

Modeling the age of Baltic Seawater masses: Quantification and steady state sensitivity experiments

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[1] Ages of Baltic Seawater masses for the period 1903–1998 were quantified using a three-dimensional (3-D) coupled ice-ocean model. Therefore an additional Eulerian tracer for the age of seawater was embedded. The age is the time elapsed since a water particle left the sea surface. Median ages of the bottom water between 1 year in the Bornholm Basin and 7 years in the northwestern Gotland Basin were found. During 1903–1998 the oldest bottom water of about 11 years appeared at Landsort Deep. In the halocline of the deeper basins a secondary age maximum was calculated. In the eastern Gotland Basin 3 stagnation periods (in the 1920/1930s, 1950/1960s, and 1980/1990s) with ages exceeding 8 years were found. Further, the sensitivities of modeled salinity and age on freshwater supply, wind speed, and amplitude of the sea level in Kattegat were investigated. In steady state the average salinity of the Baltic is most sensitive to perturbations of freshwater inflow. Increased freshwater inflow and wind speed result both in decreased salinity whereas increased amplitude of the Kattegat sea level results in increased salinity. The average age is most sensitive to perturbations of the wind speed. Especially, decreased wind speed causes significantly increased age of the deep water. On the other hand, the impact of changing freshwater or sea level in Kattegat on the average age is comparatively small, suggesting invariance of stability and ventilation in steady state approximately. A simple conceptual model for the Baltic deep water ventilation was applied to explain the 3-D model results.

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1. Introduction

[2] The Baltic Sea is a strongly stratified semi-enclosed sea with a total volume of $21,500 \text{ km}^3$ (without Kattegat). Horizontal and vertical salinity gradients are the result of the high freshwater supply from rivers and net precipitation and of the reduced water exchange with the world ocean (Figure 1).

[3] During the twentieth century the total mean net freshwater inflow amounts to about $15,000\text{--}16,000 \text{ m}^3 \text{ s}^{-1}$ [e.g., Meier and Kauker, 2003a]. To understand the time evolution of Baltic Sea salinity, long-term investigations are necessary because the typical internal “response timescale” is estimated to be of the order of 30 years [e.g., Winsor *et al.*, 2001; Meier and Kauker, 2003a; Döös *et al.*, 2004]. Assuming that the recirculation of Baltic Sea water is the dominant process for the variability of basin-wide averaged salinity [e.g., Rodhe and Winsor, 2002; Stigebrandt, 2003], the “response time” is approximately determined by the turnover time of the freshwater content in the Baltic Sea. Winsor *et al.* [2003] and Meier and Kauker [2003a] calculated the freshwater content from observations

($1.67 \times 10^4 \text{ km}^3$) and model results ($1.77 \times 10^4 \text{ km}^3$), respectively. Thus the corresponding turnover timescales are 33 and 35 years, respectively. Another approach is to estimate the residence time of freshwater in the Baltic. Döös *et al.* [2004] calculated residence times of various Baltic water masses using Lagrangian particles released either at the sills in the Baltic entrance area or at the mouth of the river Neva. They found, for the entire Baltic, residence times of 26–29 years. The time evolution of the residence of the Neva water in the Baltic is very similar to the water for the entire Baltic [Döös *et al.*, 2004, Figure 3].

[4] To understand the long-term variability of salinity in the Baltic Sea, two fundamentally different cases need to be considered. In the first case, the timescale of the forcing anomaly is smaller than the response timescale of about 30 years. Precipitation, river runoff, and zonal wind over the Baltic Sea show pronounced decadal oscillations with timescales (i.e., half periods) between 4 and 16 years, approximately, causing phases of high and low average salinity [Meier and Kauker, 2003a]. During low-saline phases the deep water in the Baltic proper is poorly ventilated. However, even during high-saline phase, reduced deep water ventilation might occur. These phases are called stagnation periods. Usually, strong inflows of high-saline (and very often oxygen rich) water from the Kattegat occur randomly

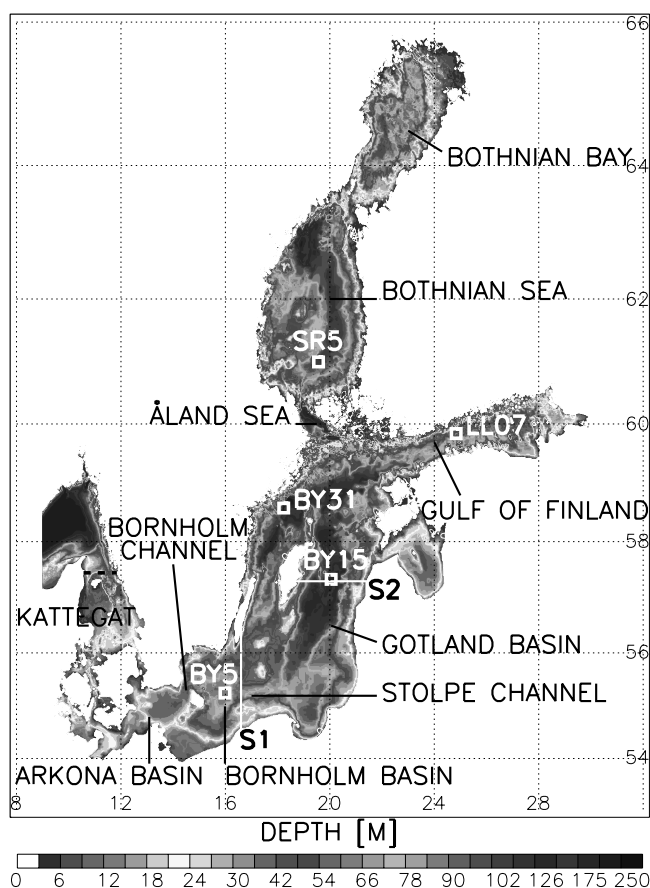


Figure 1. Bottom topography of the Baltic Sea. The domain of the Rossby Centre Ocean model (RCO) is limited with open boundaries in the northern Kattegat (dashed line). Selected stations (BY5, BY15, BY31, LL07, SR5) and cross sections (S1 and S2, white lines) are depicted additionally. See color version of this figure at back of this issue.

at intervals of one to several years and eventually ventilate the deep water [e.g., Matthäus and Franck, 1992; Fischer and Matthäus, 1996]. During the last century, two long-lasting periods of exceptionally low average salinity were identified analyzing model simulations for 1903–1998 [Meier and Kauker, 2003a]. This finding is supported by available observations [e.g., Fonselius and Valderrama, 2003]. Meier and Kauker [2003a] found that about half of the decadal variability of the average salinity of the Baltic is related to the accumulated freshwater inflow. Another significant part of the decadal variability of salinity is caused by the low-frequency variability of the zonal wind. The wind stress anomaly is balanced by a sea-level slope anomaly between Kattegat and the central Baltic. Consequently, an anomalous barotropic pressure gradient hampers saltwater inflows through the Danish straits [Meier and Kauker, 2003a].

[5] In the second case, the timescale of the forcing anomaly is larger than the response timescale of 30 years. In this case the system is shifting into a new quasi steady state. For instance, this might be the consequence of anthropogenically induced climate change in about 100 years. In some regional scenarios a significant increase of precipitation in the Baltic catchment area was calculated

[Räisänen *et al.*, 2004]. Consequently, the salinity of the Baltic Sea was projected to decrease considerably; that is, the change of the mean salinity is larger than the range of natural variability [Meier and Kauker, 2003b]. On the basis of model simulations, the sensitivity of the average steady state salinity on the external forcing has been estimated in several studies [e.g., Stigebrandt, 1983; Gustafsson, 1997, 2000b; Meier and Kauker, 2003b; Stigebrandt and Gustafsson, 2003]. It was found that the sensitivity of the steady state salinity on the freshwater supply is nonlinear. In different model approaches the results agree rather well. The sensitivity of the three-dimensional (3-D) Rossby Centre Ocean model (RCO) [Meier and Kauker, 2003b] is only slightly higher than the sensitivity of the process-oriented model by Stigebrandt [1983]. Even with 100% increased freshwater supply the Baltic cannot be classified as a freshwater sea. Further, the sensitivity in different subbasins was studied when the changing freshwater supply is nonuniform [Stigebrandt and Gustafsson, 2003]. In contrast to these investigations, Rodhe and Winsor [2002] [see also Rodhe and Winsor, 2003] found that the sensitivity on freshwater inflow is much larger. An explanation could be that in the empirical model of Rodhe and Winsor [2002] the freshwater content anomaly depends only on the freshwater supply and not on the wind forcing as found by Meier and Kauker [2003a] [see Meier and Kauker, 2003b, section 5].

