q_{DARRS}, should be parameterized according to:

$$Q_{SKAG} = g\beta \frac{(S_{2_{KAT}} - S_{1_{KAT}})}{2f} H_{1_{KAT}}^2 \quad [m^3\ s^{-1}] \qquad (17)$$

$$Q_{DARSS} = g\beta \frac{(S_{2_{BELT}} - S_{1_{BELT}})}{2f} (H_{DARSS} - H_{1_{BELT}})^2 \quad [m^3\ s^{-1}], \qquad (18)$$

where indexes 1 and 2 represent the surface and the deep water layer, respectively, S is salinity, $H_{1_{KAT}}$ surface layer thickness in the Kattegat, $H_{1_{BELT}}$ surface layer thickness in the Belt and H_{DARSS} the sill depth at Darss.

The deep water flow out of the Arkona Basin through the Bornholm Channel, Q_{BORN}, is also parameterized by assuming a steady, geostrophic and hydrostatic flow. The flow reads:

$$Q_{BORN} = g\beta \frac{(S_{2_{ARK}} - S_{1_{ARK}})}{2f} (H_{BORN} - H_{1_{ARK}})^2 \quad [m^3\ s^{-1}], \qquad (19)$$

where H_{BORN} is the sill depth in the Bornholm Channel and $H_{1_{ARK}}$ the well mixed surface layer thickness in the Arkona Basin.

Eq. (19) states that the deep flow through the Bornholm Channel may, to a first approximation, be considered as rotationally-baroclinically controlled. This is supported by the fact that the Bornholm Channel is rather wide and that the downstream halocline is below the sill depth. See also the hydrographical observations presented by Petrén and Walin (1976). Frictional effects may be of importance, according to Lundberg

(1983), and reduce the flow. However, these effects are not dealt with.

Besides the exchange of volume, eqs. (13)–(19), the flows also exchange salt and heat. The flow of properties is associated with the assumed exchange levels. The outflowing water brings salt and heat from the upstream basin and stores it in the downstream basin. The outflow levels of brackish water are assumed to be fixed. In the case of inflowing Kattegat water to the Belt Sea and Öresund, the flow is modelled partly as surface and partly as deep Kattegat flow, see measurements from the northern Öresund (Cederlöf and Djurfelt, 1977). The inflow levels are calculated by considering the inflowing water density and assuming that all water will reach the corresponding density level without mixing.

Finally, it should be noted that water level data from one station alone, Viken, is needed as external forcing in the present parameterizations. This will, of course, reduce the complexity of the problem.

3.5. Consideration of applicability

Cooling in the ocean is often controlled by the net heat loss to the atmosphere and the depth of the well mixed surface layer. The importance of these two factors can easily be illustrated by considering the heat energy equation, eq. (1). Assuming no vertical velocity, the vertically integrated heat energy equation becomes:

$$\frac{\Delta \bar{T}}{\Delta t} \approx \frac{\Delta F}{\rho C_p D},$$

where $\Delta \bar{T}$ is the temperature change within the mixed surface layer D, Δt is time period considered and ΔF net heat loss through D.

With a net heat loss of 100 W m^{-2}, which is a typical value for autumn cooling conditions, one finds that a 5 m thick surface layer is cooled 4°C during 10 days, while a 50 m thick surface layer during the same period cools a mere 0.4°C. These factors are modelled by employing a transient Ekman or channel flow boundary layer theory, in which stratification caused by the exchange of salt and heat through the basins is considered, as well as the net heat loss at the ocean/atmosphere interface.

Horizontal inhomogeneities in the entrance area are treated by the assumption that the area can be characterized by 4 sub-basins. The division of the sub-basins considers natural boundaries caused by geometrical and dynamical restrictions. In the present problem, at least two time scales are of importance. First, the synoptic time associated with the meteorological forcing; and second, the residence time associated with the estuarine circulation. In the entrance of the Baltic Sea, the time scales can be estimated to days, and to years, respectively. This implies that the wind mixing and the net heat loss, as well as the barotropic forcing, should be treated as transient processes, while the baroclinic exchange may be regarded as steady. Also worth noticing is that the horizontal length scale of each sub-basin is about one order of magnitude smaller than typical mid-latitude synoptic weather systems.

These considerations indicate that the mathematical model may be suitable when modelling cooling in the entrance of the Baltic Sea. The comparison with field data in Section 5 will serve as a quantitative test of the applicability.

4. Details of calculations

Eqs. (1)–(11), (13)–(19), and the continuity equation determining the vertical velocities from the in- and outflows, form a closed system. The equations in their finite difference form, together with boundary conditions, were integrated forward in time, using an implicit scheme and a standard tri-diagonal matrix algorithm (Svensson, 1986). The time considered covers two periods: the first from December 1984 to January 1985, and the second from November 1985 to February 1986.

