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Sensitivity of Baltic Sea salinity to large perturbations in climate

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ABSTRACT: Potential future changes in the salinity of the Baltic Sea must be included in safety assessments of nuclear waste repositories. The salinity affects the potential recipient ecosystems, water turnover along the coast, as well as the groundwater hydrology and groundwater chemistry. The time-scales of nuclear waste repositories are extremely long, and climate forecasts must therefore encompass extreme variations. This study presents a model that enables computation of Baltic Sea salinity for different sea level positions and freshwater supplies, thus making it possible to assess the impact of a given climatic change. A modest rise in global sea level (+1 m) would lead to a salinity increase from 8 to 9 in the southern Baltic Proper. An increase in the freshwater supply by about 2000 $m^3 s^{-1}$ (approximately +10%) would result in a similar salinity change. Further, a sea level drop of about 5 m or an increase in the freshwater supply by a factor of 3 would reduce the salinity in the southern Baltic Proper to <1, i.e. large parts of the Baltic would become limnic. A 50% decrease in freshwater supply would double the salinity in the southern Baltic Proper to about 15; the effect is more pronounced in the Bothnian Sea and Bothnian Bay, where the salinity would increase from 6 to 13 and from 3.5 to 10, respectively.

KEY WORDS: Baltic Sea · Salinity · Climate · River runoff · Sea level

1. INTRODUCTION

Important background information for safety assessments of nuclear waste repositories adjacent to the Baltic Sea is the salinity under different conditions. The salinity affects groundwater hydrology, water turnover in the coastal zone, groundwater chemistry and the ecosystem. The first two influence spreading of radionuclides from a leaking repository. Corrosion of the canisters containing the radioactive waste may be influenced by changes in the groundwater chemistry. Finally, it is of interest to know the structure of the ecosystem that is the potential recipient of leaking radionuclides.

This study describes computations of salinity in the Baltic Sea that can be used to complement scenarios of future fate and effects of radionuclide exposures. It continues an investigation of salinity variations in the Baltic Sea during the past 8500 years (Gustafsson & Westman 2002). In that study, proxy data for salinity were compared with computations using a simple hydrodynamic model. That hydrodynamic model was found to be quite suitable for analyzing changes in Baltic Sea salinity on these time-scales, and here it is developed further to assess extreme values of Baltic Sea salinity that might occur until the next glaciation.

Specific climate scenarios for the period preceding the next glaciation are available (e.g. Morén & Påsse 2001), but they are not detailed enough to estimate the driving forces of Baltic Sea salt exchange, particularly with regard to freshwater supply. This study explores water circulation and equilibrium salinity in the Baltic Sea during periods of drastically different climatic conditions and MSL (mean sea level), which can be useful when discussing effects from different climate scenarios. In this paper the term MSL is used for the large-scale and long-term average sea level.

1.1. Oceanography of the Baltic Sea

Water exchange between the Baltic Sea and the Atlantic Ocean is limited by the shallow sills and by a

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Fig. 1. Overview of the Baltic Sea

large transitional area comprising the Danish Straits (Belt Sea and Öresund) and Kattegat; see Fig. 1. The Baltic Sea has a freshwater surplus of about 16 000 m^3 s^{-1} at present. The combined effect of limited exchange and high freshwater supply results in low salinities, ranging from about 12 in the deep waters of the Baltic Proper to about 3 in the surface waters of the Bothnian Bay. For the same reason the Baltic is permanently salt-stratified. The intermittency and variability of saline water inflows from the North Sea, and limited diffusive vertical exchange frequently lead to stagnation of the deep waters of the Baltic Proper, with oxygen depletion as a consequence.

The sensitivity of the salinity to changes in freshwater supply and oceanic exchange has been quantified by Stigebrandt (1983), Gustafsson (1997a,b, 2000b), Omstedt et al. (2000) and Gustafsson & Westman (2002). However, these investigations dealt with relatively small changes in environmental conditions. Rodhe & Winsor (2002) and Stigebrandt & Gustafsson (2003) have explored how large the freshwater supply would need to become to transform the Baltic Sea into a freshwater lake. Rodhe & Winsor (2002) found that the salinity would already become negligible at a 25% increase in freshwater supply, whereas Stigebrandt & Gustafsson (2003) found that the supply would have to increase by a factor of 4; the latter has been confirmed by Meier & Kauker (2003a).

The Baltic Sea is undergoing a gradual change due to postglacial uplift and sea level changes. About 8500 yr BP the Baltic became brackish when the Danish Straits opened and seawater intruded into the Baltic Basin. Since then Baltic Sea salinity increased from zero to 10–15 around 6000–5000 yr BP, and thereafter declined to 8–10 by 3000 yr BP. During the past 1500 yr the salinity seems to have been relatively constant. Gustafsson & Westman (2002) showed that the salinity variations are not only caused by postglacial uplift and sea level variations, but also by variations in freshwater supply, which contributed especially to the salinity maximum about 5000–6000 yr BP.

The present-day surface salinity is 7–8 in the Baltic Proper and less in the Bothnian Sea, Bothnian Bay and Gulf of Riga. The halocline of the Baltic Proper is located at 60 to 80 m depth and here the salinity increases to 11–13 (see Samuelsson & Stigebrandt 1996 for a discussion of salinity variations during the last 40 yr).

Inflows of salt water through the Danish Straits are the source of salt for the Baltic Sea, and since the halocline is located deeper than the sill depth, Baltic deep water cannot return directly through the Danish Straits. The flows through the Straits oscillate with an amplitude that is an order of magnitude larger that the average flow. This oscillating flow is the main carrier of salt to and from the Baltic Sea. The Belt Sea and Öresund are usually strongly stratified because of the fluxes of low salinity surface water from the Baltic Sea and of high saline deep water from the Kattegat.

The hydrography of the Kattegat is characterized by a sharp halocline at about 15 m depth dividing the surface mixed layer with a salinity of 15 to 25 from the deep water with a salinity of 32 to 35 (Svansson 1975). The deep-water layers are supplied from the north with Skagerrak water and they are successively entrained into the surface layer by wind generated turbulence. The Kattegat–Skagerrak front separates the surface waters of the seas. The position of the front may vary, but it is mostly heading from Skagen towards the northeast (Gustafsson & Stigebrandt 1996). In general, frontal dynamics force the outflow from the surface layer of the Kattegat to enter the Skagerrak along the Swedish coast.

Changes in vertical stratification of the Baltic Sea, and consequently changes in stagnation periods, can be anticipated from changes in freshwater supply and wind conditions. Gustafsson & Andersson (2001) described the forcing mechanisms of major water inflows to the Baltic Sea, and Stigebrandt (2003) recently discussed the mechanisms regulating the ventilation of the deep-water in the Baltic Proper, but it is difficult to model how the Baltic deep water will react to climate change; preliminary results have been presented by Gustafsson & Stigebrandt (2003), but further research is necessary to obtain reliable results.

2. MODEL DESCRIPTION

As the salinity of the Baltic Sea is sensitive to the freshwater supply as well as saltwater exchange with the ocean, water exchanges between the Baltic basins need to be coupled with a model of exchange with the North Sea. Stigebrandt & Gustafsson (2003) presented computations of the response of the horizontal salinity distribution to changing freshwater supply. They used a simpler model of salinity in the Kattegat than the one presented here, and they did not analyse the effects of changing MSL. The response of the horizontal salinity distribution to changed mixing and wind forced circulation within the Baltic Sea is not covered by the model.

Any long-term forecast remains uncertain since detailed information on future climate development is lacking. Therefore, we present a sensitivity analysis that gives the equilibrium Baltic Sea salinity for wide ranges of freshwater supply and MSL.

2.1. Barotropic water exchange

To analyse how topographic changes in the Danish Straits affect Baltic Sea salinity it is necessary to calculate the changes of the oscillating flow. This is done with a simple model of channel flow. The flow of homogeneous water through a channel is primarily forced by the pressure gradient caused by the difference in water level between the adjacent basins. During steady-flow conditions, the pressure gradient force is balanced by frictional forces at the walls and bottom of the channel and by large scale topographic drag due to acceleration in the narrowest sections (Stigebrandt 1980, 1992). The latter causes turbulence downstream of the constriction. In the case of a stratified channel there may be an additional pressure gradient due to differences in density between the basins on either side of the channel (Mattsson 1996) and additional resistance due to transfer of energy to internal waves (Stigebrandt 1999). These effects of stratification are detectable, but quite small, and have therefore been disregarded in this investigation.

