The major Baltic inflow in January 2003 and preconditioning by smaller inflows in summer/autumn 2002: a model study

OCEANOLOGIA, 46 (4), 2004.
pp. 557–579.

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KEYWORDS
Baltic Sea
Numerical modeling
Major Baltic inflow
Climate variability
Bottom boundary layer

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Manuscript received 29 September 2004, reviewed 15 November 2004, accepted 23 November 2004.

Abstract

Using the results of the Rossby Centre Ocean model (RCO) the Baltic inflows in summer/autumn 2002 and January 2003 have been studied. The model results were extracted from a long simulation with observed atmospheric forcing starting in May 1980. In RCO a bottom boundary layer model was embedded. Both the smaller inflows and the major inflow in January 2003 are simulated in good agreement with observations. We found that a total of 222 km^3 water entered the Baltic in January; the salinity of 94 km^3 was greater than 17 PSU. In August/September 2002 the outflow through the Sound and inflow across the Darss Sill were simulated. The net inflow volume amounted to about 50 km^3.

1. Introduction

The deepwater of the Baltic Sea is ventilated by inflows of high saline water from the Kattegat through the narrow and shallow Danish Straits

(Fig. 1). Usually only larger, so-called major Baltic inflows renew the deepwater in the Gotland Basin significantly (e.g. Matthäus & Franck 1992, Fischer & Matthäus 1996). During the last century such events occurred more or less randomly at intervals of one to several years. However, during the past two decades the frequency of major inflows has decreased. Significant inflows have occurred only 3 times: in 1983, 1993, and 2003. The lack of saltwater inflows caused stagnation periods with decreasing oxygen and increasing hydrogen sulphide concentrations. Stagnation periods have

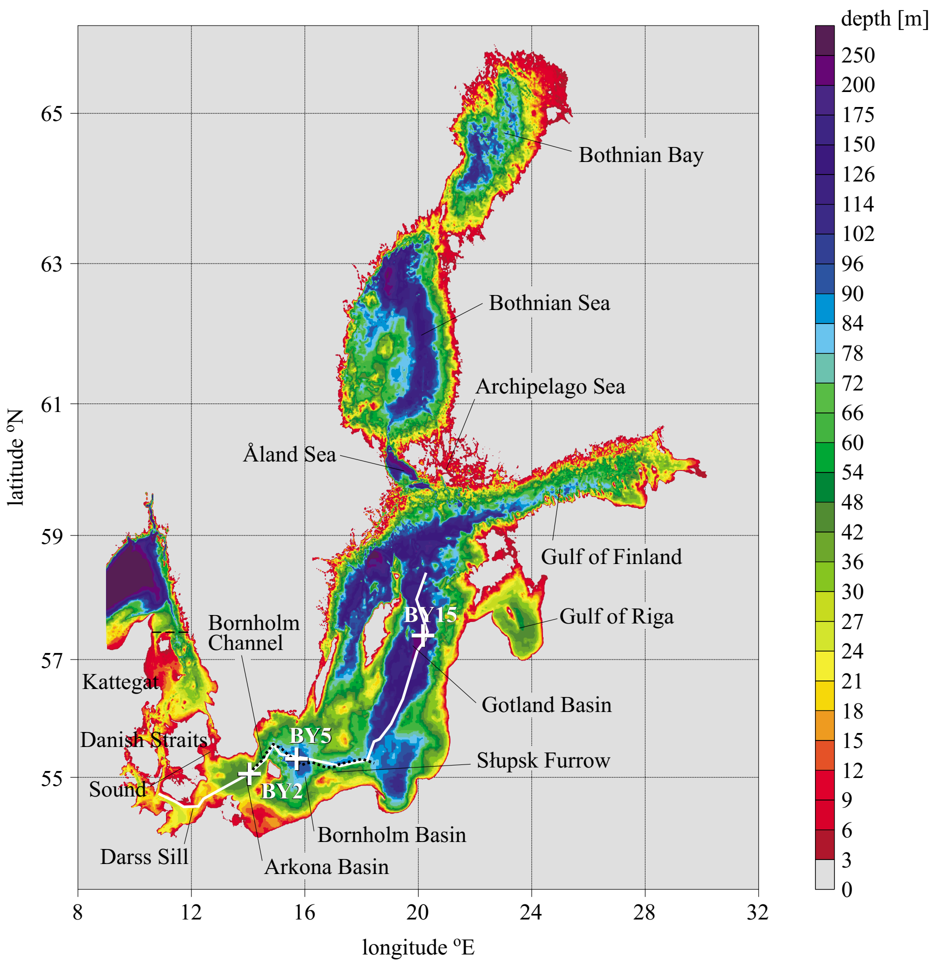

Fig. 1. Bottom topography of the Baltic Sea. The domain of the Rossby Centre Ocean model (RCO) is delimited by the open boundaries in the northern Kattegat (dashed black line). Selected monitoring stations (BY2, BY5, BY15) and the cross-sections from Fig. 5 (solid white line) and Fig. 6 (dotted black line) are additionally depicted