Air temperature, relative humidity, wind velocity and mean cloudiness where extracted every third hour from three synoptic weather stations at the Swedish west coast (Fig. 1). These weather data were horizontally averaged, and bulk formulas for the net heat loss as well as the wind stress were used in accordance with Omstedt et al. (1983). Daily mean water level data were extracted from one water level station, Viken, in the Kattegat.

During the two periods, a measuring system was moored in Öresund (see Fig. 1). The instrument consisted of a current meter, type Aanderaa, supplied with conductivity and temperature sensors. The current speed was

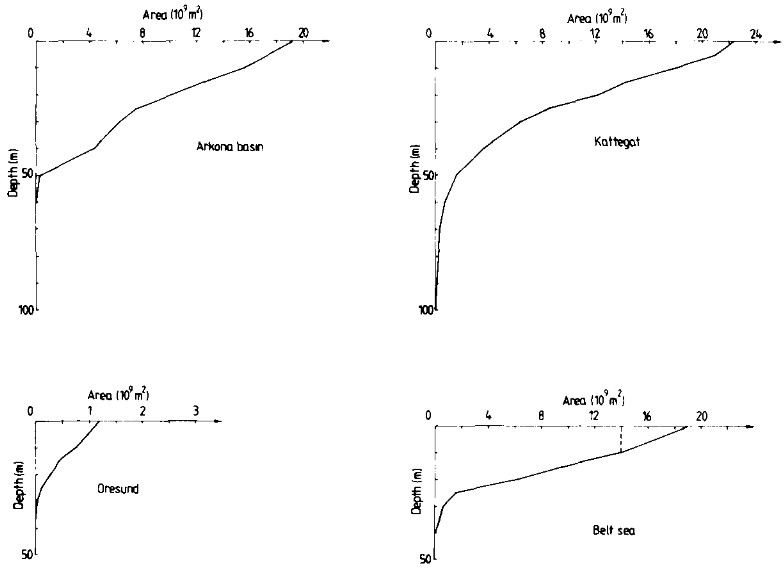

Fig. 5. The horizontal area vs depth for the sub-basins considered. The broken line in the Belt Sea indicates the distribution used in the calculation.

measured by use of a rotor, the temperature by a thermistor, the conductivity by an inductive type of conductivity sensor, and the current direction by a magnetic compass. The data were sampled from 5 m depth every 15 min and stored in a recording unit. The area/depth distribution in each sub-basin is calculated according to Ehlin et al. (1974). In the Belt Sea the distribution is slightly modified by neglecting the area/depth distribution in the shallow waters (Fig. 5).

5. Results and discussion

In this section, the results are presented. The first calculation to be considered is from December 1984 to January 1985. Fig. 6 shows measured and numerically simulated data from that period. Fig. 6a and 6b illustrate the sea level variations in the Kattegat and in the Arkona Basin. The calculated sea level variations in Kattegat follow those at Viken. The response in the Arkona Basin is given in Fig. 6b. The calculated values, which represent the mean sea levels of the Baltic Sea, are compared with measured data from the Arkona Basin (Klagshamn, North 55°31′, East 12°55′). The measured data show a larger fluctuation than

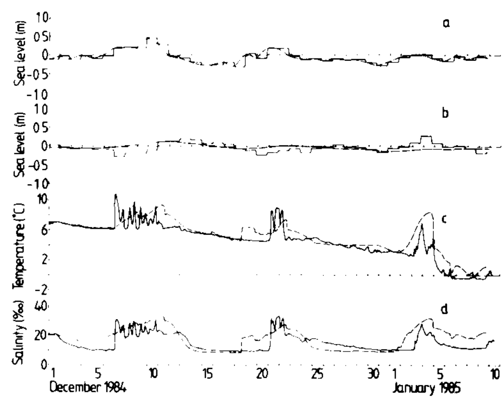

Fig. 6. Measured and calculated data from Kattegat (a), Arkona Basin (b) and from 5 m depth in Öresund (c and d). Fully drawn lines represent the measured data, while broken lines represent calculated values.

calculated values. This is what one can expect, as the water level variations of Klagshamn do not represent the mean sea level variations in the Baltic Sea. However, the general trends of rising and sinking mean sea levels are also observed at Klagshamn.

In Fig. 6d, both measured and simulated salinities show three distinct salt inflow events. These events are intrusions of saline Kattegat water, separated by periods with outflowing

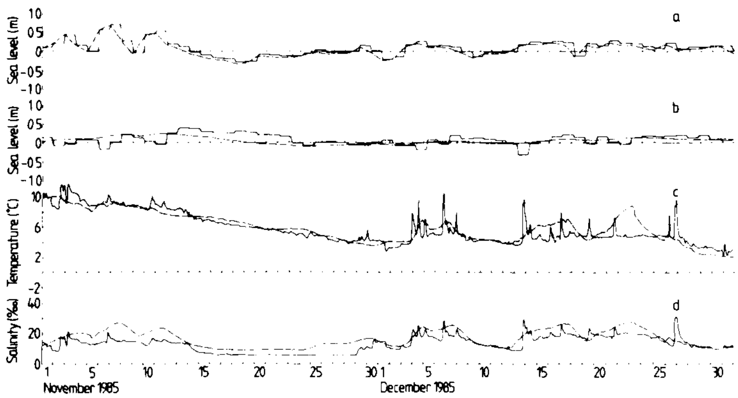

Fig. 7. Measured and calculated data from Kattegat (a), Arkona Basin (b) and from 5 m depth in Öresund (c and d). Fully drawn lines represent the measured data, while broken lines represent calculated values.