Following e.g. Stigebrandt (1992) the current speed through a channel is

$$u|u| = \frac{g\Delta\eta}{\frac{1}{2}+C_\mathrm{d}L\frac{W+2H}{WH}} \tag{1}$$

where g is the acceleration of gravity, $\Delta\eta$ is the sea-level difference between the basins, C_d is a drag coefficient, and L is the length, W is the width and H is the depth of the channel. The water flow, Q, is computed by multiplying current velocity by the cross-sectional area, as Eq. (1) is already averaged across the channel.

$$Q = C_T\frac{\Delta\eta}{\sqrt{\Delta\eta}} \tag{2}$$

where the transmission coefficient C_T is given by

$$C_T = WH\sqrt{\frac{g}{\frac{1}{2}+C_\mathrm{d}L\frac{W+2H}{WH}}} \tag{3}$$

In the case of present-day flow across the Darss and Drogden Sills, the transmission coefficients have been determined empirically by tuning Eq. (2) using observations of volume flow and sea levels (Stigebrandt 1992, Mattsson 1995, 1996, Jakobsen et al. 1997, Jakobsen & Ottavi 1997, Carlsson 1998, Jakobsen & Trebuchet 2000). The most recent values for the Danish Straits are

$$C_T = \begin{cases} C_{TS} = 0.63\times10^5\,\mathrm{m}^{5/2}\,\mathrm{s}^{-1}\ \text{Öresund} \\ C_{TB} = 1.67\times10^5\,\mathrm{m}^{5/2}\,\mathrm{s}^{-1}\ \text{Belt Sea} \end{cases} \tag{4}$$

A comparison between the empirical values in Eq. (4) and the values obtained by inserting the dimensions of the channels into Eq. (3) reveals that (1) the frictional force is an order of magnitude greater than the resistance due to constriction, and (2) friction must be of a considerable magnitude over the whole length of the Straits, not only at the sill. Further, as the width of the Straits is much larger than the depth, and since frictional forces are dominant, Eq. (3) can be approximated with good accuracy by

$$C_T = WH\sqrt{\frac{gH}{C_\mathrm{d}L}} \tag{5}$$

By assuming the channel length and the drag coefficient to be constant, variations in the oscillating barotropic flow can be calculated from Eqs. (2) & (5) based on time series of topographic changes. For simplicity a time-varying topographic factor can be defined so that

$$C_T = \alpha C_{T0} \tag{6}$$

where C_{T0} is the present flow resistance (zero in subscript denotes present day conditions) and the topographic factor α is defined as

$$\alpha = \frac{WH^{3/2}}{W_0 H_0^{3/2}} \quad (7)$$

or

$$\alpha = \frac{A_C\sqrt{H}}{A_{C0}\sqrt{H_0}} \quad (8)$$

where A_C is the cross-section area and H is mean depth.

Changes in topography due to MSL change were computed from a new digital topographic chart with approximately 1 km resolution (compiled by T. Seifert, F. Tauber and B. Kayser, Baltic Sea Research Institute, Warnemuende, Germany; data and data description available from www.io-warnemuende.de). The effect on barotropic flow was computed similarly as in Gustafsson & Westman (2002), that is, the flow is given by

$$Q = (\alpha_S C_{TS} + \alpha_B C_{TB})\frac{\eta_K - \eta_B}{\sqrt{|\eta_K - \eta_B|}} \quad (9)$$

where η_B and η_K are the sea levels in the Baltic Sea and the Kattegat, respectively, C_{TS} and C_{TB} are transmission coefficients for the Öresund and Belt Sea for present-day topography (Eq. 4), and α_S and α_B are topographic factors that change with MSL according to Eq. (8). Thus, Eq. (9) enables separate computation of the flows through Öresund and Belt Sea.

When the MSL changes, the changing flow resistance will modify the effect on sea level in the Baltic Sea. Therefore, it is necessary to evaluate the net change in magnitude of the oscillating flow by using a simple time-dependent model of the Baltic sea level. Following Stigebrandt (1992), the Baltic sea level is given by

$$A_B \frac{d\eta_B}{dt} = Q + Q_F \quad (10)$$

where A_B is the surface area, η_B is the sea level and Q_F is the freshwater supply to the Baltic Sea. Q is given by Eq. (9). In this model we assume that the sea level that forces the flow is the same as the sea level given by Eq. (10). Thus, we disregard spatial sea level variations due to the wind regime and variations in atmospheric pressure over the Baltic Sea. Carlsson (1998) presented a more refined model that took these effects into account.

Variations of the surface area of the Baltic Sea must be taken into account as well. For MSL changes of less than ±10 m, a simple linear relation approximates the surface area (cf. Fig. 2).

$$A_B = \begin{cases} A_{B0}(1+0.013z) & z<0 \\ A_{B0}(1+0.01z) & z>0 \end{cases} \quad (11)$$

where A_{B0} is the present surface area of the Baltic Sea and z is the MSL.

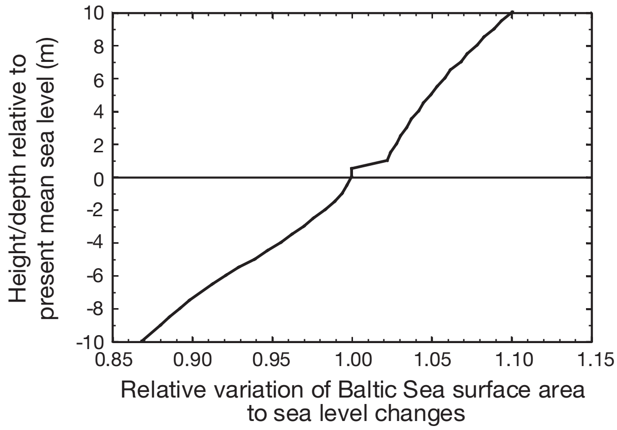

Fig. 2. Relative change of the surface area of the Baltic Sea due to changing global mean sea level (MSL)

The sea level model defined by Eqs. (9) to (11) is forced by an observed daily time-series of sea level in the Kattegat, for a length of time sufficient to provide accurate statistics of the flow through the Straits. Note that the results also vary with freshwater supply via Eq. (10), so computations must be repeated for each combination of MSL and freshwater supply change.

2.2. Steady-state salinity model

The model for the salinity of the Baltic Sea is a modification of the model in Gustafsson (1997b) and Gustafsson & Westman (2002). All parameter values are given in Table 1. The salinity model computes surface salinities in the Baltic Sea and the upper layer salinity and depth in the Kattegat (see Fig. 3).

As pointed out by Stigebrandt & Gustafsson (2003), the model in Gustafsson (1997b) and Gustafsson & Westman (2002) does not take into account the feedback on the variability of the flow through the Danish Straits from large changes in freshwater supply. In this study we directly compute average inflow (Q_{IN}) and outflows (Q_{OUT}) through the Danish Straits using the model described by Eqs. (9) & (10). The salt balance of the Baltic Sea can then be written as

$$S_0 = S_{IN}\frac{Q_{IN}}{Q_{OUT}} \quad (12)$$

where S_0 is the salinity of the outflowing Baltic water (i.e. the surface salinity of the southern Baltic Proper), S_{IN} is the salinity of the water flowing into the Baltic as described below. Daily volume flows are computed using the channel model described above (Eqs. 9 & 10). The water volume that enters the Baltic during a specific inflow event, V_i, can be integrated from the volume flow time series

Table 1. Parameters used in the model in the standard configuration

Parameter name	Symbol	Value
Acceleration of gravity	g	9.8 m s^{-2}
Contraction coefficient of salt	β	8×10^{-4}
Salinity in the deep Kattegat	S_D	33.5
Coriolis parameter	f	1.2×10^{-4} s^{-1}
Mixing efficiency	m_0	0.6
Mean depth of the Kattegat	h_m	15.5 m
Friction velocity in the Kattegat	u_*	0.013 m s^{-1}
Surface area of the Kattegat	A_0	43 300 km^2
Frontal volume	V_F	10 km^3
Mixing ratio in the Kattegat	γ	0.5
Freshwater supply to the Baltic Proper	Q_{f0}	9280 m^3 s^{-1} (58%)
Freshwater supply to the Bothnian Sea	Q_{f1}	3200 m^3 s^{-1} (20%)
Freshwater supply to the Bothnian Bay	Q_{f2}	3520 m^3 s^{-1} (22%)
Sill depth in the Southern Kvark	h_{SK}	40 m
Sill depth in the Northern Kvark	h_{NK}	20 m
Sill width in the Northern Kvark	W	3300 m
Surface area of the Baltic Sea	A_{B0}	370 000 km^2

$$V_i = \int_{\tau_i} Q\,\mathrm{d}t \tag{13}$$

If V_i is larger than a given volume, V_F, water with salinity S_{IN} is transported into the Baltic. V_F represents the volume of southern Baltic proper water that resides at the sills of Darss and Drogden after an outflow and needs to be transported back into the Baltic before the salty water can enter. So Q_{IN} is computed from

$$Q_{IN} = \frac{1}{T}\sum_i \max\left[V_i - V_F, 0\right] \tag{14}$$

A similar expression for Q_{OUT} is easily derived.