a significant impact on the marine ecosystem. Therefore, their causes need special attention. The last major inflow in January 2003 was described in great detail (Feistel et al. 2003a, Piechura & Beszczyńska-Möller 2003). The January 2003 inflow was enhanced through smaller saltwater inflows in August/September 2002, November 2002, and March 2003 (Feistel et al. 2003b). Especially the event in summer 2002 gained attention. Although this inflow was not driven by westerly gales, the exceptionally warm water left traces even in the deepwater of the western Gotland Basin (Feistel et al. 2003b). Tidal pumping was suggested as a possible driving mechanism (Feistel et al. 2003b, 2004). However, due to the lack of observations the dynamic details are still unclear. This is a challenge for model applications. In previous modeling studies saltwater inflows like the one in January 1993 were studied intensively (e.g. Huber et al. 1994, Lehmann 1995, Meier 1996, Andrejev et al. 2002, Meier et al. 2003, Meier & Kauker 2003, Lehmann et al. 2004). It was concluded that state-of-the-art ocean general circulation models (OGCMs) with geopotential vertical coordinates simulate major inflows into the Baltic rather well but smaller inflows into the Bornholm Basin, for instance, are systematically underestimated in long simulations (Meier et al. 2003). Therefore, in this study a bottom boundary layer model was embedded into an OGCM to study the major Baltic inflow in January 2003 and the connected smaller inflows.

2. Methods

2.1. Ocean model description

The simulations were performed with the Rossby Centre Ocean model (RCO) (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. A k-turbulence closure scheme for vertical mixing is embedded. The model domain covers the Baltic Sea including the Kattegat (Fig. 1). In the present study, RCO was used with a horizontal resolution of 2 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 May 1980. Hourly sea level data were prescribed at the open boundary in the Kattegat. In the case of inflow climatological temperature and salinity profiles were nudged at the open boundary. In the case of outflow a radiation condition was used. Simulations were performed for May 1980 to December 2003 with monthly mean river runoff calculated with a large-scale hydrological model (Graham 1999). The atmospheric forcing was calculated from the 3-hourly sea level pressure, 2 m air temperature, 2 m specific humidity, precipitation, and total cloudiness fields available at the Swedish Meteorological and Hydrological Institute (SMHI). For further details of the model setup the reader is referred to Meier et al. (2003).

2.2. Bottom boundary layer

The bottom boundary layer in RCO is based on the approach of Beckmann & Döscher (1997), refined by Döscher & Beckmann (2000). The approach addresses the known problem of artificial dilution of water masses descending down a sill in geopotential ('z') coordinate models with step-like topography such as RCO. In more detail, dense water passing over a sill cannot spread along its isopycnals. It is violently mixed as a result of static instability situations. A convection algorithm or an increased vertical diffusivity leads to a vertical homogenization of participating model boxes. In many cases, this results in a bottom density distinctly lower than is the case in observed situations. Artificial dilution can in general be prevented by resolving the bottom boundary layer (computationally unrealistic for typical Baltic Sea or large-scale applications) or, as in this study, by parameterizing the dense water flow. A detailed description of the bottom boundary layer model (BBL) used in RCO is given in the Appendix.

2.3. Experimental strategy

From a long hindcast simulation for May 1980 to December 2003 model results of the major Baltic inflows in January 1993 and 2003 are analyzed and compared. Especially the periods May 1992 to December 1993 and May 2002 to December 2003 are considered. In addition to the analysis of the major inflows, the consequences of the smaller summer inflows in August/September 2002 and November 2002 for the deepwater ventilation are discussed. For comparison, a sensitivity experiment without bottom boundary layer parameterization (section 2.2) is performed.

During the whole integration period of about 23.5 years neither restart from observations nor other data assimilation techniques were utilized. The strategy of this study is to evaluate the performance of a state-of-the-art climate model during saltwater inflow events rather than to present 'perfect' model results. The model results might be improved if observations were assimilated successively during the long integration.