[6] For the marine ecosystem in the Baltic, it is crucial to understand the processes controlling ventilation, age, and residence time of the deep water. Phosphorus and nitrogen sinks are sensitive to the redox state which is determined by the concentration of dissolved oxygen [Wulff and Stigebrandt, 1989]. The latter is determined by a balance between the vertical flux of organic matter and the horizontal flux of oxygen-enriched new deep water. Increased supply of organic matter and the lack of saltwater inflows may cause oxygen depletion together with an increase of hydrogen sulphide and accumulation of organic matter in the sediment. Below a critical concentration of 2 mg O₂ L⁻¹, higher forms of life are excluded. In addition, if the near-bottom water is getting anoxic, phosphorus will be released from the sediments [Caraco *et al.*, 1989]. As a consequence, massive cyanobacteria blooms may occur. Therefore, important questions are: How well is the Baltic deep water ventilated during present climate and how sensitive is the age to changes of climatological factors, like freshwater inflow, wind speed, and sea level fluctuations in Kattegat? In the following, the concept of age in marine modeling is used to quantify the ventilation of the Baltic Sea deep water [e.g., Deleersnijder *et al.*, 2001; Gustafsson, 2000a]. Thereby, a passive Eulerian tracer based upon an advective-diffusive equation was embedded into a 3-D ocean circulation model. The sensitivities of modeled salinity and seawater age on the external forcing are investigated.

2. Method

2.1. Model Description

[7] The simulations were performed with the Rossby Centre Ocean model (RCO) [e.g., Meier *et al.*, 1999; Meier, 2001; Meier and Faxén, 2002; Meier *et al.*, 2003]. RCO is a

Bryan-Cox-Semtner primitive equation model with a free surface and open boundary conditions. It is coupled to a Hibler-type sea ice model. An improved turbulence closure scheme is embedded. The model domain covers the Baltic Sea including Kattegat (Figure 1). In the present study, RCO was used with a horizontal resolution of 6 nautical miles and with 41 vertical levels with layer thicknesses between 3 and 12 m. RCO was started from rest with initial temperature and salinity observations from November 1902. In case of inflow a constant deep water salinity of 33.2‰ was nudged at the open boundary in all simulations. In the standard experiment, daily sea level data were prescribed at the open boundary in Kattegat and observed monthly mean river runoff was used. The atmospheric forcing was calculated from daily reconstructed sea level pressure and monthly surface air temperature, dew-point temperature, precipitation, and cloudiness fields [Kauker and Meier, 2003]. For further details of the model setup the reader is referred to Meier and Kauker [2003a].

2.2. Age of Seawater

[8] Different timescales are used in geophysical applications to characterize exchange processes, for example, turnover time, average age, and average transit time (residence time) [Bolin and Rodhe, 1973]. The turnover time is the ratio of the total mass in a reservoir to the total mass flux. In general, the age of a particle of a seawater constituent is defined to be the time elapsed since the particle under consideration left the region, in which its age is prescribed to be zero [Deleersnijder et al., 2001]. The average age is the first moment of the concentration distribution function with respect to age. Finally, the residence time is defined to be the first moment of the concentration distribution function of particles leaving the reservoir. In the case of steady state the residence time is identical with the turnover time [Bolin and Rodhe, 1973]. However, the average age is usually different from the residence time. While the age is estimated as the solution of a direct partial differential problem (see below), the residence time obeys an inverse problem [e.g., Holzer and Hall, 2000].

[9] To track specific water masses a passive tracer might be utilized [Deleersnijder et al., 2001]. At time t and location $\vec{r}$, the concentration of the latter, $C(t, \vec{r})$, would obey the following equation:

$$\frac{\partial C}{\partial t} + \nabla \cdot (\vec{v}C - K \cdot \nabla C) = 0, \quad (1)$$

where $\vec{v}$ is the water velocity and K denotes the diffusivity tensor. The age concentration, $\alpha(t, \vec{r})$, of the water mass under study is the solution of

$$\frac{\partial \alpha}{\partial t} + \nabla \cdot (\vec{v}\alpha - K \cdot \nabla \alpha) = C. \quad (2)$$

Finally, the age is then given as the ratio $a(t, \vec{r}) = \alpha(t, \vec{r})/C(t, \vec{r})$. For further details of the concept of age in marine modeling the reader is referred to Deleersnijder et al. [2001]. Applications of (1) and (2) were presented, for example, by Hirst [1999] and Deleersnijder et al. [2002].

[10] As the major constituent of seawater is pure water, i.e., the concentration of pure water is close to unity, a good approximation for the age of seawater is calculated from the equation of the age of pure water [Deleersnijder et al., 2001],

$$\frac{\partial a}{\partial t} + \nabla \cdot (\vec{v}a - K \cdot \nabla a) = 1. \quad (3)$$

Tests were performed by the present author which showed that the biases applying (3) instead of (1) and (2) are relatively small. The advantage is that only one additional passive tracer has to be considered.

[11] In this study, equation (3) was utilized. The initial age was set to zero. At the sea surface the age was relaxed to zero with a timescale of 1 day. At the open boundary in Kattegat the same radiation condition as used within RCO for temperature and salinity was utilized [Meier et al., 1999]. In case of inflow the age at the open boundary was relaxed to zero with a timescale of 1 day. Test studies revealed that the details of the chosen surface and lateral boundary conditions are not important for the results of this study.

2.3. Experimental Strategy

[12] In this study, the standard experiment is a hindcast simulation for 1903–1998 using reconstructed atmospheric surface fields by Kauker and Meier [2003], as described in section 2.1. It has been shown earlier that the hindcast accurately reproduces temperature, salinity, sea level, sea ice, volume, and salt fluxes on timescales of days to decades, mainly because the reconstructed atmospheric surface data and the other forcing fields have a good quality [Meier and Kauker, 2003a]. The sensitivity experiments were built on the standard experiment but they were performed with increased or reduced freshwater inflow (runoff and precipitation), wind speed, or sea level elevation at the open boundary in Kattegat. Changes of ± 15 and $\pm 30\%$ were applied. Thus a total number of thirteen 96-year-long experiments (including the standard run) was performed. In all sensitivity experiments the average salinity of the Baltic Sea is approximately in a new steady state after 50 years of integration. Hence, median salinities and ages of the last 48 years of the simulation period were used for the analysis. The period 1951–1998 comprises periods of both high and low salinities.

3. Results

3.1. Natural Variability of Salinity and Age 1903–1998

[13] Kattegat saltwater which crosses the sills of the entrance area flows through Bornholm Channel into Bornholm Basin with a maximum depth of more than 90 m (Figure 1). Mainly owing to smaller inflows, the interannual and decadal variability of the deep water below the halocline is high (Figure 2). At the bottom of Bornholm Deep (BY5) the median age amounts to 1.2 years (1.3 years in 81 m depth) (Table 1). Interestingly, a thin layer within the upper halocline (but below the seasonal thermocline) shows a relatively long median age of 0.5 years whereas the surface layer and the layer in about 60 m depth are well ventilated (Figure 2).