The salinity of the inflowing water is given by similar parameterisations as in Gustafsson (1997b) and Gustafsson & Westman (2002). In the model, the upper layer of the Kattegat and Belt Sea is treated as a single water mass (box), while the lower layer is treated as a dynamically passive source of water and salt (see Fig. 3). In steady state the volume and salt conservation of the box are

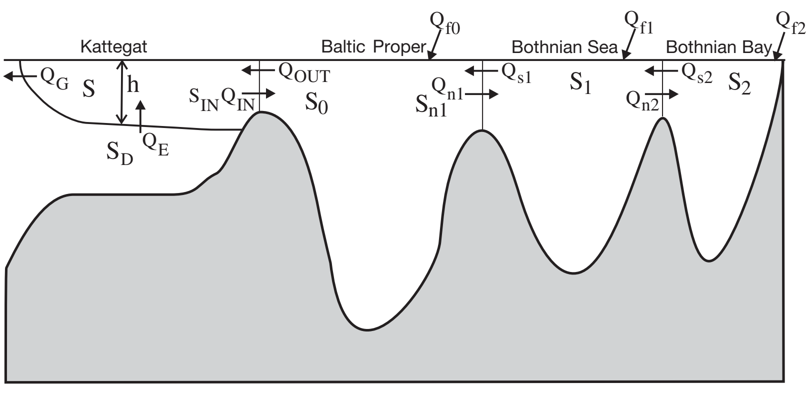

Fig. 3. Diagram of volume flows and state-variables of the salinity model

$$-Q_G + Q_E + Q_F = 0 \tag{15}$$

$$-SQ_G + S_D Q_E + S_0 Q_{OUT} - S_{IN} Q_{IN} = 0 \tag{16}$$

Eq. (15) expresses volume conservation of the Kattegat/Belt Sea box, where Q_G is the geostrophic outflow to the Skagerrak, Q_E is the entrainment flow from the lower layer and Q_F (= $Q_{f0}+Q_{f1}+Q_{f2}$) is the net supply from the Baltic Sea (freshwater input). Eq. (16) expresses the conservation of salt in the upper layer of salinity S, where S_D is the salinity of the lower layer.

The geostrophic flow and the entrainment velocity, w_E, are calculated from

$$Q_G = \frac{g\beta(S_D - S)h^2}{2f} \tag{17}$$

$$w_E = \frac{2m_0 u_*^3}{g\beta(S_D - S)h} \tag{18}$$

where h is the depth to the pycnocline in the Kattegat–Belt Sea sector, f is the Coriolis parameter, g is the constant of gravity, β is the contraction coefficient of seawater due to addition of salt, m_0 is the efficiency of turbulent mixing with respect to work against the buoyancy forces, and u_* is the friction velocity.

The hypsographic function of the Kattegat–Belt Sea shows that the horizontal area $A(h)$ decreases approximately linearly with depth, and at a depth of 15.5 m the area is 50% of the area at the sea surface. However, the hypsographic function of the Kattegat–Belt Sea needs to be adjusted to changes in MSL. The hypsography shows that a linear relationship between area and depth is valid within the range of MSL changes we consider here. This means that the entrainment flow in Kattegat/Belt Sea increases when sea level rises, since the horizontal area of the halocline increases.

Using Eqs. (12), (15) to (18) and the hypsographic function one can deduce analytical expressions for salinity and halocline depth in the Kattegat–Belt Sea. These are

$$h = \frac{C_M(2h_m + z) + FQ_F}{C_M + Q_F} \tag{19}$$

$$S = S_D\left[1 - \frac{F}{h}\right] \tag{20}$$

where S is expressed by the definition of freshwater height (F) (see e.g. Gustafsson & Stigebrandt 1996). In this case, F is given by

$$F = \sqrt{\frac{2fQ_F}{g\beta S_D}} \quad (21)$$

and C_M is a parameter defined by

$$C_M = \frac{m_0 u_*^3 A_0}{g\beta S_D h_m} \quad (22)$$

where A_0 is the surface area of the Kattegat.

The inflowing salinity, S_{IN}, is a complex function of the dynamics in the Danish Straits and Kattegat. Gustafsson & Andersson (2001) devised semi-empirical relationships that demonstrate that much of the variance is due to advection of the north-south horizontal salinity gradients. The advection is driven by currents associated with the inflows and outflows to the Baltic. However, those relationships are rather complex and tuned to present climate. Therefore they are not really applicable in this investigation of future scenarios. Other dynamic models (e.g. Omstedt & Axell 1998, Gustafsson 2000a,b) require more complex forcing data and this makes them less suitable for demonstrating the main features of the system. Knowing that we need a better parameterisation, we will here apply the simplification that

$$S_{IN} = \gamma S + (1-\gamma) S_0 \quad (23)$$

where γ is a constant for the mixing ratio in the Kattegat, based on present day conditions. Thus, we have 2 unknown empirical parameters, γ and V_F, that are most probably functions of both MSL and freshwater supply. Further investigations are needed to determine their parameterisation. However, we discuss below the sensitivity of the results to changes in these parameters. Tuning the model with present day MSL and freshwater supply we find that $\gamma = 0.5$ and $V_F = 10 \times 10^9$ m^3 give results consistent with observations.

The salinity difference between the surface waters in neighbouring basins is due to the dynamics of the strait connecting the basins. The dynamics of the straits connecting sub-seas within the Baltic Sea are simpler than the dynamics of the Danish Straits and therefore salinity differences between sub-basins are easier to predict.

The flow through the Southern Kvark, that is the passage between the Bothnian Sea and the Baltic Proper, is baroclinic and in geostrophic balance. This is similar to the flow from the Kattegat to the Skagerrak (Eq. 17), but since the halocline in the Bothnian Sea is located deeper than the sill, the flow is controlled by sill depth rather than halocline depth. Thus, the flow is computed from

$$Q_{s1} = \frac{g\beta(S_{n1} - S_1)(h_{SK} + z)^2}{2f} \quad (24)$$

where S_{n1} is the salinity flowing in from the northern Baltic Proper, S_1 is Bothnian Sea salinity, and h_{SK} is a representative depth of the Southern Kvark. Using conservation equations similar to Eqs. (15) & (16), the salinity of the Bothnian Sea is given by

$$S_1 = S_{n1} - \sqrt{\frac{2fS_{n1}(Q_{f1} + Q_{f2})}{g\beta(h_{SK} + z)^2}} \quad (25)$$

where Q_{f1} and Q_{f2} are the freshwater supplies to the Bothnian Sea and Bothnian Bay, respectively. The salinity in the northern Baltic Proper, S_{n1}, is substantially lower than in the south, S_0. The salinity gradient is due to lateral and vertical mixing, and to water circulation patterns in the Baltic. Accurate modelling of the salinity gradient requires a model with greater complexity than the present one; therefore, we make a simplifying assumption that results for the Bothnian Sea depend on unchanging lateral and vertical mixing in the Baltic Proper. Following present conditions, we assume that the water of the northern Baltic Proper comprises an equal mix of water of the southern Baltic Proper and Bothnian Sea, that is $S_{n1} = \frac{1}{2}(S_1 + S_0)$.

In quite narrow straits the effect of the earth's rotation can be neglected. This is the case in the strait between the Bothnian Sea and Bothnian Bay, Northern Kvark, and here the exchange will be limited by baroclinic hydraulic control (Stommel & Farmer 1953). This concept implies that there exists a so-called critical section in the strait where

$$\frac{u_1^2}{g\beta(S_1 - S_2)h_1} + \frac{u_2^2}{g\beta(S_1 - S_2)h_2} = 1 \quad (26)$$

where $u_{1(2)}$ is the velocity and $h_{1(2)}$ is the thickness of the upper and lower layers, respectively. The conservation of volume and salt in the Bothnian Bay is given by

$$Q_{s2} = Q_{n2} + Q_{f2} \quad (27a)$$

$$S_1 Q_{n2} = S_2 Q_{s2} \quad (27b)$$

Combining Eq. (26) with the conservation laws yields the following equation (e.g. Stigebrandt 1981):

$$P^3\left(1 + \frac{\eta^3}{(1-\eta)^3}\right) - 2P^2 + P = \frac{\eta^3}{F_e^2} \quad (28)$$

where $P = Q_{s2}/Q_{f2}$ (and $Q_{s2} = u_1 h_1 W$ is the outflow of water from the Bothnian Bay), $\eta = h_1/(h_{NK} + z)$, $h_{NK} + z = h_1 + h_2$, W is the width of the strait, and the estuarine Froude number F_e is defined by

$$F_e^2 = \frac{Q_{f2}^2}{g\beta S_1 (h_{NK} + z)^3 W^2} \quad (29)$$

Eq. (28) has 2 real roots only when $F_e \leq 1$. The baroclinic flow is maximal when the 2 roots become equal. This does not necessarily occur, depending on the rate of mixing in the estuary and the topography of the strait, but when it occurs the estuary is defined as overmixed (Stommel & Farmer 1953). A baroclinic control of this type may be superseded by strong barotropic flows. This is the case, for example, in the southern Öresund. Stigebrandt (2001) showed that the exchange of Bothnian Bay water through the Northern Kvark can be explained if the Bothnian Bay is regarded as an overmixed estuary. Here we use Eq. (28) to compute the flow Q_{s2} and then the salinity S_2 is computed using