To analyze the ventilation of the deepwater a passive age tracer is implemented. The age a of a particle of seawater is defined as the time elapsing since the particle under consideration left the region, in which its age is prescribed to be zero (Deleersnijder et al. 2001):

$$\frac{\partial a}{\partial t} + \nabla\ (\vec{v}\ a - K\ \nabla\ a) = 1, \qquad (1)$$

where t – time, $\vec{v}$ – the velocity, and K – the diffusivity tensor. In this study, the age on May 1, 2002, the age at the open boundary in the Kattegat in case of inflow, and the age at the sea surface were relaxed to zero. For further details of the implementation of the age tracer in RCO the reader is referred to Meier (2004).

3. Results

3.1. Comparison with observations

The Landsort sea level is a good measure of the volume of the Baltic and has frequently been used to estimate the volume of inflowing water during inflow events (e.g. Matthäus & Franck 1992). We found good agreement between model results and observations (Fig. 2). Based upon 2-daily data mean error, the root mean square error, correlation coefficient, and explained variance amount to ME=0.1 cm, RMSE=4.4 cm, R=0.96, and VAR=0.92, respectively. The transports through the Sound and across the Darss Sill cannot usually be validated owing to the lack of operational transport measurements. However, a simple hydraulic model carefully calibrated with ADCP measurements (Håkansson 2004) is run operationally by SMHI. The results are accessible via the homepage of BOOS (Baltic Operational Oceanographic System, see http://www.boos.org/). In Fig. 3 the accumulated inflow through the Sound calculated with RCO is compared with data of the hydraulic model by Håkansson (2004). As the long-term trend of the hydraulic model is not reliable, only the detrended record

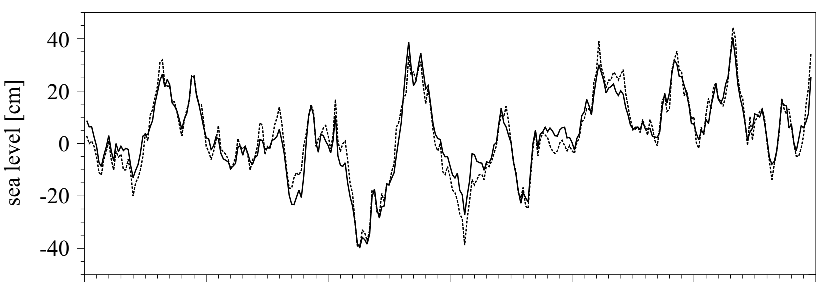

Fig. 2. 2-daily sea level at Landsort (in cm): model (solid), observations (dotted). The time axis starts on May 1, 2002

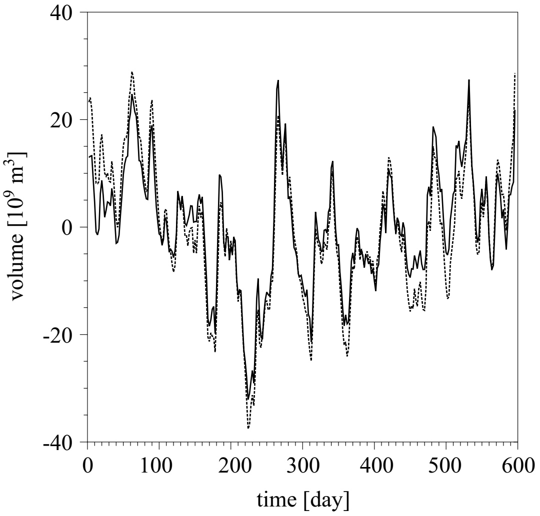

Fig. 3. 2-daily detrended accumulated inflow (in 10^9 m^3) through the Sound: RCO (solid), hydraulic model by Håkansson (2004) (dotted). The time axis starts on May 1, 2002

of the inflow period 2002/2003 is shown. The agreement between the two models is satisfactory. Based upon 2-daily data mean error, the root mean square error, correlation coefficient, and explained variance amount to ME=0.9 km^3, RMSE=4.1 km^3, R=0.95, and VAR=0.89, respectively.