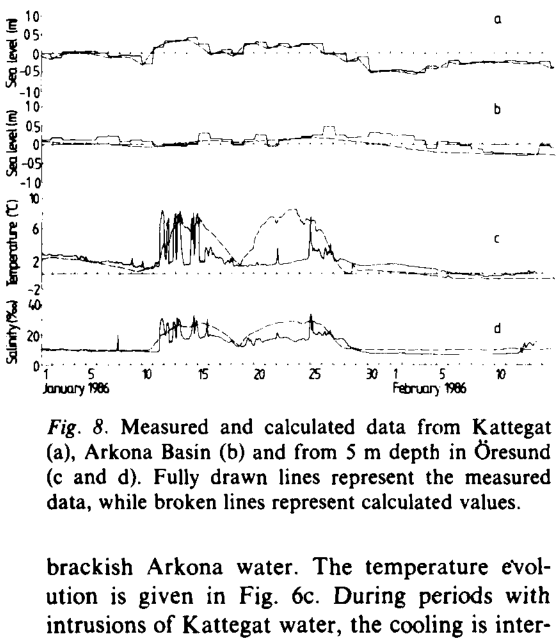

Fig. 8. Measured and calculated data from Kattegat (a), Arkona Basin (b) and from 5 m depth in Öresund (c and d). Fully drawn lines represent the measured data, while broken lines represent calculated values.

brackish Arkona water. The temperature evolution is given in Fig. 6c. During periods with intrusions of Kattegat water, the cooling is interrupted, whereas during periods with outflowing Arkona water, the cooling proceeds. From the time-temperature and time-salinity curves one can observe that the general development is reproduced in the model. To test the model further, a second calculation from November 1985 to February 1986 is considered below.

In Figs. 7 and 8, the results from measured and numerically simulated data from that period is given. Again, in Figs. 7b and 8b, one can observe how the measured water level variations in Klagshamn show more rapid changes, but they follow the general trends of rising and sinking mean sea levels of the Baltic Sea. In Figs. 7d and 8d the measured and calculated salinities are illustrated. The time-temperature development is shown in Fig. 7c and 8c. The measured temperatures illustrate again that the cooling is associated with outflowing water from the Arkona Basin, while inflowing water from the Kattegat interrupts cooling. In the calculations, only horizontal average values are considered and short-period oscillations are thus not dealt with. The measured data in Figs. 6, 7 and 8 illustrate the complex time-temperature evolution during autumn cooling in the entrance of the Baltic Sea. In the light of the simplifications introduced in the mathematical model, the results are most promising.

6. Summary and conclusions

The objective of the present work was to examine autumn cooling in the entrance of the Baltic Sea. For this purpose, a mathematical model was formulated and tested against measured data. The approach was to divide the entrance of the Baltic Sea into 4 sub-basins and to treat each sub-basin as horizontally homogeneous. In each basin, a transient boundary layer theory was assumed, and the sub-basins were coupled through parameterized in- and outflows. Also the net heat loss to the atmosphere was considered. The mathematical formulation has been tested against measurements from two periods: from December 1984 to January 1985 and from November 1985 to February 1986. The periods represent the prelude of two severe winters in the entrance of the Baltic.

The external parameters used in the model were observed weather data in the net heat loss and the wind stress calculations and observed sea level variations from one water level station in the barotropic flow calculations.

The conclusions may be summarized as follows.

(i) Measured temperatures and salinities from Öresund clearly demonstrate the importance of considering the exchange through the entrance of the Baltic Sea when modelling autumn cooling. The measurements illustrate that cooling is associated with outflowing brackish water from the Arkona Basin, while events with interrupted cooling are associated with periods of inflowing saline water from the Kattegat.

(ii) The mathematical model describes the exchange through the entrance of the Baltic Sea and the general development of the cooling in a satisfactory way.

(iii) Short-period fluctuations, less than 1 day, were measured particularly during inflow events. This effect was not dealt with in the model.

The results demonstrate that a one-dimensional boundary layer analysis can serve as a meaningful starting point when modelling surface water cooling, even in complex sea areas. The success of the analysis depends on how well the turbulent processes and the energy exchange at the water/air interface are modelled. Also, the subdivision of the area studied and the parameterization of the horizontal exchange are important. For these last two factors, consideration of geometrical and dynamical restrictions may be helpful.

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