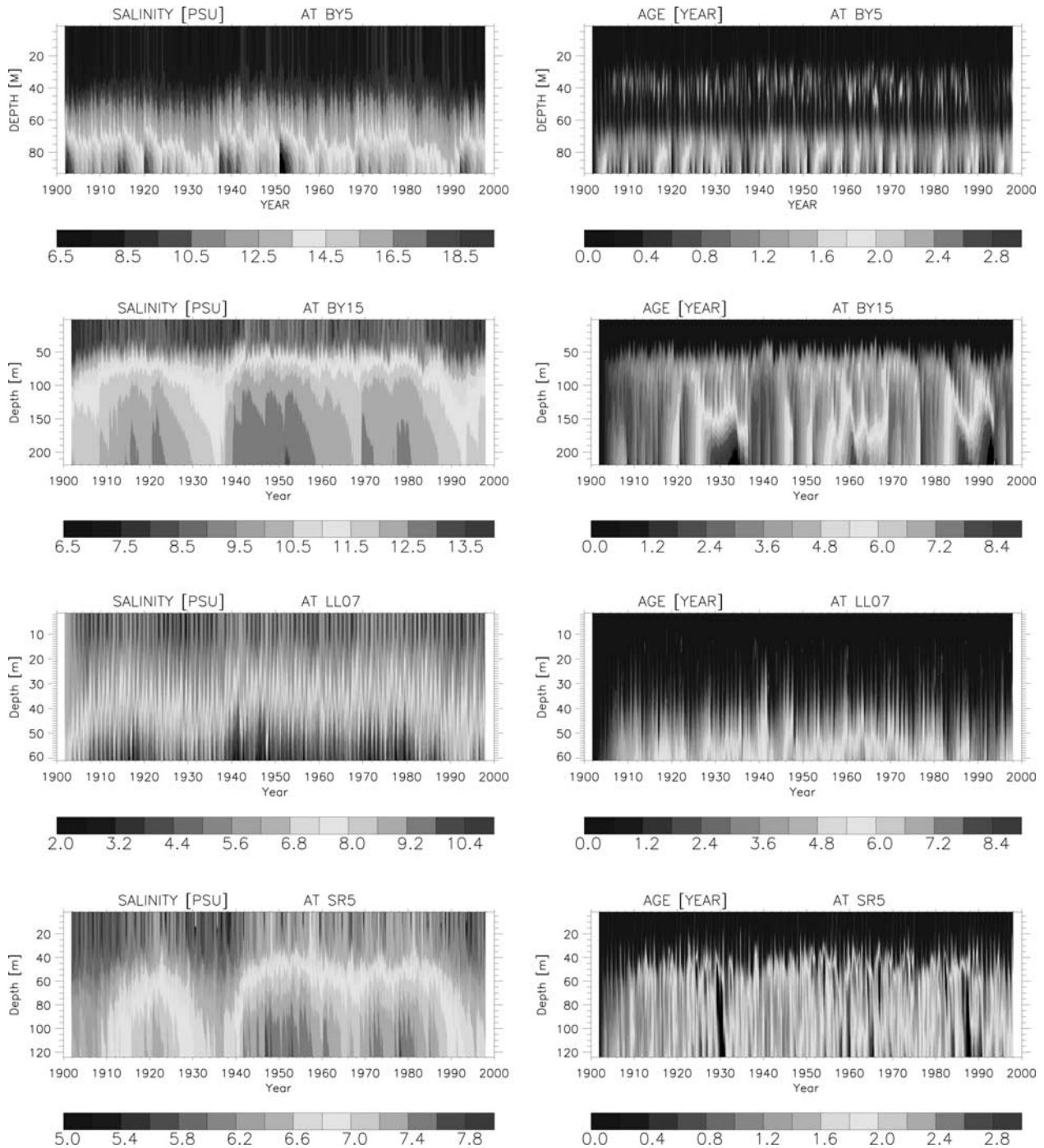


Figure 2. (left) Salinity and (right) age as a function of depth and time at selected stations (see Figure 1). See color version of this figure at back of this issue.

[14] When the salinity of the inflowing water is higher than the bottom salinity at BY5, the older water may be lifted above the sill, which has a depth of 60 m, separating Bornholm Basin and Stolpe Channel (Figure 1). Stolpe Channel (or Šlupsk Furrow) is the only connection between Bornholm Basin and the eastern Gotland Basin. Such water will flow farther into Gotland Deep (BY15). The eastern Gotland Basin is with a maximum depth of 250 m much deeper than Bornholm Basin. Temporal variations of the salinity in the deep water are much larger than the variability in the upper

layer, such as in Bornholm Basin (Figure 2). In the deep water of the eastern Gotland Basin the maximum age amounts to 9.9 years. Thus only rare inflows with relatively high salinities have the potential to renew the bottom water. For the whole simulation period the median age at the bottom of BY15 amounts to about 4.5 years. A local minimum of the age in about 100 to 130 m depth is clearly visible. This minimum corresponds approximately to a local maximum of the volume transport into the Gotland Basin found in observations [Elken, 1996] and model results (Figure 3).

Table 1. Median and Maximum Ages at Selected Stations at the Secondary Maximum Within the Halocline and at the Bottom for 1903–1998^a

	BY5	BY15	BY31	LL07	SR5
<i>Halocline</i>					
Depth	41	81	100	-	53–61
Median age	0.5	3.8	7.4	-	1.6
Maximum age	5.8	5.8	9.7	-	4.2
<i>Bottom</i>					
Depth	93	219	231	61	124
Median age	1.2	4.5	7.1	5.3	1.8
Maximum age	2.8	9.9	10.7	8.3	3.8

^aAges are given in years. See Figure 1 for station locations. Depths are given in meters.

[15] The halocline in about 70 to 100 m depth is poorly ventilated, with a median age of 3.8 years (Table 1). The corresponding minimum of interleaving in about 60 m depth is located somewhat higher in the water column (Figure 3). During the twentieth century, three stagnation periods (in the 1920/1930s, 1950/1960s, and 1980/1990s) with ages exceeding 8 years are found. Thereby, the latest stagnation period is the most pronounced one. As mentioned in section 1, the stagnation periods during the 1920/1930s and 1980/1990s are explained by stronger than normal freshwater inflow and zonal wind velocity [Meier and Kauker, 2003a]. They coincide with low salinities. The nature of the less pronounced stagnation period during the 1950/1960s (a period of high salinities) is not clear yet. Perhaps it might be caused by the very strong inflow in 1951 which filled the Baltic deep water with very high saline water. Consecutive saltwater inflows might not be saline enough to replace these earlier water masses.

[16] The high-saline water is propagating from Gotland Deep in two branches farther into Landsort Deep (BY31) in the northwestern Gotland Basin and into the Gulf of Finland. In general, at LL07 in the Gulf of Finland a pronounced halocline is observed (Figure 2). As the Gulf of Finland is relatively shallow with a maximum depth decreasing almost monotonically from 80–100 m at the entrance area to 20–30 m in the eastern part of the gulf, the halocline sometimes disappears due to estuarine flow reversal caused by relatively strong southwesterly winds [Elken *et al.*, 2003]. This does not happen very often, and consequently at LL07 a maximum age of 8.3 years is found (Table 1). However, during the stagnation period of the 1980/1990s the stratification is strongly reduced for a relatively long time. Between 1988 and 1994 the age at the bottom is significantly reduced. For the whole simulation period the median age amounts to 5.3 years.

[17] Between the Baltic proper and the Åland Sea, there is a wide sill with a depth of only 40 m intersected by a narrow channel with a sill depth of about 70 m [Stigebrandt, 2001]. Therefore, high-saline water cannot flow from the Gotland Basin via the Åland Sea into the Bothnian Sea. The stratification in the Bothnian Sea is much weaker compared to the other already discussed subbasins (Figure 2). During stagnation periods the stratification is significantly reduced. The variability of the age of the water masses between halocline and bottom is large. However, during the whole simulation period the age at SR5 never exceeds 4.2 years

(Table 1). The median age at the bottom amounts to 1.8 years. As in Bornholm Basin and Gotland Basin, in the halocline a secondary age maximum is found. It is believed that the deep water in the Gulf of Bothnia is mainly renewed by major inflows of Baltic proper surface water and not via deep thermal or haline convection [Marmefelt and Omstedt, 1993].

3.2. Sensitivity of Salinity and Age

[18] Increased amplitude of the sea level in Kattegat causes increased saltwater inflow into the Baltic. Stigebrandt [1983] found that approximately one half of the salt transport into the Baltic is carried out by the dispersive mode associated with barotropic fluctuations. Consequently, the salinity in the Baltic proper increases in the upper and lower layers (Figure 4).

[19] The impact of reduced freshwater inflow is similar, but the sensitivity of the salinity is much larger in terms of the relative change of the two forcing functions (but not necessarily in terms of transports). The reversed situation is found for reduced sea level oscillations and increased freshwater inflow. In these experiments the difference between the salinity change at the surface and the salinity change at the bottom is small compared to the magnitude of the changes. Thus, in a first approach, the salinity profile is just shifted toward a new salinity range with constant stability. This finding will be used below to estimate changing recirculation in steady state in the Baltic utilizing a simple conceptual model. If the freshwater inflow or the amplitude of the sea level is altered by 15%, the resulting change of the salinity profile is still outside the range of natural variability at all depths (not shown).

[20] The situation differs when the wind speed is altered. Increased (decreased) wind speed causes salinity to decrease (increase) at all depths (Figure 4). However, changes of the sea surface salinity (SSS) are rather small. If the wind speed changes by $\pm 15\%$, SSS will change within the range of natural variability only (not shown). Much larger changes are found for the salinity in the halocline and at the bottom, affecting stability significantly.

[21] The salinity profile at BY15 (Figure 4) may be used immediately to calculate the average salinity approximately taking the hypsographic function of the entire Baltic into account [Winsor *et al.*, 2001]. However, this approach was not applied here. Instead, the salinities of all grid boxes of the model domain (without Kattegat) were averaged (Figure 5).

[22] The sensitivity of the average salinity on freshwater inflow is largest compared to the sensitivities on wind speed and on Kattegat sea level. However, especially decreased wind speed causes a significant increase of the average salinity. SSS at BY15 is less sensitive to wind speed than average salinity or deep water salinity, in agreement with results by Stigebrandt [1983] (Figure 4). However, the sign of the response in process-oriented models is reversed [e.g., Stigebrandt, 1983; Gustafsson, 2000b]. Increased wind speed causes in our experiments decreased SSS at BY15 but increased SSS at LL07 in the Gulf of Finland and also in the Bothnian Bay (not shown).