The volume of the Baltic Sea is governed by the balance between river runoff, precipitation minus evaporation, and the inflows through the Sound and across the Darss Sill. For the period May 2002 to December 2003 the accumulated discharge calculated by the large-scale hydrological model amounts to 569 km^3 in 600 days. The sum of this value and the accumulated net precipitation determines the overall trend of the accumulated inflow through the Danish Straits (Fig. 4). In the climatological mean, the net precipitation is a minor part of the total freshwater inflow to the Baltic (c. 11%) (Meier & Döscher 2002). For the 600 days selected in this study we found a contribution of 14%. The total volume budget for May 2002 to December 2003 shows that 35% of the water flows through the Sound and 65% through the Great Belt and Little Belt (Fig. 4). For the period May 1992 to December 1993 (including the major inflow in January 1993) the corresponding contributions are 29 and 71%. The inflowing volumes during the precursor and inflow period of the events in January 1993 and 2003 are listed in Table 1. For the major inflow in 2003 we summed the transports between December 31, 2002 and January 22, 2003 and found a total net inflow of 222 km^3. 42% of this inflowing water has a salinity

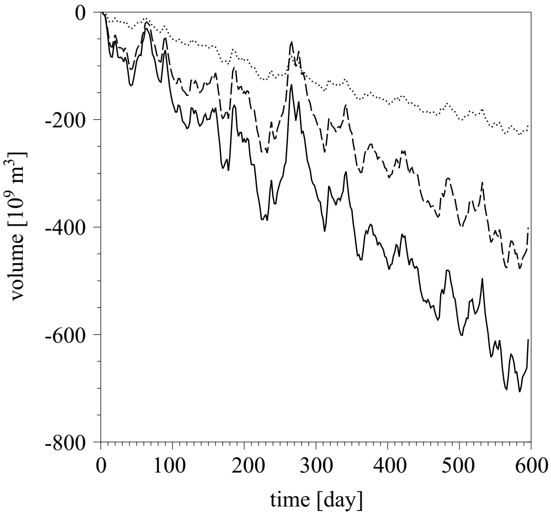

Fig. 4. Accumulated inflow (in 10^9 m^3) through the Sound (dotted), across the Darss Sill (dashed), and the sum (solid). The time axis starts on May 1, 2002

$>$ 17 PSU. The ratio of the total inflowing volumes through the Sound and across the Darss Sill was about 2:9. RCO results are close to estimates from observations (Table 1).

Table 1. Comparison of inflow volumes in January 1993 and 2003 through the Sound (S) and across the Darss Sill (DS). V_d and V are calculated during the inflow period with $S > 17$ PSU and during the precursor and inflow periods (all salinities), respectively

	Volume					
	V_d [km^3]			V [km^3]		
Reference	DS	S	Sum	DS	S	Sum
1993 Meier et al. (2003)	66	71	137	210	88	298
1993 Fischer & Matthäus (1996)	93	66	159	–	–	–
1993 (this study)	91	72	163	233	71	304
2003 Feistel et al. (2003a)	65	32	97	–	–	–
2003 (this study)	51	43	94	182	40	222

In the following, the temperature and salinity results shown in Figs 5 and 6 are compared to corresponding observations presented by Feistel et al. (2003a), Feistel et al. (2003b), and Piechura & Beszczyńska-Möller (2003). A sequence of salinity sections from the Belt Sea to the north-eastern Gotland Basin between January and July 2003 illustrates the ventilation