$$S_2 = S_1 \frac{Q_{s2} - Q_{f2}}{Q_{s2}} \tag{30}$$

3. RESULTS

The change in topography of the Danish Straits due to MSL change was computed for several cross-sections (see Fig. 4). The cross-section areas, mean depths and widths were computed from the digitised topography for different MSLs and the results are illustrated in Figs. 5 & 6. MSL changes result in changes in the frictional resistance according to Eq. (8); the changes in topographic factor α for each cross-section are shown in Fig. 7. The various cross-sections in the Belt Sea have similar values of α, thus any single α can be representative for the overall flow resistance. In the Öresund, α is clearly smaller for negative MSL changes in the Drogden cross-section, while for increased MSL α is lower at the other sections. The topographic information has limited accuracy; the standard deviation of the depth is 1 to 4 m in each cell and the horizontal resolution is about 1 km. Thus, an absolute error of about 1 m in depth and 1 km in width is anticipated. Errors have their largest effect for the lowest sea level positions since the relative error increases with reduced depth and width of the straits. The model is tuned to give correct results for zero MSL change. Therefore, it is actually the error in change of topography, i.e. α, that propagates into the results, and a definitive error estimate of the results is difficult.

Using the minimal α found for each strait and MSL change, Q_{IN} and Q_{OUT} are computed using present-day external sea level forcing. Computations were made for a freshwater supply between 0 and 60 000 $m^3\ s^{-1}$, and the result is shown in Fig. 8 as the ratio between Q_{IN} and Q_{OUT}. As indicated by Eq. (12), this ratio determines the difference between inflowing and outflowing salinity. In the extreme scenario, with low MSL and high freshwater supply, the flow becomes nearly unidirectional and the Baltic becomes a lake with outlets through the Straits.

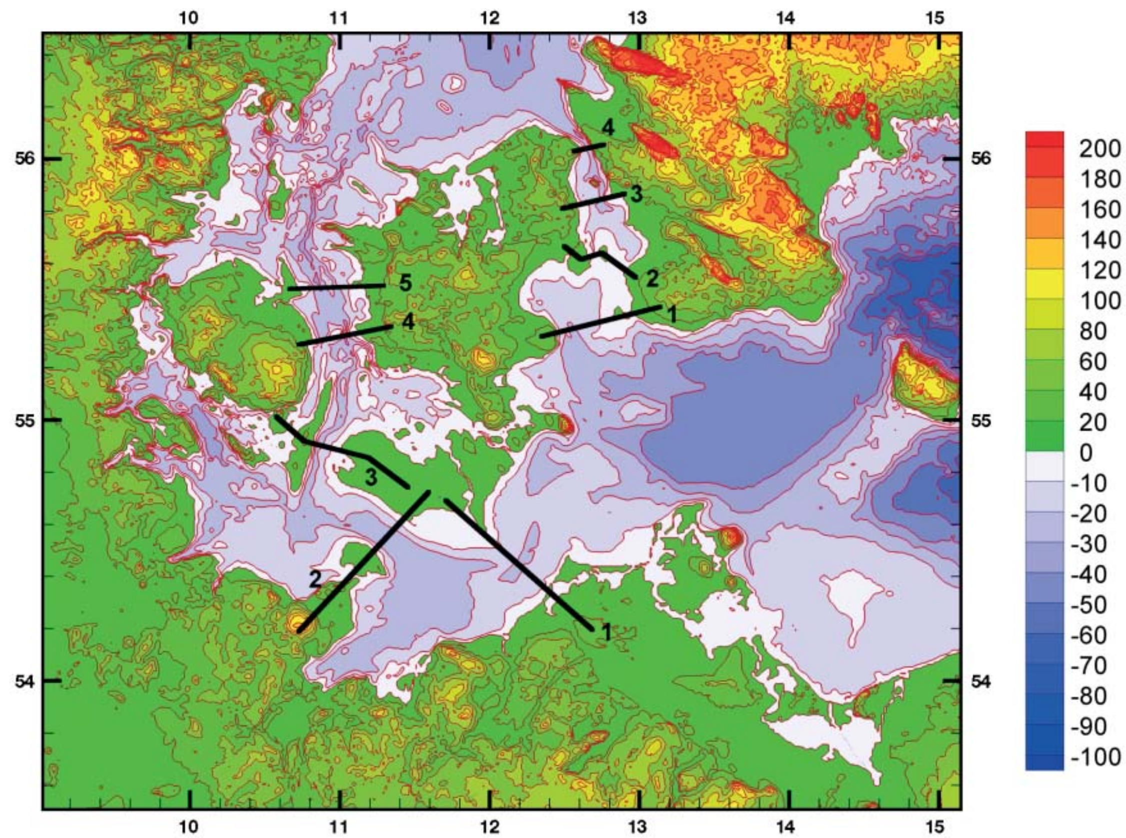

Fig. 4. Approaches to the Baltic Sea and cross-sections for which topographic changes were modeled

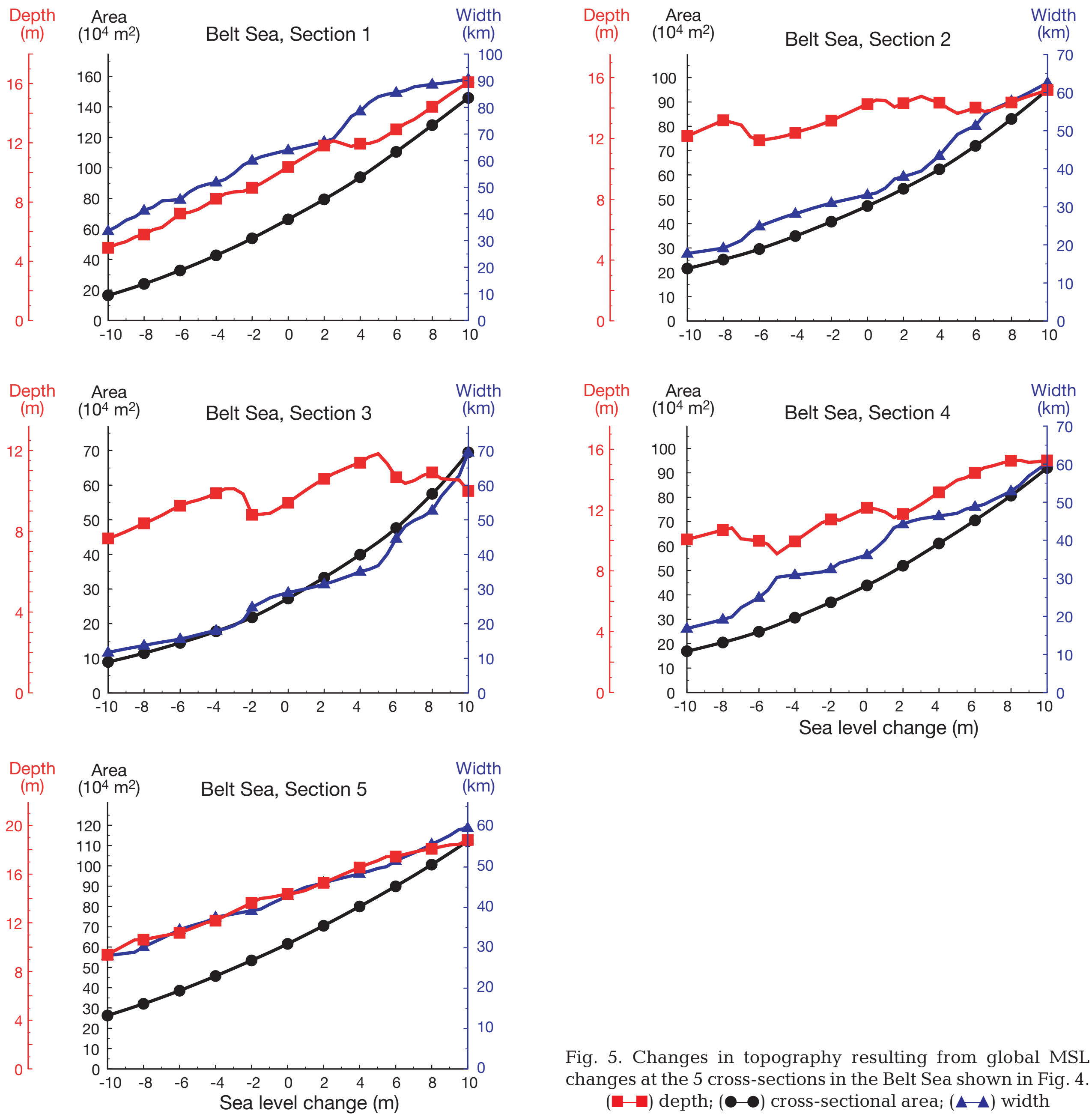

Fig. 5. Changes in topography resulting from global MSL changes at the 5 cross-sections in the Belt Sea shown in Fig. 4. (■—■) depth; (●—●) cross-sectional area; (▲—▲) width

The change in ratio is larger for decreased than for increased MSL, due to the effect of the shallow and wide cross-sections in the Öresund and Belt Sea, which limit the flow for decreased MSL, whereas for increased MSL these cross-sections become less limiting than deeper and narrower cross-sections (see Fig. 7).