[23] In all three types of sensitivity experiments (freshwater inflow, wind speed, and sea level in Kattegat) the changes of the age as function of depth and time at BY15

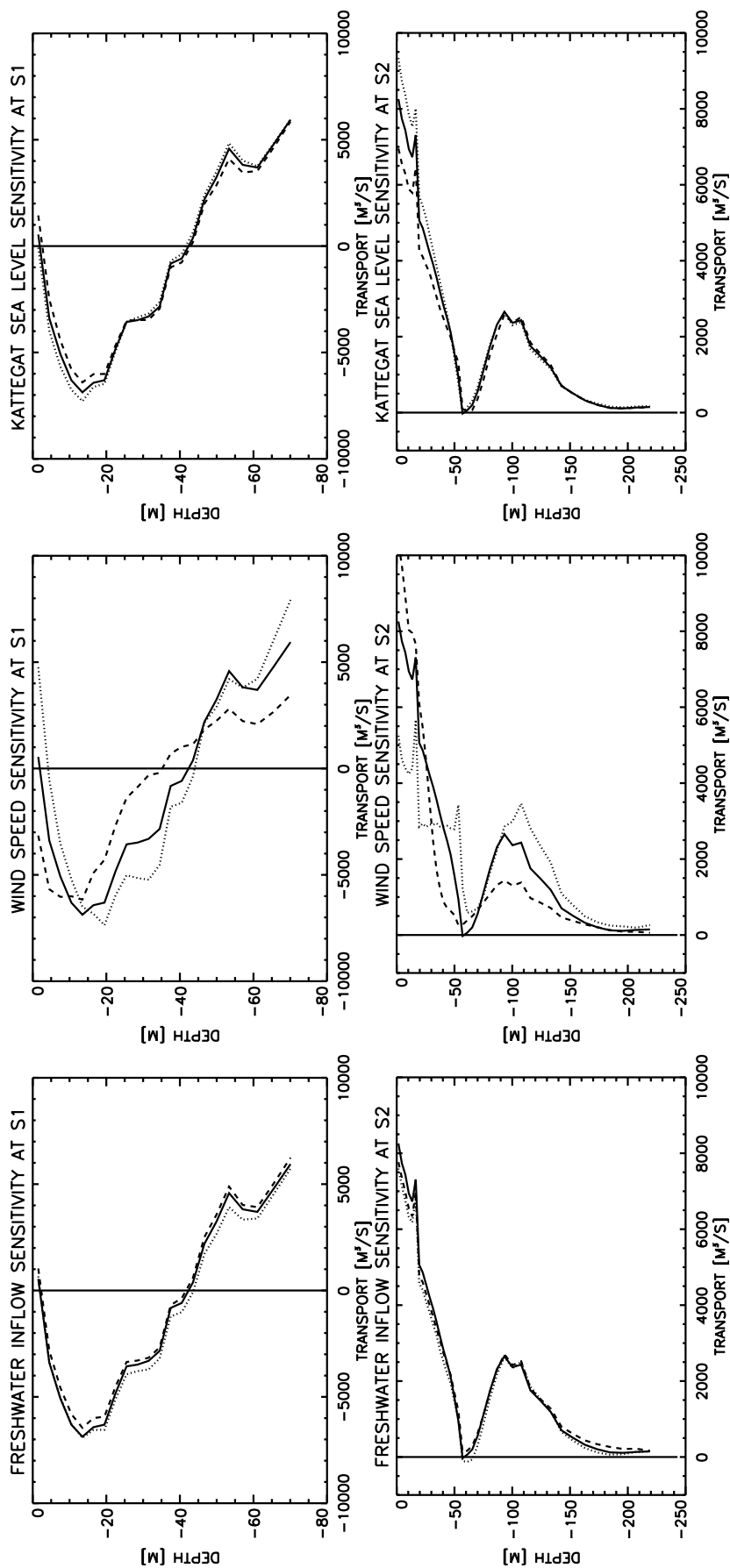


Figure 3. Mean horizontally integrated transports (in $\text{m}^3 \text{s}^{-1}$) across a basin-wide section at Stolpe Channel (S1) and in the eastern Gotland Basin between Gotland and Latvia (S2) for 1951–1998: standard run (solid line) and 30% increased (dotted line) and reduced (dashed line) freshwater inflow, wind speed, and sea level in Kattegat. At S1 (S2), eastward (northward) transports are counted positive. The positions of the sections are depicted in Figure 1.

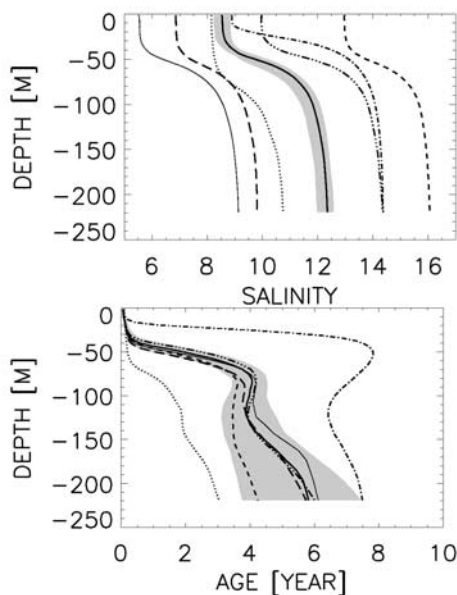


Figure 4. Median salinity and age profiles for 1951–1998 at BY15: standard run (solid line, shaded areas indicate the range between the first and third quartiles), increased (thin solid line), and reduced (dashed line) freshwater inflow, increased (dotted line) and reduced (dash-dotted line) wind speed, and increased (dash-triple-dotted line) and reduced (long-dashed line) amplitude of the sea level in Kattegat. Changes amount to $\pm 30\%$.

are largest during stagnation periods in the deep water below 100 m and during the whole simulation period in the upper part of the halocline (not shown).

[24] During 1951–1998 the vertical median age profile at BY15 has a local maximum within the perennial halocline of about 4 years (Figure 4). The largest median age of about 6 years is found at the bottom with a variability range between 4 and 8 years, approximately (note that for the whole period 1903–1998 the median age at the bottom of BY15 is about 1.5 years shorter; see Table 1). The vertical median age profiles of the sensitivity experiments with $\pm 30\%$ altered freshwater inflow and $\pm 30\%$ altered sea level in Kattegat are within the range of natural variability at all depths. However, in the experiments with increased or decreased wind speed, changes much larger than the natural variability are found (Figure 4). Similar results are calculated for the average age of the Baltic Sea (Figure 5). Thus the major factor affecting the ventilation of the Baltic deep water and, consequently, the age in steady state is the wind speed. With respect to freshwater inflow, the average age has a maximum in present climate (Figure 5). Both increasing and decreasing freshwater supplies result in shorter average ages.

[25] Horizontal distributions of the vertically averaged salinities and seawater ages of the standard experiment and of the sensitivity experiments with 30% increased external forcing are shown in Figures 6 and 7. In the standard experiment the vertically averaged salinity has a maximum at Gotland Deep (Figure 6, top left panel). Largest ages are found in the northwestern Gotland Basin and in the northern Bothnian Sea (Figure 6, top right panel). The distribution of the age is mainly determined by the topography. Increased

freshwater inflow has the largest impact on salinity in the Belt Sea (Figure 6, bottom left panel). The changes at the outermost ends of the gulfs are smallest. In the Arkona Basin, Bornholm Basin, Gotland Basin, Bothnian Sea, and western Gulf of Finland the differences are surprisingly homogeneous and amount to about 2‰. By contrast, the vertically averaged age does not change significantly (Figure 6, bottom right panel). Only in the northern Bothnian Bay, reductions of 1–2 years at the maximum are found. Increased saltwater inflow through the Danish straits causes an increase of the vertically averaged salinity of 1–2‰ in the entrance area and in the deeper parts of the Arkona Basin, Bornholm Basin, Baltic proper, and western Gulf of Finland (Figure 7, top left panel). In this case the impact on the vertically averaged age is even smaller compared to the case of increased freshwater inflow (maximum differences are less than half a year) (Figure 7, top right panel). Finally, increased wind speed has a significant impact on salinity (Figure 7, bottom left panel). In the deeper parts of the Baltic the vertically averaged salinity decreases most with up to 1–2‰. However, in the coastal areas, in the Baltic entrance area, and in the Gulf of Riga the vertically averaged salinity increases. The largest increase of salinity is found again in the Belt Sea. In the entire model domain, vertically averaged ages are smaller than 2 years (Figure 7, bottom right panel). A maximum reduction is found in the northwestern Gotland Basin. Summarizing, in the cases of altered saltwater and freshwater inflows the salinities change significantly but the ages are not much affected. By contrast, increased wind speed affects both salinities and especially ages considerably.

3.3. Transient Behavior in Sensitivity Experiments

[26] To better understand the physical mechanisms involved, in this section the transient behaviors of salinity (Figure 8) and volume transport into the Baltic deep water (Figure 9) in the sensitivity experiments are discussed.