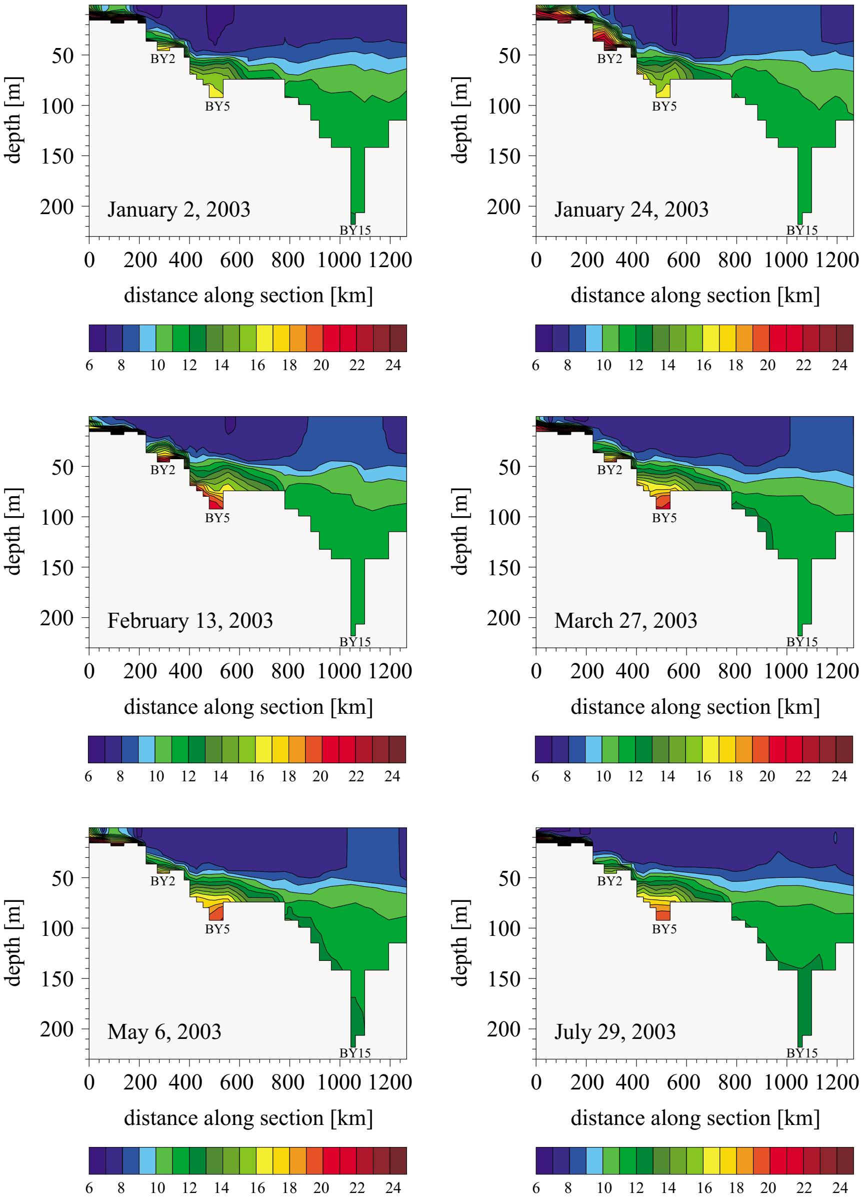

Fig. 5. Cross-section of simulated salinity (in PSU) from the Belt Sea to the northern Gotland Basin on selected dates (cf. Feistel et al. (2003a), their Figs 4 to 7, and 9). The position of the section is depicted in Fig. 1

of the Baltic deepwater as a consequence of the major inflow in January 2003 (Fig. 5). The inflow started at the Darss Sill between January 16 and 25. The first signs of the inflow arrived at the Bornholm Deep (BY5) and Gotland Deep (BY15) on January 28 and April 16, respectively (not shown).

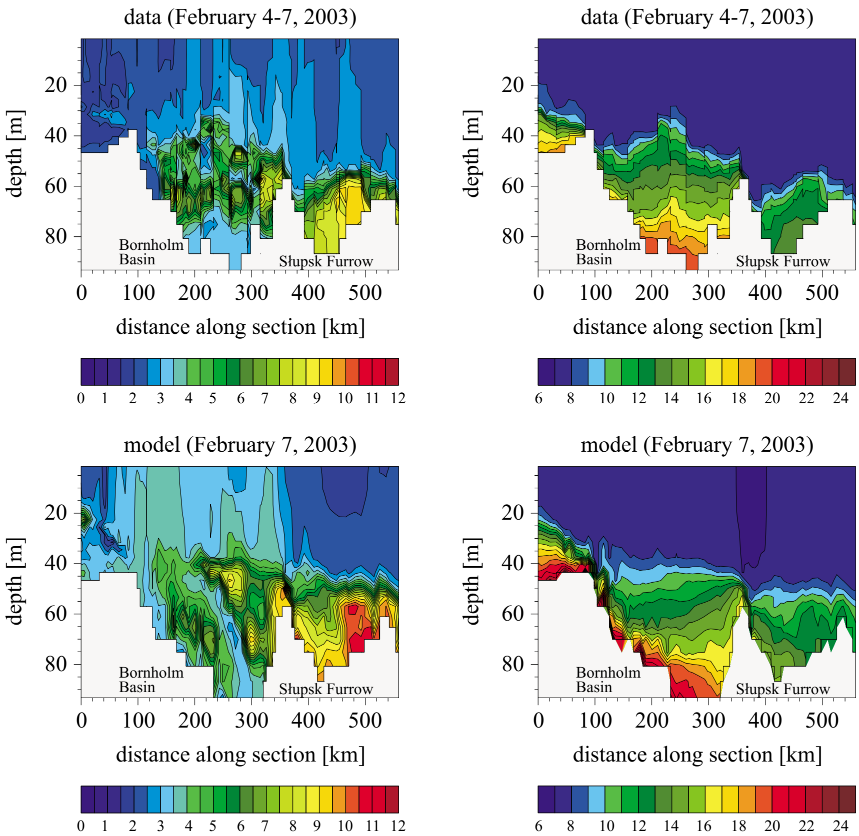

Fig. 6. Cross-sections of observed (upper panels) and simulated (lower panels) temperatures (in °C, left panels) and salinities (in PSU, right panels) from the Bornholm Channel to Słupsk Furrow on February 7, 2003 (February 4–7 for the observations; cf. Piechura & Beszczyńska-Möller (2003), their Fig. 6). The position of the section is depicted in Fig. 1. Exact agreement between observations and model results cannot be expected owing to the slightly different sample times and positions