The computed southern Baltic salinity, S_0, as a function of MSL change and freshwater supply, is shown in Fig. 9. The salinity change is larger for a sinking than for a rising MSL, as is expected from the flow ratio shown above. With present day freshwater supply (16 000 m^3 s^{-1}) a rise in MSL of 1 m would cause the salinity to increase by 0.8, whereas an MSL drop of 1 m would cause a decrease in salinity by 1.5. Further, an MSL drop of about 5 m would reduce the salinity of the Baltic to <1 and obviously force the conditions in the northern Baltic to a limnic state. The results show that the Baltic environment might not be influenced very much by the rise in MSL hypothesised to occur in the near future due to the greenhouse effect. The changes due to variations in freshwater supply are significant, but transformation of the

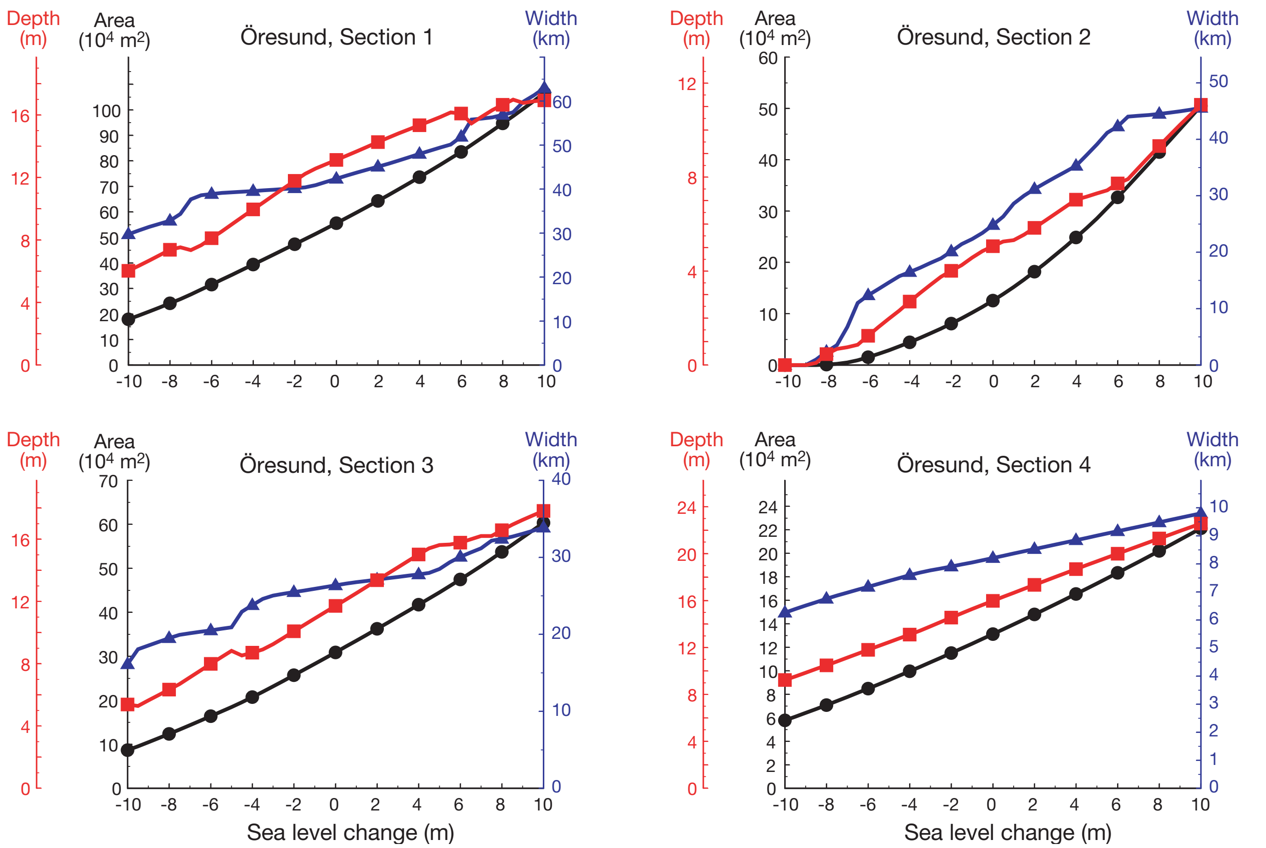

Fig. 6. Changes in topography resulting from global MSL changes at the 4 cross-sections in the Öresund shown in Fig. 4. Same symbols as Fig. 5

Baltic into a lake would require a 4-fold increase in runoff. A more probable scenario could be the decrease in freshwater supply due to the significantly colder climate preceding a new glaciation. Fig. 9 shows that with a 50% reduction of the freshwater supply, which could correspond to the cessation of runoff to the northern Baltic Sea, the surface salinity in southern Baltic would increase to about 15.

The salinities in the Bothnian Sea and Bothnian Bay for different MSL and freshwater runoff are shown in Figs. 10 & 11. Here the freshwater supply is changed in proportion to the distribution of present day river runoff between the main basins of the Baltic (see Table 1). For modest changes in MSL and freshwater supply, the absolute changes in salinity closely resemble the results for the southern Baltic Proper. However,

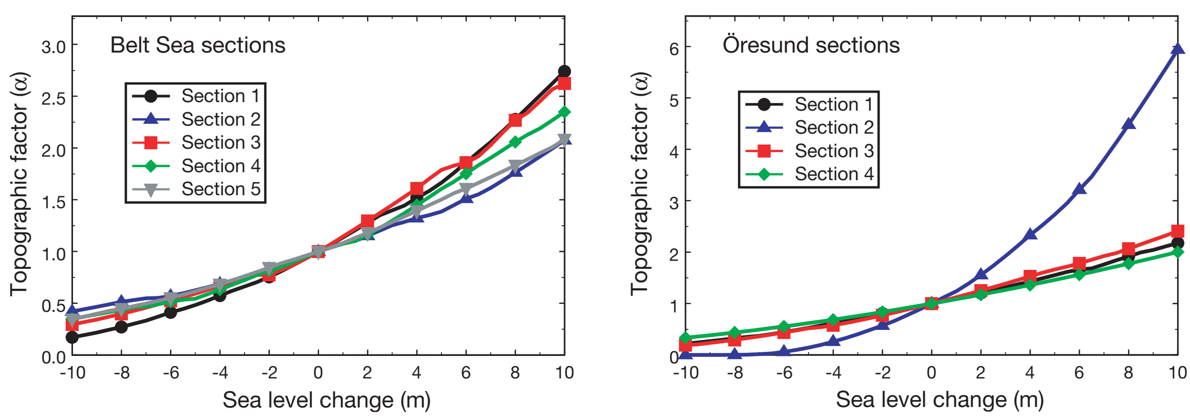

Fig. 7. Computed topographic factor α for the cross-sections in the Belt Sea and Öresund

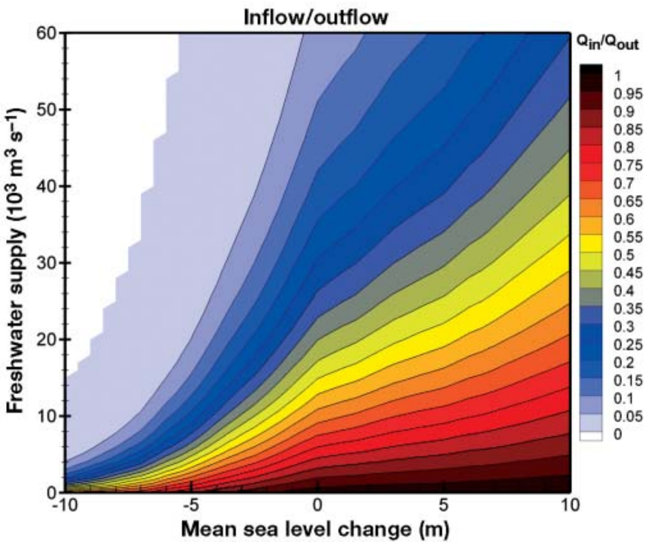

Fig. 8. Ratio between Baltic Sea inflow and outflow through the Danish Straits for different combinations of freshwater supply and MSL change