3.3.1. Freshwater Inflow

[27] When the freshwater supply to the Baltic is increased, the additional volume is flushed out of the Baltic at once and the horizontal salinity fronts in Kattegat move toward north. Thus, during the first months of the integration, the largest salinity change occurs in the Belt Sea due to the changing fronts whereas the sea level in the Baltic inside the Danish straits does not change significantly (not shown). As the inflowing water in the Danish straits has a lower salinity in the sensitivity experiment compared to the reference experiment, the volume transport into the Baltic deep water is significantly smaller (Figure 9). During this first phase after the freshwater supply has been increased, the deep water is less frequently ventilated, causing stagnation. The reduced salt transport is giving rise to decreasing deep water salinity. Indeed, surface and bottom salinities at Gotland Deep decrease with the same decay rate, approximately (Figure 8). During the second phase the volume transport into the Baltic deep water is recovered to the level of the reference experiment with an e-folding time of about 28 years. Thus the dominant process at the beginning of the sensitivity experiment is an increased dilution of inflowing high-saline Kattegat water with outflowing Baltic Sea water in the Danish straits, i.e., an increased recirculation. The recovery of the volume transport into the Baltic deep water

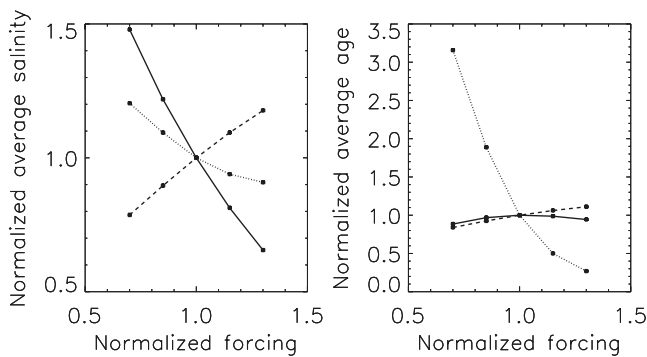


Figure 5. Normalized total mean average salinity and age for 1951–1998 in sensitivity experiments with modified forcing: freshwater inflow (solid line), wind speed (dotted line), and sea level in Kattegat (dashed line). (Note the different scales of normalized average salinity and age.)

might be explained by the changing horizontal salinity gradients caused by the overall dilution. Owing to decreasing salinity in the Baltic Sea, water of less intensive saltwater inflows is now flowing into the Baltic proper into that depth where its buoyancy disappears. In the present climate, these smaller inflows are not able to ventilate the deep water. This “salinity inflow feedback” has been described by Meier [2002] already.

3.3.2. Sea Level Kattegat

[28] Qualitatively similar results were found in the experiments with decreasing amplitude of the sea level in Kattegat compared to the experiments with increasing freshwater inflow. At the beginning the volume flow of high-saline water is reduced (Figure 9), causing the salinity in the Baltic to decrease. However, the reduction of the deep water salinity is larger than the reduction of the SSS (Figure 8), which might be explained by the following hypothesis. Since the salt flux which is associated with barotropic fluctuations and which leaves the Baltic is also reduced, in the new steady state the vertical net flux of salt must be smaller. Consequently, the stability in the new steady state will be somewhat reduced compared to the reference experiment. After the initial drop the volume transport at the bottom of Stolpe Channel is recovered with an e-folding time of about 11 years.

3.3.3. Wind Speed

[29] The results of the sensitivity experiments for the age of the Baltic deep water (Figures 4 and 5) can be explained by the results for the horizontally integrated transports at Stolpe Channel and Gotland Deep (Figure 3). Changes of the freshwater and saltwater inflow cause the system to shift into a new quasi steady state in which interleaving does not differ very much from the standard experiment. In these experiments the halocline depth does not change significantly. By contrast, changes of the wind speed have large impacts on the vertical stratification and on the halocline depth. This is not astonishing, as wind-induced mixing scales with the third power of the surface wind speed. For instance, when the wind speed is increased, at the beginning the vertical flux of salt will increase. This will cause the halocline in the entire Baltic to deepen with increasing SSS and decreasing deep water salinity (Figure 8). After the initial transition period the vertical salt flux will decrease

again. In the new steady state, SSS and the stability at BY15 are lower compared to the standard experiment (Figure 4). Increased wind speed causes increased volume transport of less saline water in the bottom layer at Stolpe Channel and increased interleaving at Gotland Deep mainly in about 100–110 m depth (Figures 3 and 9). Even deeper layers down to the bottom are affected. Although salinity in the upstream basin (Bornholm Basin) also decreases, the salt transports of the new inflowing water into the Gotland Basin deep water are slightly increased with a shift of the interleaving maximum toward greater depths (not shown). Owing to the enhanced volume transport, the age of the deep water is reduced compared to the standard run (Figure 4). The fact that the fresher inflowing water is younger in the new steady state cannot explain the younger deep water in the Gotland Basin. At BY5 the age in 60 m depth (the sill depth) is reduced by less than half a year in case of 30% increased wind speed (not shown) whereas at BY15 the ages in 100 m depth (the maximum of interleaving) and at the bottom are reduced by about 2.5 and 3 years, respectively (Figure 4). Thus the dominant effect is the increase of horizontal volume transports into the Gotland Basin (Figure 3).

[30] After about 20–30 years the decrease of SSS and bottom salinity is significantly weakened (Figure 8). In the bottom layer the salinity (Figure 8) and transport anomalies (Figure 9) show pronounced decadal oscillations which are anticorrelated to decadal anomalies of the zonal wind velocity [Meier and Kauker, 2003a]. The decadal oscillations of the volume transport in the sensitivity experiment are relatively stronger compared to the reference experiment because the wind speed is multiplied with a constant factor; that is, the low-frequency component of the zonal wind velocity is also increased.

4. Conceptual Model Results

[31] For the discussion of the results of the previous section the conceptual model of Stigebrandt [2003] is applied (Appendix A). The model is based upon the conservation of volume and salinity and includes important processes like the recirculation of Baltic proper surface water and entrainment of ambient water causing the volume flow of new deep water to increase on its path from the sills toward the deep water in the Baltic proper when the buoyancy disappears. The model will enable calculation of the volume flows and salinities of the inflowing and outflowing water of the Baltic and even the salinity of the Baltic proper deep water at interleaving if the freshwater supply, the salinity of the Kattegat deep water, the volume flow of Kattegat deep water into the Baltic, and the rates of recirculation and entrainment are known (see Appendix A). The model implies that SSS of the Baltic proper is independent of the rate of recirculation and wind-induced mixing in the Baltic. The latter conclusion is approximately confirmed by RCO results because SSS is only slightly dependent on the wind speed (Figure 4). At least, the sensitivity of SSS on wind speed (not shown) is much smaller compared to the sensitivity of the average salinity on wind speed (Figure 5).

[32] The results of the sensitivity experiments using RCO indicate that the stability (i.e., the difference between