Thus, the inflowing water arrived at the Bornholm Deep after 12 days and at the Gotland Deep after about 3 months. The remarkably rapid movement from the Darss Sill and Drogden Sill into the Baltic deepwater is confirmed by observations (Piechura & Beszczyńska-Möller 2003). In the first stage of this event the speed is estimated to be at least 30 cm s^{-1}, whereas the usual speed of inflowing water is about 3–5 cm s^{-1} (Piechura & Beszczyńska-Möller 2003). Between January 2 and January 24 the maximum salinity of the bottom water in Arkona Basin (BY2) increased from more than 17 PSU to more than 23 PSU (Fig. 5). This figure is confirmed by observations (Feistel et al. 2003a). On February 13 the simulated maximum salinity of the Bornholm Basin deepwater exceeded 20 PSU (in observations 19 PSU). On March 27 water with salinities of more than 12 PSU propagated in RCO from Słupsk Furrow (see Fig. 1) to Gotland Deep in agreement with observations. Snapshots from May 6 and July 29 show the subsequent filling of the Gotland Basin with new water.

Available observations indicate that the inflow is well simulated (Fig. 6). At the beginning of February the saline water was flowing downslope into the Bornholm Deep. Thereby the old, warm bottom water was raised and broke up into small patches concentrated at 45–55 m and 60–70 m depth. We also found this patchy layer structure in the simulation. However, in the model the salinity of the inflowing water into the Bornholm Deep and Słupsk Furrow is somewhat higher than in the observations. The measurements shown in Fig. 6 were cast between February 4 and 7, whereas the model result is a snapshot from February 7. Hence, some discrepancy is expected as the hydrography was changing daily. In Słupsk Furrow old water with relatively high temperatures from intermediate layers of the Bornholm Basin pushed the local water eastwards and upwards. Even this process is reproduced in the model.

During February wind speeds were quite low (mainly below the annual mean of 7.1 m s^{-1} for 2003, e.g. Nausch et al. (2004)). Consequently, the deepwater transport through the Słupsk Furrow was not correlated with the wind forcing (Krauss & Brügge 1991) but was mainly density driven. In the model, a large portion of the high saline water in the Arkona Basin descended from about 40 m in the Bornholm Channel down to about 60 m depth generating a strong anticyclonic circulation in the Bornholm Basin. In the well pronounced core of this current, which followed first the Swedish coast and later left the Bornholm Basin at the sill of the Słupsk Furrow, we found velocities of about 10–20 cm s^{-1}. During February the anticyclonic circulation within the halocline was relatively persistent. By contrast, observations between January 21 and 26 showed a cyclonic circulation at 50 and 60 m depth (see Piechura & Beszczyńska-Möller 2003, their Fig. 5). Further, during this period with a relatively strong easterly bottom transport in the Słupsk Furrow highly saline water in the Bornholm Basin from below sill level withdrew up to the sill and spilled over into the channel (Fig. 6). Both observations and model results show this process. We suggest that the mechanism controlling the uphill flow is based upon selective withdrawal, i.e. the transport of highly saline water through the Bornholm Channel might control the amount of withdrawn water below the sill depth at Słupsk Furrow (see discussion). For the study period 2002/2003 we found a long lasting uphill flow in the model only during February after the major inflow.