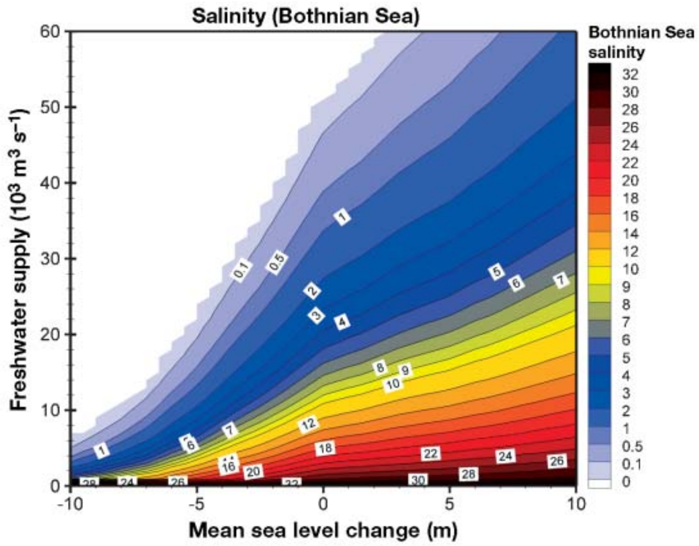

Fig. 10. Salinity in the Bothnian Sea for different combinations of freshwater supply and MSL change

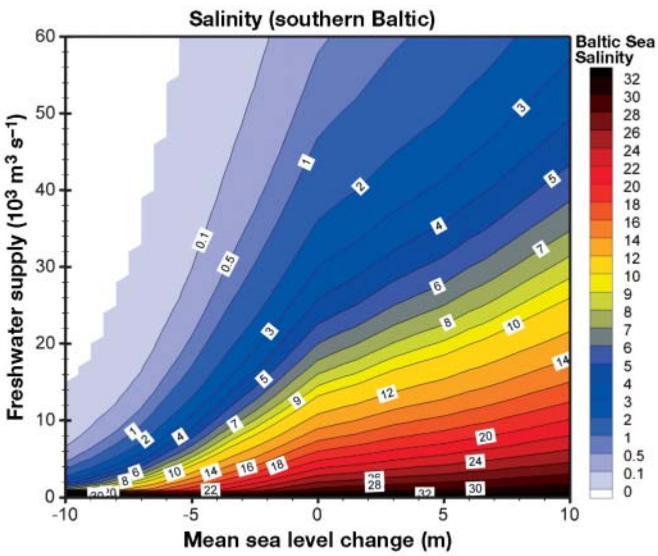

Fig. 9. Salinity in the southern Baltic Proper for different combinations of freshwater supply and MSL change

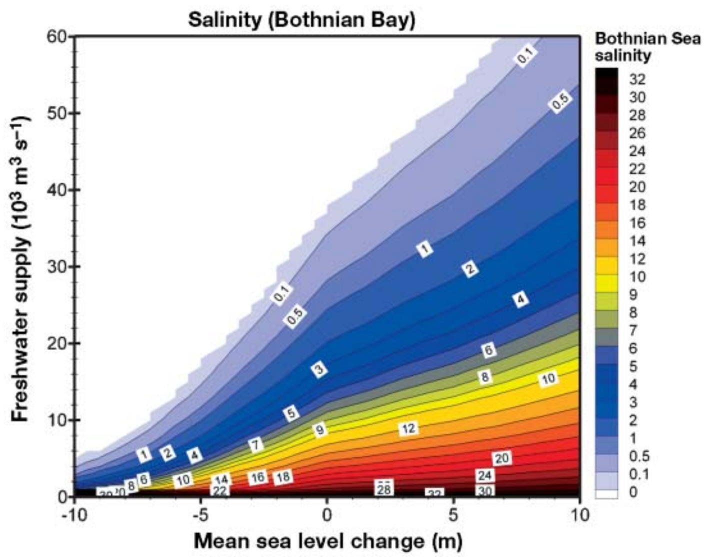

Fig. 11. Salinity in the Bothnian Bay for different combinations of freshwater supply and MSL change

the salinity difference between the Baltic Proper and the Bothnian Sea or Bothnian Bay increases with increasing MSL and decreasing freshwater supply. Due to the low salinities in the Bothnian Bay the relative change in salinity here is large even for modest changes in MSL and runoff.

We also computed the sensitivity of the salinity in the southern Baltic Proper due to changes in wind mixing, deep water salinity and sea level variability in the Kattegat. Fig. 12 shows salinity in the southern Baltic Proper against friction velocity (converted to wind speed for simplicity), deep water salinity and standard deviation of sea level variations in the Kattegat. The friction velocity is computed from the cubic root of the mean wind cubed, i.e. $\overline{w_f} = \sqrt[3]{\overline{w^3}}$, where the overbar represents average over time. Present conditions are characterized by a rather high wind speed (11 m s^{-1}). The amplitude of sea level forcing of the model is varied by multiplying the time-series by a factor. Thus the frequency characteristics of the sea level time-series are retained. Fig. 12 shows that salinity is much more sensitive to decreased rather than to increased wind mixing. The Baltic salinity decreases to 4.5 for a decrease in wind speed of 50%, while it only increases by 0.5 for a 50% increase in wind speed.

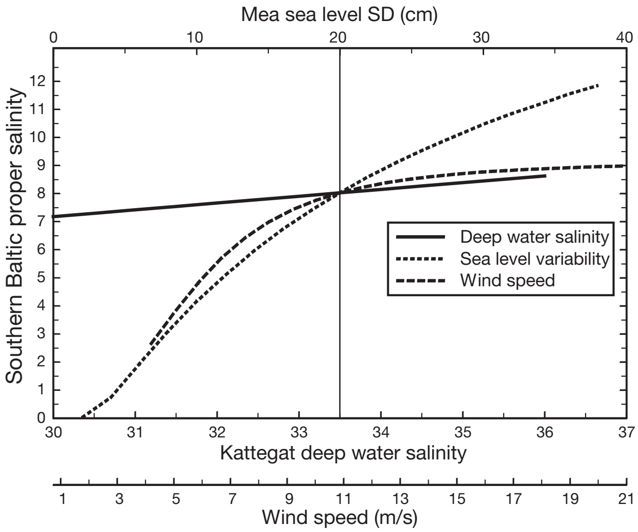

Fig. 12. Salinity in the southern Baltic Proper as function of wind-forced mixing, sea level variability (standard deviation), and deep water salinity in the Kattegat. The vertical reference line indicates present-day conditions (parameters and values in Table 1)

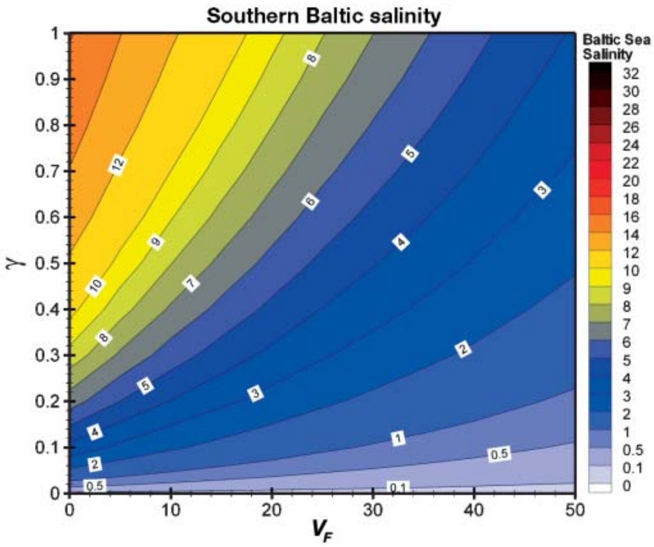

Fig. 13. Sensitivity of the salinity in the southern Baltic Proper to changes in the model parameters. γ: mixing ratio in the Kattegat; V_F: frontal volume

Fig. 12 shows that a change in sea level variability (standard deviation) in the Kattegat by 50% reduces the salinity to 4 (or increases it to 10.5). The sensitivity for smaller changes is a salinity of 1 per 10% change. This is higher than reported by Gustafsson & Westman (2002), probably because of the new method to compute the efficient exchange.

Salinity may change in the deep layer of the Kattegat, S_D, due to changing conditions in the North Sea and in the Atlantic Ocean. The model results shown in Fig. 12 demonstrate that at S_D = 30 to 36, the change in salinity in the southern Baltic Proper is 0.24 per 1 salinity unit change in S_D. Thus, salinity in the Baltic is rather insensitive to changes in S_D, confirming (Stigebrandt 1983).