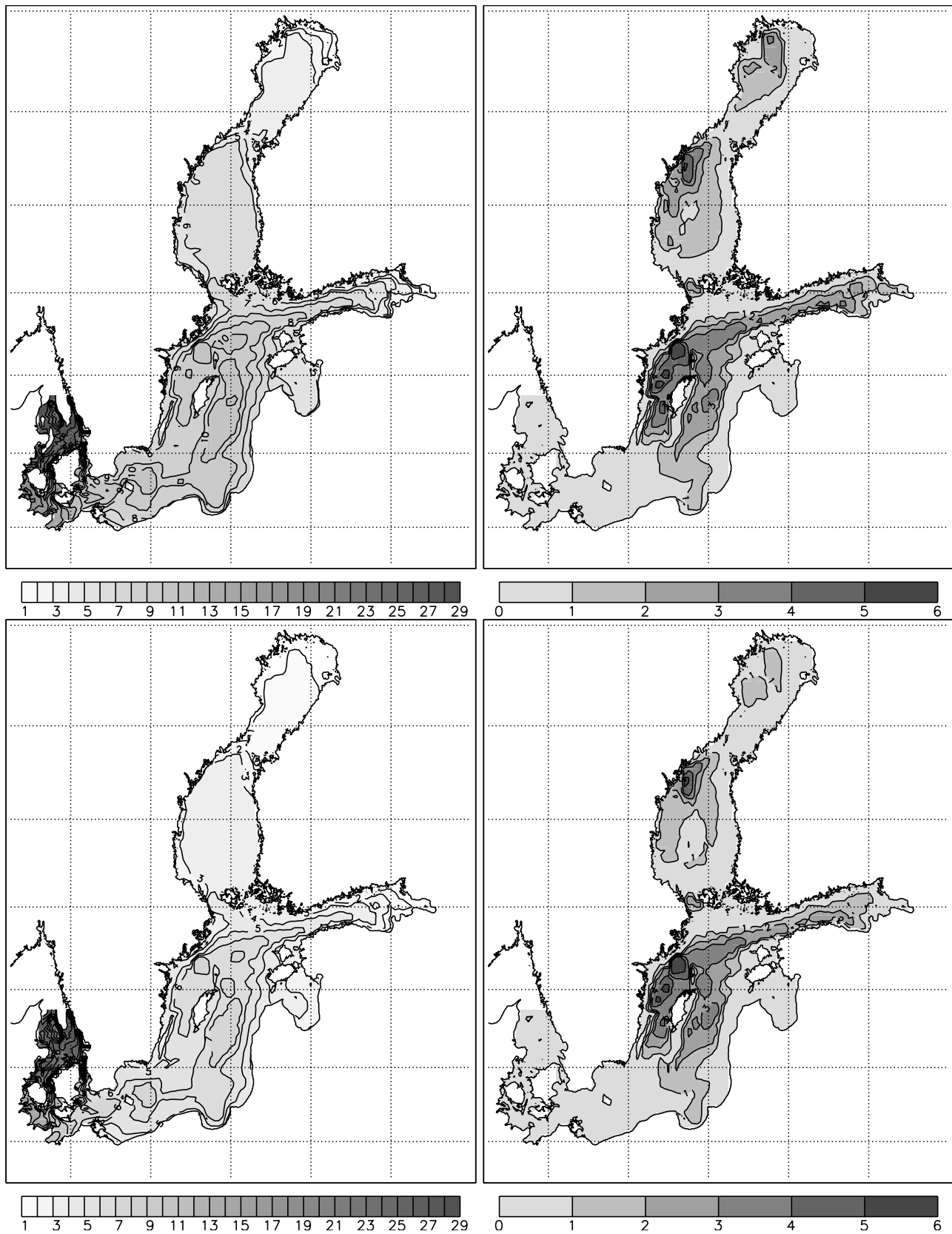


Figure 6. (left) Vertically averaged salinity (in ‰) and (right) age (in years) for 1951–1998; (top) standard experiment and (bottom) sensitivity experiment with 30% increased freshwater inflow.

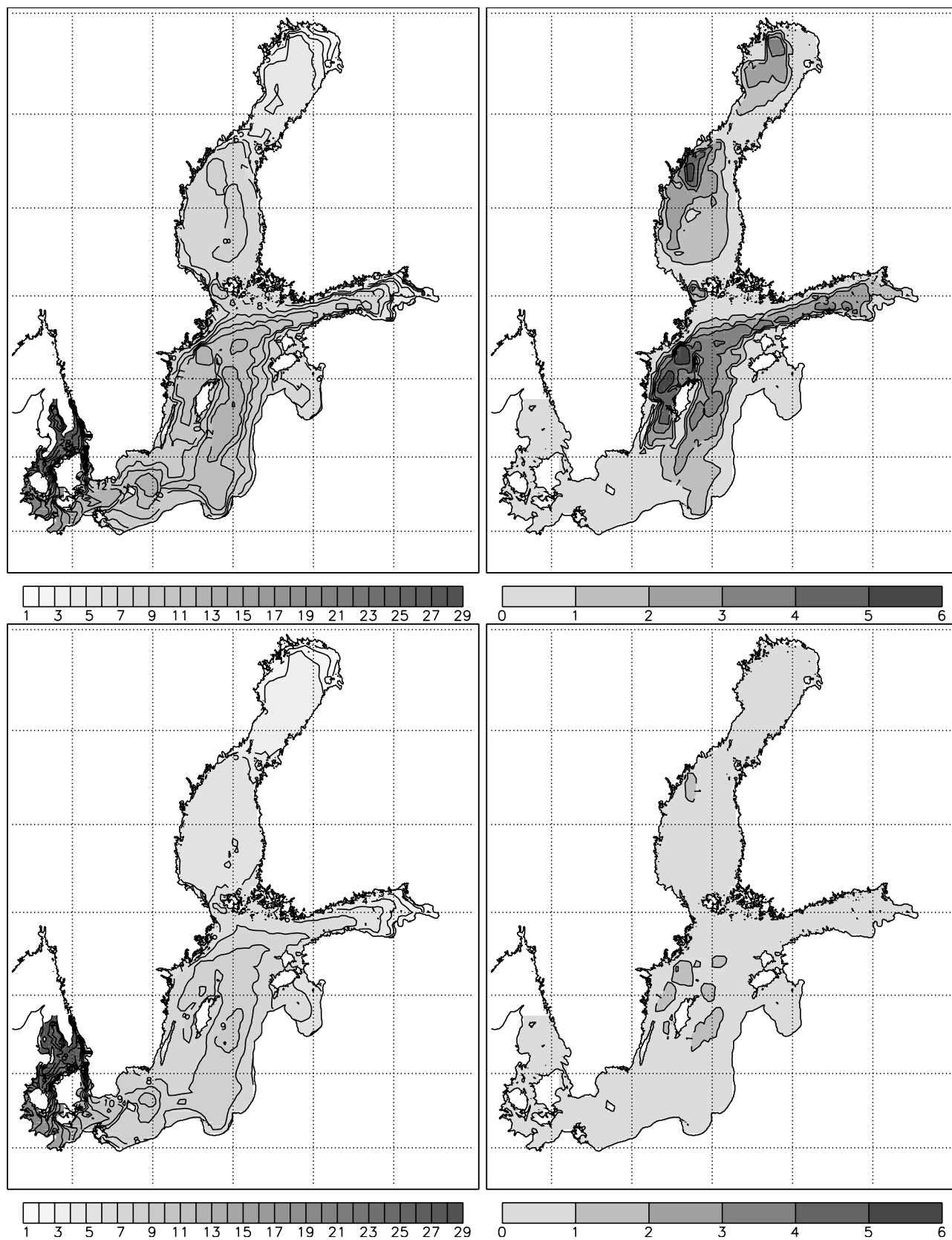


Figure 7. Same as Figure 6 but for sensitivity experiments with (top) 30% increased sea level amplitude in Kattegat and (bottom) 30% increased wind speed, respectively.

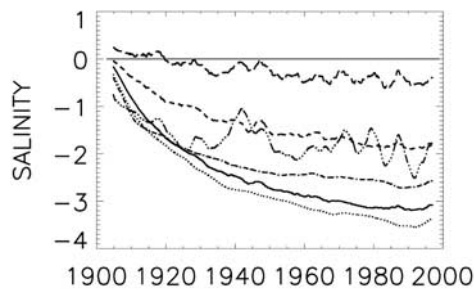


Figure 8. Four-year running mean surface and bottom salinity differences at BY15 between sensitivity experiments and the reference experiment: 30% increased freshwater inflow (surface: solid line, bottom: dotted line), 30% reduced sea level in Kattegat (surface: dashed line, bottom: dash-dotted line), and 30% increased wind speed (surface: long-dashed line, bottom: dash-triple-dotted line).

surface and deep water salinity) will approximately be invariant in steady state if the saltwater and freshwater inflows change within a range of less than $\pm 30\%$. Stability seems to be stabilized via a negative feedback. For instance, if freshwater supply is increased, SSS will decrease. As the entrainment rate is inversely proportional to stability [Stigebrandt, 1987] (see Appendix A), entrainment and consequently the flow rate of new deep water will decrease. Thus the deep water salinity will be reduced. The opposite argumentation chain may be applied if the freshwater supply is decreased. Since in the sensitivity experiments the shape of the profiles changes slightly with time compared to the standard experiment (Figures 4 and 8), the system may be not yet in steady state. In addition, the simplified approach of this section may not consider all the nonlinearities of the more complex RCO model. Nevertheless, assuming that in steady state, stability is invariant, the recirculation factor as a function of the freshwater supply can easily be calculated (Appendix B). The result suggests that the recirculation factor is not very sensitive to the freshwater inflow if the absolute change does not exceed 30%. (Figure 10). However, the impact of decreased freshwater inflow is larger than the impact of increased inflow. The sensitivities of both SSS and deep water salinity are smaller in the analytical model than in RCO (Figure 11).

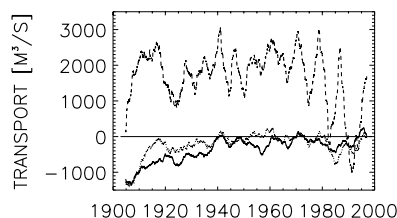


Figure 9. Four-year running mean horizontally integrated transport differences (in $\text{m}^3 \text{s}^{-1}$) at the bottom of Stolpe Channel across a basin-wide section (S1) between sensitivity experiments and the reference experiment: 30% increased freshwater inflow (solid line), 30% reduced sea level in Kattegat (dotted line), and 30% increased wind speed (dashed line).