3.2. Comparison of the major inflows in 1993 and 2003

In January 1993 the volume of inflowing highly saline water amounts to 163 km^3 which is 73% more than 2003 (Table 1). Although the inflow in January 2003 is significantly weaker than that in January 1993, it has been estimated to be the 25th strongest inflow since 1897 following the criteria of Fischer & Matthäus (1996) (see Feistel et al. 2003a). The 1993 inflow is the fourth strongest. The RCO results of the inflowing highly saline water during January 1993 presented by Meier et al. (2003) are somewhat lower than in the present study, mainly because of different initial conditions and the altered bottom friction. During both events the highly saline water in the Arkona Basin is not significantly affected by the newly implemented BBL (Figs 7 and 8). As the flow in the Danish Straits is mainly barotropic during major inflows, this behavior is expected. However, in the Bornholm Basin the inflowing highly saline water is significantly increased with BBL. Actually, the impact is larger in the case of the smaller inflow in 2003 than in the case of the 1993 event. For 2003 the maximum volume of highly saline water is 43% higher with BBL than without BBL. For 1993 the increase still amounts to 20%. Comparison with observations shows that for the whole period 1980 to 2003 the results of the Bornholm Basin in the simulation with BBL are more realistic than without BBL (not shown). It is interesting to note that in the case with BBL after the first rapid increase, the volume of highly saline water further increases but at a slower rate during the following

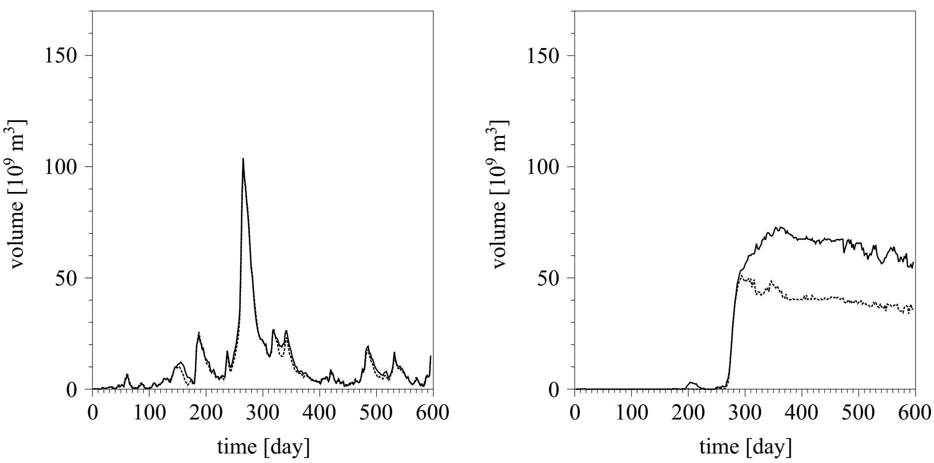

Fig. 7. Volume (in km^3) of highly saline water ($S > 17$ PSU) in the Arkona Basin (left) and the Bornholm Basin (right) during the 2003 inflow event: reference run (solid line), experiment without BBL (dotted line). The time axis starts on May 1, 2002

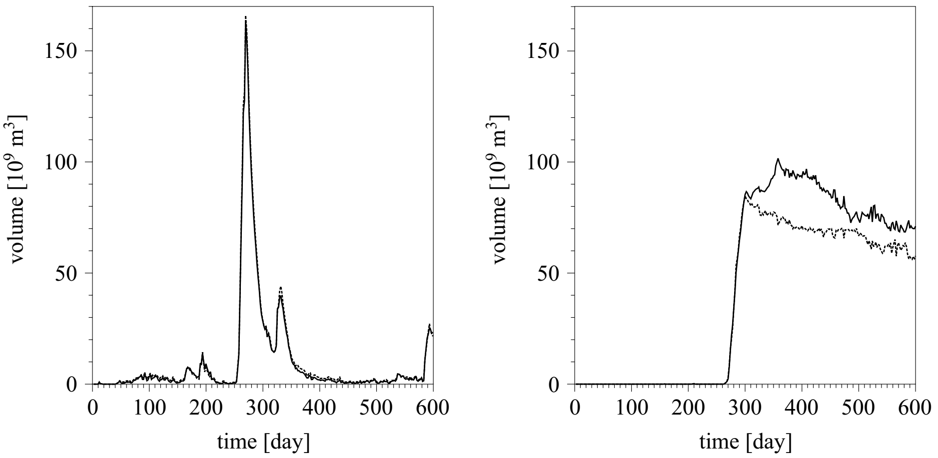

Fig. 8. As Fig. 7 but for the 1993 inflow event. The time axis starts on May 1, 1992

month, whereas it drops immediately after the first rapid increase when BBL is not used. This behavior is found during both events.

3.3. Smaller inflows in summer and autumn 2002

In the following, the time evolution of the simulated smaller inflows in summer and autumn 2002 will be discussed in comparison to observations (Figs 9 to 11).