4. VALIDITY OF THE RESULTS

4.1. Sensitivity to model parameters

There are 2 major parameters in the model: γ, which represents the mixing ratio between Kattegat surface water and water from the southern Baltic Proper in the water that flows into the Baltic (Eq. 23), and V_F, which represents the volume of water that must pass the sills before effective salt exchange occurs (Eq. 14). Fig. 13 shows the Baltic salinity for different combinations of γ and V_F. The model results shown in the previous section are for $\gamma = 0.5$ and $V_F = 10 \times 10^9$ m^3, giving a salinity of 8 in the southern Baltic Proper. Fig. 13 shows that a salinity of 8 also results from other combinations of γ and V_F as well, ranging from $\gamma = 0.27$ and $V_F = 0$ to $\gamma = 1$ and $V_F = 25 \times 10^9$ m^3. It is not surprising to find that there is a range of combinations that reproduce the present-day salinity, because the 2 processes have a similar effect, changing the efficiency of salt exchange. To further elucidate the model's sensitivity to these parameters, the salinity of the southern Baltic Proper was computed for the same range of freshwater supply and MSL change as above, but using the 2 extreme combinations of γ and V_F given above. The results (Fig. 14) show that the model gives similar results for both combinations; however, the salinity is more sensitive to changes in MSL and freshwater supply in the case $\gamma = 1$, $V_F = 25 \times 10^9$ m^3 and less sensitive in the case $\gamma = 0.27$, $V_F = 0$ compared to the standard case. Thus, the model results are not very sensitive to the choice of these parameters as long as they do not vary with freshwater supply and/or MSL change. In the case of changing freshwater supply, we compared the model results with the 3D ocean model simulations by Meier & Kauker (2003a). The responses of the 2 models are quite similar (Table 2). The differences can partly be explained by the difference between salinity in the southern Baltic Proper and spatially averaged Baltic Sea salinity. This comparison indicates that it is reasonable to assume that γ and V_F remain fairly constant when freshwater supply changes.

The bottom drag coefficient, C_d in Eq. (5), is considered to be constant. In reality, theory and observations indicate that the drag coefficient varies with the ratio between water depth and bottom roughness. For example Li et al. (2004) present recent observational evidence that the bottom drag coefficient increases with decreasing depth. In principle, one could implement a variation of the drag coefficient due to changes in MSL. The problem is, however, that we can expect bottom roughness to change as well when parts of the bed emerge or when adjacent land becomes flooded. It might be possible to quantify changes in bottom roughness by compiling information on the spatial distribution of bottom types, but it seems hardly worthwhile for this simple model approach. Bottom roughness at any given location may also vary in time due to changes in vegetation and mussel beds (e.g. Green et al. 1998) or abiotic changes such as sedimentation/ erosion, and these effects are virtually impossible to take into account. In addition, there might be additional drag due to conversion from barotropic to baroclinic motions both in the Belt Sea (Jakobsen & Trebuchet 2000) and in the Öresund (Stigebrandt 1999).

It is possible to estimate within an order of magnitude how a depth dependent drag coefficient will affect the results. From Eq. (5) we see that the transmission coefficient is proportional to $C_d^{-0.5}$ and to $H^{1.5}$. Assuming that C_d is approximately inversely proportional to water depth, as indicated by Li et al. (2004), the transmission coefficient would be proportional to H^2 instead. Thus, a more elaborate modelling of the bottom drag would likely increase the sensitivity of the Baltic salinity to changes in MSL. Changes in the structure of the atmospheric boundary layer, e.g. changed stability, influence the momentum transfer from the atmosphere to the ocean and perhaps also infuence how efficiently the turbulent kinetic energy drives mixing. The magnitude of these influences can be estimated using Eq. (22). Changes in boundary layer structure would influence in the same direction as changes in wind speed, but the sensitivity would be less because of the cubic influence from the wind while change in momentum transfer would influence through a power of $3/2$ and mixing efficiency, m_0, has a linear influence.

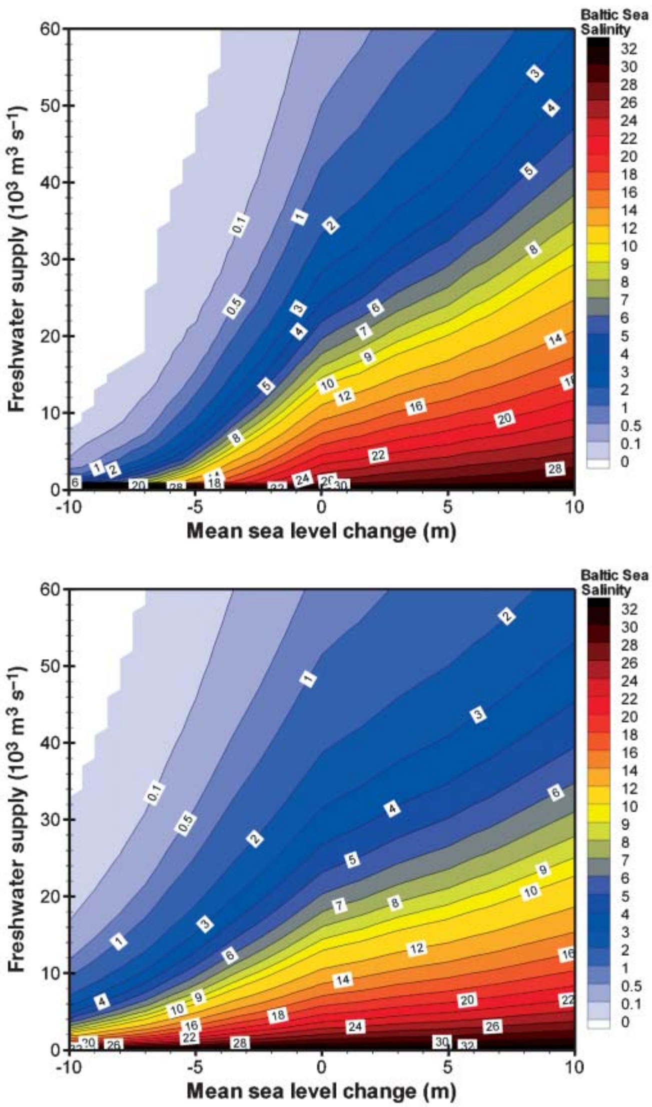

Fig. 14. Response of the salinity in the Baltic Proper to changes in freshwater supply and MSL for 2 extreme settings. Top: $\gamma = 1$, $V_F = 25 \times 10^9$ m^3; bottom: $\gamma = 0.27$, $V_F = 0$

Table 2. Salinities in the southern Baltic Proper for different freshwater supplies in comparison with corresponding spatially averaged Baltic Sea salinities computed by Meier & Kauker (2003a)

Freshwater supply ($m^3 s^{-1}$)	Present model	Meier & Kauker (2003a)
5000	19.62	22.99
12736	10.37	10.99
16115	7.96	7.98
21655	5.34	4.56
32230	2.59	1.36
64460	0.34	0.03

4.2. Other climatic effects

Variability in sea level in the Kattegat will probably be affected by large deviations in overall MSL. Gustafsson & Andersson (2001) showed that the sea level variations in the frequency band that predominantly drives Baltic Sea salt exchange is closely related to the regional east–west winds (i.e. north–south air pressure difference). This indicates that salt exchange is mostly due to the wind regime. The simplest wind regime would be a constant wind blowing over a closed basin. In this case the pressure force due to sea level slope in the basin is adjusted to balance the wind stress, i.e.

$$g\eta_x = \frac{\tau^{\text{wind}}}{\rho H} \quad (31)$$

where η_x is the surface slope, g is acceleration of gravity, H is the depth of the basin, ρ is the density and τ^{wind} is the wind stress. This indicates that for a given wind forcing and basin geometry the sea surface slope, and hence the sea level variations, will increase when depth decreases, i.e. if MSL drops. This could counteract the reduction in Baltic salinity that results from the shallower sills. However, an assessment would require an accurate evaluation of the causes for sea level variability in the Kattegat. Changes in the flow through the Straits are proportional to α, which in turn is proportional to $H^{1.5}$ (the depth of the strait raised to the power 3/2) and to the linear width of the strait (Eq. 7), but only proportional to the square root of the sea level difference between the Kattegat and the Baltic (Eq. 9). Thus, if sea level variability changes according to Eq. (31), it will only influence exchange by a factor of $H^{0.5}$, which is less than the change in resistance in the Straits.

Present tides are quite weak in the Kattegat (amplitudes of 5 to 20 cm). This is partly due to the dissipation of the tidal energy in the North Sea, and partly due to the location of an amphidromic point just outside the Skagerrak. Am amphidromic point is where the tides diminish due to interaction of 2 or more tidal waves. In this case, one tidal wave propagates along the shores of the North Sea and other through the Norwegian trench and these meet just outside the Skagerrak. If MSL changes, this pattern may change as well, and tides in the Kattegat could become significant; tidal amplitudes of several meters are conceivable if the bathymetry becomes favourable for tidal resonance. This would lead to complete vertical tidal mixing in the Kattegat. Exchange with the Skagerrak would probably remain geostrophic, but Eq. (17) would apply to the entire depth of the Kattegat, instead of being restricted to the surface layer. The entrainment flux would be replaced by a compensating horizontal flow of Skagerrak water. Under these conditions, the salinity in the northern Kattegat is given by

$$S = S_D\left(1 - \frac{F}{2h_m + z}\right) \tag{32}$$

where F is given by Eq. (21). Assuming present day conditions, i.e. Q_F = 16 000 m^3 s^{-1}, γ = 0.5 and Q_{IN}/Q_{OUT} = 0.5, the salinity in the southern Baltic Proper can be computed from Eqs. (12), (21), (23) & (32) and complete vertical mixing in the Kattegat would result in a salinity increase from 8 to 9.8. This calculation demonstrates the sensitivity of Baltic Sea salinity to such a change, but it is hardly realistic since a dramatic change in tidal amplitudes cannot occur without a significant MSL change. A prognosis of tides and tidal mixing for different MSL variations is beyond the scope of this study, but can readily be performed using a 2-dimensional model.