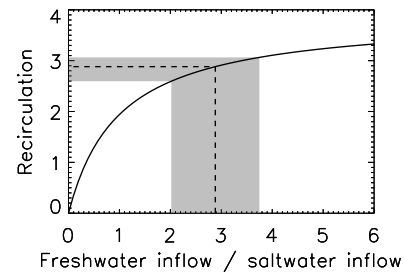


Figure 10. Recirculation factor as a function of the normalized freshwater inflow, i.e., the ratio between freshwater and saltwater inflows. Dashed lines show present climate using standard figures [Stigebrandt, 2003]. The shaded area indicates the range between 30% increased and reduced freshwater inflows.

[33] Owing to the assumption of invariant stability the sensitivity of the deep water salinity on freshwater inflow is independent of the entrainment rate. The entrainment in Figure 11 is chosen such that the deep water salinity of the analytical model and the salinity of RCO at BY15 in 100 m depth coincide. The SSS results of the process-oriented model by Stigebrandt [1983] are in between the analytical solution and RCO results. The solutions outside the $\pm 30\%$ range should not be considered because some of the utilized assumptions will break down if the climatic perturbations are too large. For instance, the open boundary condition in Kattegat used in RCO may not be valid in case of large perturbations [Meier and Kauker, 2003b]. The same

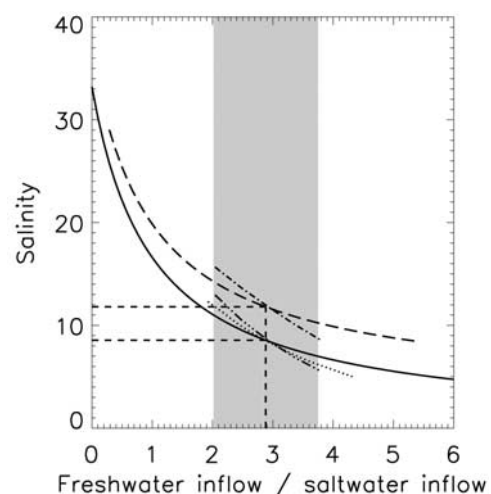


Figure 11. Total mean sea surface salinity (SSS) and deep water salinity (in ‰) as a function of the normalized freshwater inflow: SSS (solid line) and deep water salinity (long-dashed line) of the analytical model following Stigebrandt [2003], SSS of the process-oriented model by Stigebrandt [1983] (dotted line), and mean SSS in 1.5 m depth (dash-triple-dotted) and deep water salinity in 100 m depth (dash-dotted line) at BY15 of the RCO model for 1951–1998. Dashed lines show present climate using standard figures [Stigebrandt, 2003]. The shaded area indicates the range between 30% increased and reduced freshwater inflows.

might be true for the process-oriented models. Despite of discrepancies between the results of the different model approaches, the overall agreement is satisfactory. Hence the conceptual model is a useful tool to understand complex model results. However, one has to keep in mind that some of the assumptions made in this section might be questionable. For example, the volume flow of Kattegat deep water into the Baltic is very likely not constant but smaller in case of increased freshwater inflow.

5. Discussion

[34] The experiments of this study differ from the sensitivity experiments performed by *Meier and Kauker* [2003a]. First, the changes of the external forcing were applied during the whole simulation period 1903–1998, and the analysis was performed for 1951–1998 when the system is in steady state, approximately. By contrast, *Meier and Kauker* [2003a] performed experiments with either climatological monthly mean or with high-pass filtered (with a cutoff period of 4 years) external forcing. Thus the response of the Baltic Sea on decadal timescales between 4 and 16 years was studied; that is, the timescale of the forcing anomalies is smaller than the typical response timescale (see section 1). Second, in this study the wind speed was multiplied by a constant factor affecting all frequencies, whereas *Meier and Kauker* [2003a] removed only the low-frequency part of the wind spectrum.

[35] Comparing the two sets of experiments with changing freshwater supply it is found that “short” lasting anomalies cause stagnation whereas in steady state deep-water ventilation is almost invariant. In both sets of experiments with either an additional low-frequent zonal wind component of about $1\text{--}2\text{ m s}^{-1}$ [*Meier and Kauker*, 2003a] or with increased wind speed (this study), salinity in the Baltic is smaller compared to the long-term mean. In the first case, salinity decreases because an anomalous barotropic pressure gradient hampers saltwater inflows through the Danish straits. In addition, the ventilation of the deep water is significantly reduced because the lower layer volume transport through Stolpe Channel is reduced by the compensating flow of an anomalous Ekman transport [*Meier and Kauker*, 2003a]. Although this process causes the vertical salt flux in the western Baltic to increase, the average salinity is very likely not affected. By contrast, in the sensitivity experiments of this study, salinity of the Baltic is in quasi steady state and ventilation is increased because the increased high-frequency part of the wind speed causes the lower layer volume transport through Stolpe Channel to increase. This happened after a few years of integration.

[36] To elucidate the mechanism further, an additional sensitivity experiment was performed. During the whole integration period 1903–1998 a stationary zonal wind anomaly of $+1\text{ m s}^{-1}$ was added. The results show a significant decrease of salinity comparable to the results with 15% increased freshwater inflow. The stagnation period in the 1920/1930s is intensified and prolonged. However, in steady state the changes of age are small. For 1951–1998, differences of the median age between sensitivity and standard experiments are less than half a year. Owing to the barotropic pressure gradient associated

with the zonal wind stress anomaly, the saltwater inflow through the Danish straits is reduced. For 1951–1998 the mean sea level difference between the sensitivity experiment with additional zonal wind anomaly and the standard experiment amounts to 3.6 cm, on average. The sea level difference between Arkona Basin and Kattegat caused by the additional zonal wind anomaly is about 2–3 cm. By contrast, for 1951–1998 the mean sea level difference between the experiment with 15% increased freshwater inflow and the standard experiment amounts only to 0.6 cm on average. Maximum sea level differences of about 1 cm are found in the Belt Sea. A monotonously increasing gradient between Kattegat and Baltic proper caused by the additional freshwater inflow does not exist confirming the hypothesis that in this case, recirculation is the dominant process for changes of basin-wide averaged salinity.

6. Conclusions

[37] 1. A passive tracer describing the age of water masses is a useful tool to analyze results of 3-D ocean circulation models. The tracer results suggest that during 1903–1998 the ventilation of the Baltic deep water is realistically simulated. The advantage of the applied 3-D method compared to basin-integrated process oriented models is that the timescales of horizontal advection and dispersion are explicitly included.

[38] 2. In steady state, simulated sensitivities of salinity to perturbations of freshwater inflow, wind speed, and sea level elevation in Kattegat utilizing a 3-D ocean circulation model are in good agreement with results from process oriented models.

[39] 3. Changes of the deep water ventilation depend on the timescale of the applied forcing anomaly. Although long-term changes of freshwater and saltwater inflows and of low-frequency wind anomalies cause the estuary to drift into a new steady state with altered salinity, stability and ventilation are invariant approximately. For that, the timescale of the perturbation needs to be long compared to the turnover time of the freshwater content. By contrast, long-term changes of the high-frequency wind affect deep water ventilation significantly.

Appendix A: Conceptual Model

[40] In steady state the conservation of volume and salinity in an estuary with a two-layer flow yields [*Knudsen*, 1900]

$$Q_1 - Q_{in} = Q_f \quad (\text{A1})$$

$$S_1 Q_1 - S_{in} Q_{in} = 0. \quad (\text{A2})$$

Here Q_f , Q_{in} , and Q_1 are the volume flows of the freshwater supply and of the inflowing and outflowing water of the Baltic, respectively. S_1 and S_{in} denote the long-term mean salinity of the outflowing surface water and of the inflowing water to the Baltic, respectively. The inflowing water is a mixture of surface water from the Baltic proper of salinity S_1 and of Kattegat deep water of salinity S_0 . If the

Table B1. Standard Figures of Present Climate Used in Appendix B^a

Symbol	Meaning	Value
Q_f	freshwater inflow	15,000 m ³ s ⁻¹
Q_0	volume flow of Kattegat deep water into the Baltic	5,203 m ³ s ⁻¹
Q_{in}	volume flow into the Baltic deep water	15,178 m ³ s ⁻¹
S_0	salinity of the Kattegat deep water	33.2‰
S_1	surface salinity of the Baltic proper	8.55‰
S_{in}	salinity of inflowing water	17‰
F	recirculation rate	2.918
E	entrainment rate	2.6

^a Q_0 , Q_{in} , and F are calculated from the other parameters.

recirculating volume flow of Baltic surface water and the volume flow of seawater are Q_r and Q_0 , respectively, the conservation of volume and salinity will be written as

$$Q_{in} = Q_0 + Q_r \quad (A3)$$

$$S_{in} Q_{in} = S_0 Q_0 + S_1 Q_r. \quad (A4)$$

From (A1) to (A4) an expression for S_1 is obtained,

$$S_1 = S_0 \frac{Q_0}{Q_f + Q_0}. \quad (A5)$$

The rate of recirculation F is defined as

$$Q_{in} = F Q_0. \quad (A6)$$

Using (A1), (A3), (A4), and (A6), it follows

$$S_{in} = \frac{1}{F} (S_0 + (F - 1) S_1). \quad (A7)$$

The inflowing new deep water Q_{dw} is modified by entrainment of Baltic proper surface water. If the rate of entrainment E is defined as

$$Q_{dw} = E Q_{in}, \quad (A8)$$

the salinity S_{dw} at interleaving will be

$$S_{dw} = \frac{1}{E} (S_{in} + (E - 1) S_1). \quad (A9)$$

As entrainment will differ according to the salinity range considered, *Stigebrandt* [2003] introduced three salinity classes, i.e., three equations of the types (A8) and (A9). This will not be done here because for the discussion, only the processes are important. The equations enable calculation of Q_{in} , Q_1 , S_{in} , S_1 , and S_{dw} as a function of Q_f , S_0 , Q_0 , F , and E . The rate of recirculation F can be estimated because stability is approximately invariant to changes of the freshwater inflow Q_f (Appendix B). Following *Stigebrandt* [1987] the entrainment velocity w_e is assumed to be related to the wind speed W by

$$w_e \propto \frac{W^3}{g \beta (S_{dw} - S_1) H}. \quad (A10)$$

Here $\beta \approx 8 \times 10^{-4}$ is the coefficient of salt contraction of seawater, g is the acceleration of gravity, and H is the halocline depth. For further details the reader is referred to *Stigebrandt* [2003].

Appendix B: Recirculation

[41] Stability σ is defined as

$$\sigma = S_1 - S_{dw}. \quad (B1)$$

Assuming that the recirculation is a function of freshwater inflow and that the stability is invariant to freshwater inflow in steady state,

$$\Delta \sigma = \frac{\partial \sigma}{\partial Q_f} \Delta Q_f + \frac{\partial \sigma}{\partial F} \Delta F = 0, \quad (B2)$$

from equations (A5), (A7) and (A9) we derived

$$\frac{dF}{dQ_f} = - \frac{\partial \sigma}{\partial Q_f} / \frac{\partial \sigma}{\partial F} = \frac{F}{Q_f} \frac{Q_0}{Q_f + Q_0}. \quad (B3)$$

Solution of (B3) is

$$\frac{F}{F_0} = \frac{x}{x+1} / \frac{x_0}{x_0+1}, \quad (B4)$$

with $x = Q_f/Q_0$ and with the constants x_0 and F describing present climate. Equation (B4) is used to calculate S_{dw} shown in Figure 11. Thereby, standard figures of present climate following *Stigebrandt* [2003] are utilized (Table B1).

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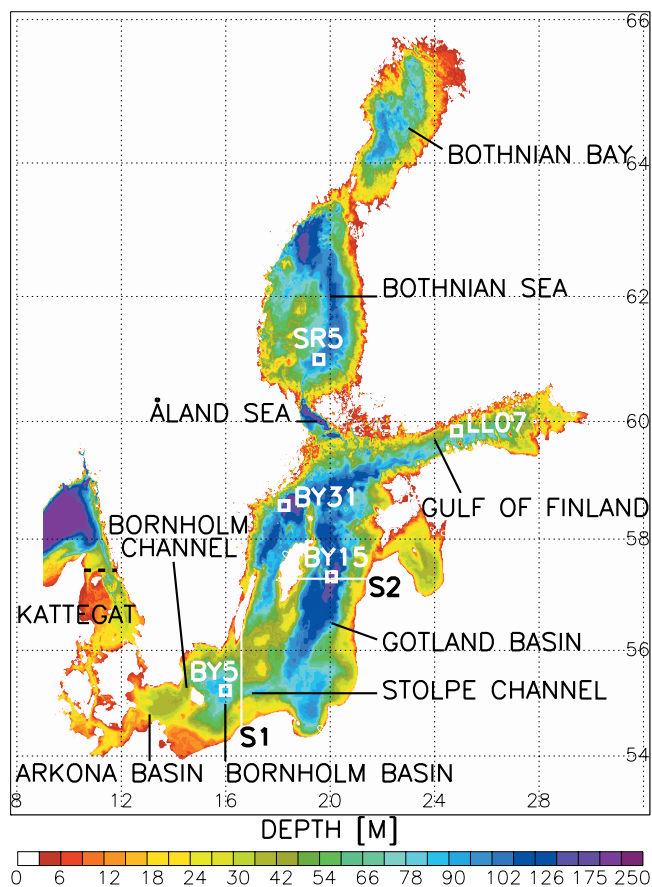


Figure 1. Bottom topography of the Baltic Sea. The domain of the Rossby Centre Ocean model (RCO) is limited with open boundaries in the northern Kattegat (dashed line). Selected stations (BY5, BY15, BY31, LL07, SR5) and cross sections (S1 and S2, white lines) are depicted additionally.

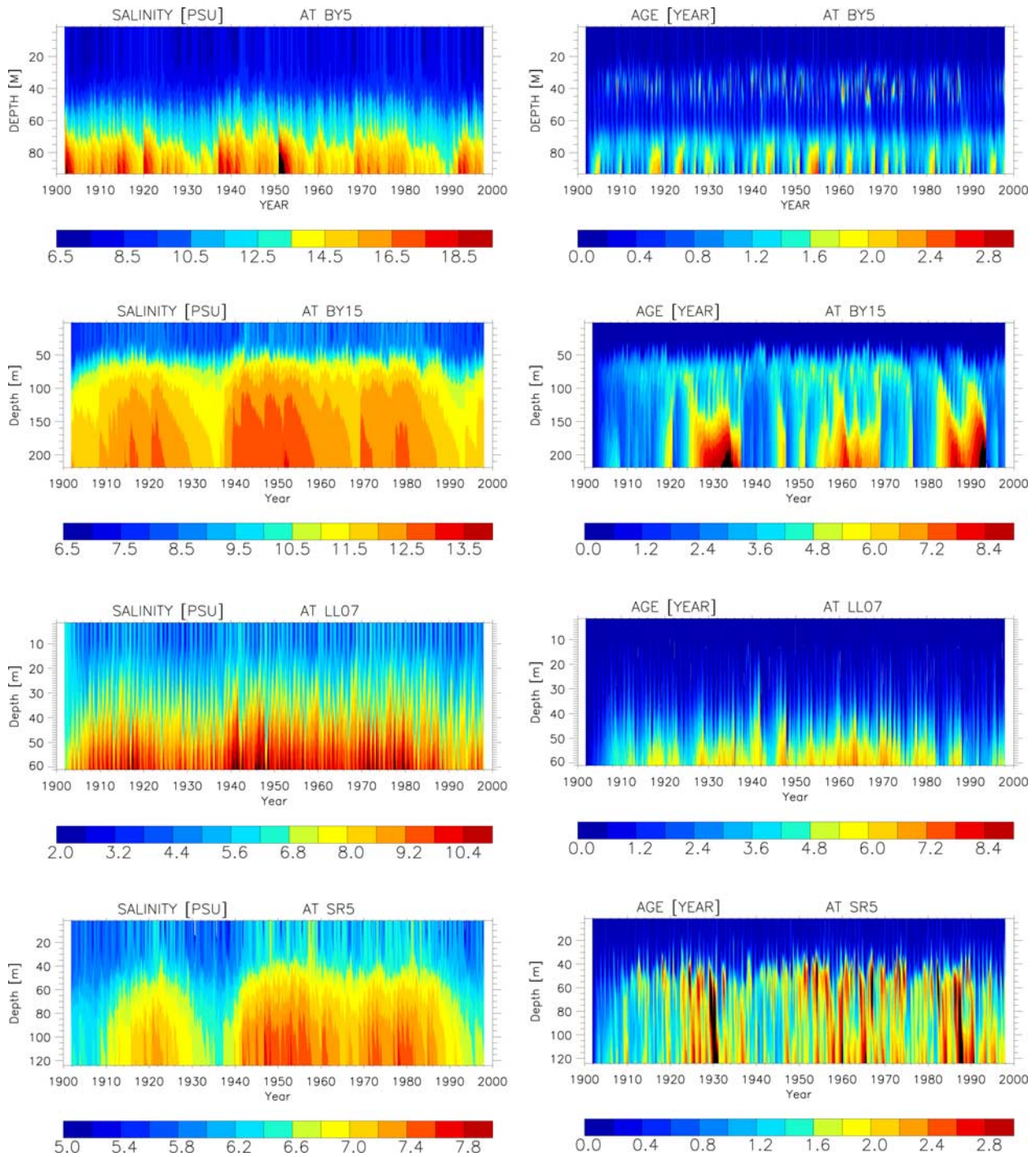


Figure 2. (left) Salinity and (right) age as a function of depth and time at selected stations (see Figure 1).