4.3. Changes in the dynamics of the Kattegat

Intuitively, one could expect that large deviations in MSL and freshwater supply would cause drastic changes in the circulation of water in the Kattegat. This is, however, not the case according to the model results, which predict relatively modest changes in pycnocline depth and smooth variations in salinity (Figs. 15 & 16). These variations are not drastic enough to change the geostrophic balance of the outflow.

If pycnocline depth were reduced significantly, deep water could intrude directly into the Baltic over the sills. This may occur occasionally today, constituting a minor source of salt for the Baltic (Stigebrandt 1983,

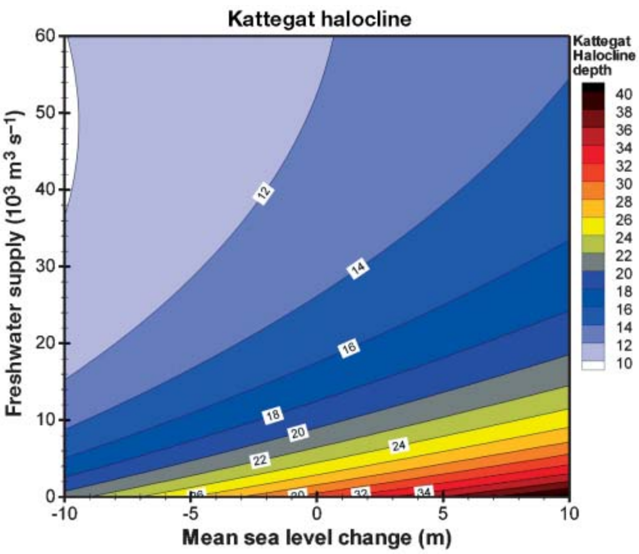

Fig. 15. Halocline depth in the Kattegat for different combinations of freshwater supply and MSL change

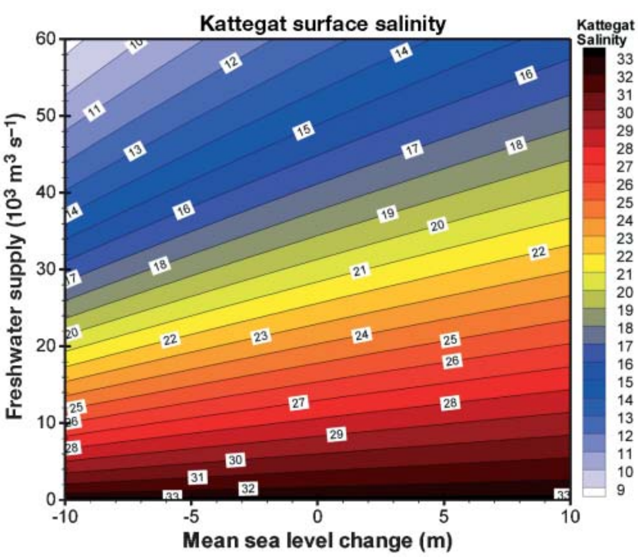

Fig. 16. Surface salinity in the Kattegat for different combinations of freshwater supply and MSL change

Gustafsson 2000b). The model results show that the pycnocline depth does not decrease much with increasing freshwater supply, so at the present MSL the inflow of Kattegat deep water over the sills will change little. For decreasing MSL, the decrease in pycnocline depth is less than the decrease in depth of the sills, so that deep water inflows would become less probable. However, for increasing MSL the pycnocline depth increases less than the depth of the sills. Thus, more high-saline deep water would intrude into the Baltic than predicted by the model. Therefore we can expect that the model result underestimates the salinity in the Baltic Sea for increased MSL.

5. DISCUSSION

This study quantifies the sensitivity of Baltic Sea salinity to variations in freshwater supply and in sea level. The aim was not to directly predict the future state of the Baltic Sea, but to present tools for quantifying the response to climatic changes. However, the result may be used to estimate Baltic Sea salinity for a given future freshwater supply and global MSL.

A quantification of the accuracy of the results is difficult, but the results gain credibility by making order of magnitude estimates of the contributions from hitherto neglected processes. Previous estimates of the sensitivity of Baltic salinity to climate change, primarily variations in freshwater supply, obtained with similar models by Gustafsson (1997b), Gustafsson & Westman (2002) and Stigebrandt & Gustafsson (2003) have been confirmed by more complex time-dependent high resolution models (Gustafsson 2000b, Omstedt et al. 2000, Meier & Kauker 2003a). Thus, more complex models only yield a small increase in accuracy.

Previous studies of the past climate have shown that variations of the freshwater supply have been the principal driver of salinity variations (Gustafsson & Westman 2002). The colder climate preceding a glaciation would reduce river runoff to the northern Baltic Sea, due to formation of glaciers. On the other hand, Gustafsson & Westman (2002) indicated that warm periods were associated with dry conditions; thus one might expect that cold periods should be associated with wetter conditions. Probably the freshwater supply would increase initially as the climate becomes colder, until the formation of glaciers becomes dominant. However, in a shorter perspective, river runoff may change in response to greenhouse gas emissions by –2% to +15% in 100 yr (Graham 2004). Thus, the simulations in that study show a likely increase in freshwater supply coherent with a warmer climate.

If global MSL declines by about 5 m, the salinity in the southern Baltic Proper will decrease to <1, and large parts of the Baltic will become limnic. The same change can be brought about by an increase in freshwater supply to 45 000 $m^3\ s^{-1}$ (a factor of 3), which is unlikely in any future climate.

The Baltic will not lose its estuarine nature and become an oceanic bay as long as the freshwater supply is positive. Negative freshwater runoff occurs for example in the Mediterranean, which is partly surrounded by deserts that supply dry and warm air favouring evaporation. Such a climate is improbable in northern Europe.There are indications that the freshwater supply was about 50% of the present supply during the warm period some 5000 yr BP, with a concommittant increase in southern Baltic salinity to about 15. If the freshwater supply decrease is evenly distributed, the model shows even more drastic changes in the Bothnian Sea and Bothnian Bay, where the salinities would attain 13 and 10, respectively.

The results show that the sea level variability in the Kattegat has a significant influence on the Baltic Sea salinity (Fig. 12) The part of the sea level spectrum that is capable of forcing salt exchange is driven by regional winds (Gustafsson & Andersson 2001). It is reasonable to assume that the mechanism for this is wind setup (cf. Eq. 31), where the sea level elevation is proportional to the wind stress on the sea surface. Since the wind stress is roughly proportional to wind speed squared, the standard deviation of the Kattegat sea level should be proportional to the variance of the wind speed. The standard deviation of the sea level should therefore be roughly proportional to the variance in the wind. The sensitivity analysis shows that reduced wind mixing in the Kattegat could reduce the salinity in the Baltic (Fig. 12). Thus, this model indicates that the combined effect of reduced mixing in the Kattegat and reduced exchange through the Danish Straits can result in significant reductions in Baltic salinity. However, the model sensitivity to changed wind-driven mixing does not conform with results from more complex numerical models, because some wind-forced processes are not described by this simple model. Gustafsson (2000b) found that the baroclinic exchange across the Darss Sill would increase considerably if mixing in the Kattegat and Belt Sea decrease, resulting in an increase in salinity in the Baltic Proper salinity. Meier & Kauker (2003b) showed that a substantial part of the low frequency (>4 yr) variations in Baltic Sea salinity is due to variations in 4 yr mean winds which affect both circulation within the Baltic Sea and vertical mixing. The response tended to be reduced salinities at increased wind velocities. Meier & Kauker (2003b) did not find a significant response to changes in the 4 yr sea level variations in the Kattegat; long-term variations in wind velocities therefore appear to affect the internal circulation in the Baltic Sea to the extent that salinity is changed.

The strength of the vertical circulation and length of stagnation periods are extremely important for quantification of the ecological response to different climate developments, especially in the Baltic proper. The main reason for leaving this out of the present investigation is that vertical circulation is influenced by day-to-day variations in weather, while the overall salinity is related to long-term climate trends. Recent research has shown that the major oceanic inflows causing renewal of the deep water in the Baltic are simply forced by the regional air pressure field (Gustafsson & Andersson 2001). This knowledge can be combined with computationally efficient time-dependent circulation models (e.g. Gustafsson 2003), to determine the response of vertical circulation to changing climatic parameters.

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