3.3.1. Darss Sill

Observations at the Darss Sill show a salty ($S > 17$ PSU) and warm (13°C<T< 18°C) bottom layer of about 5 m thickness between August 5 and October 11 (Feistel et al. 2003b). We found such a bottom layer of elevated salinity in the model as well. However, due to the lack of mixing in the model in the Great Belt and in the Belt Sea vertical gradients are overestimated (not shown). This might be the reason why the model underestimates mean temperatures and salinities (Table 2). However, comparison with observations is difficult because at the Darss Sill mast measurements from only 3 depths (7, 17, 19 m) were available (Feistel et al. 2004).

3.3.2. Arkona Deep

In the Arkona Basin the simulated bottom salinity increases at the beginning of August but bottom salinities > 16 PSU first appear in September (Fig. 9). Larger amounts of highly saline water with maximum salinity > 17 PSU flow into the Arkona Basin on October 31 in agreement with observations. During September and October the largest signal of the

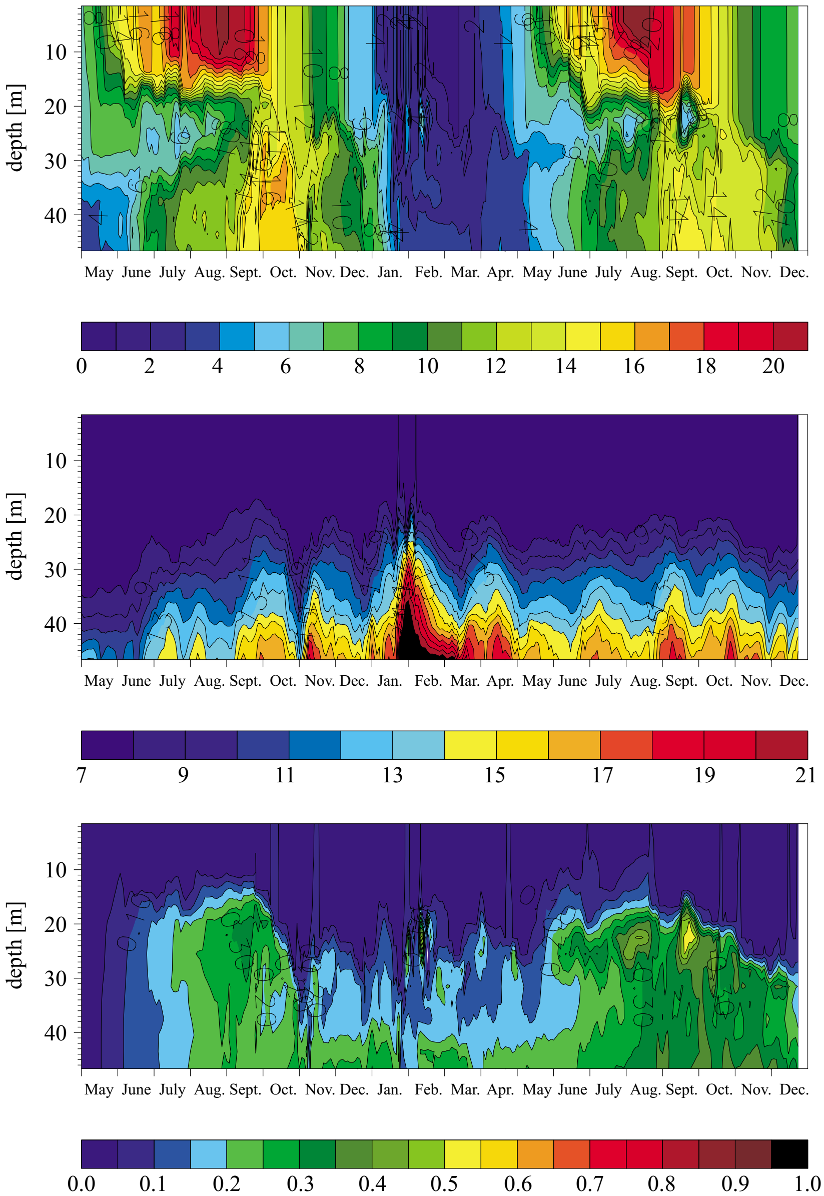

Fig. 9. Simulated temperature (in °C, upper panel), salinity (in PSU, middle panel), and age (in years, lower panel) as a function of time and depth in the Arkona Deep (BY2, see Fig. 1; cf. Feistel et al. (2003b), their Fig. 3). The time interval May 2002 to December 2003 is shown

summer inflow is found at 35 m with temperatures exceeding 17°C and with a pronounced local age minimum during mid-October. The November inflow is accompanied by a temporary increase in temperature (maximum

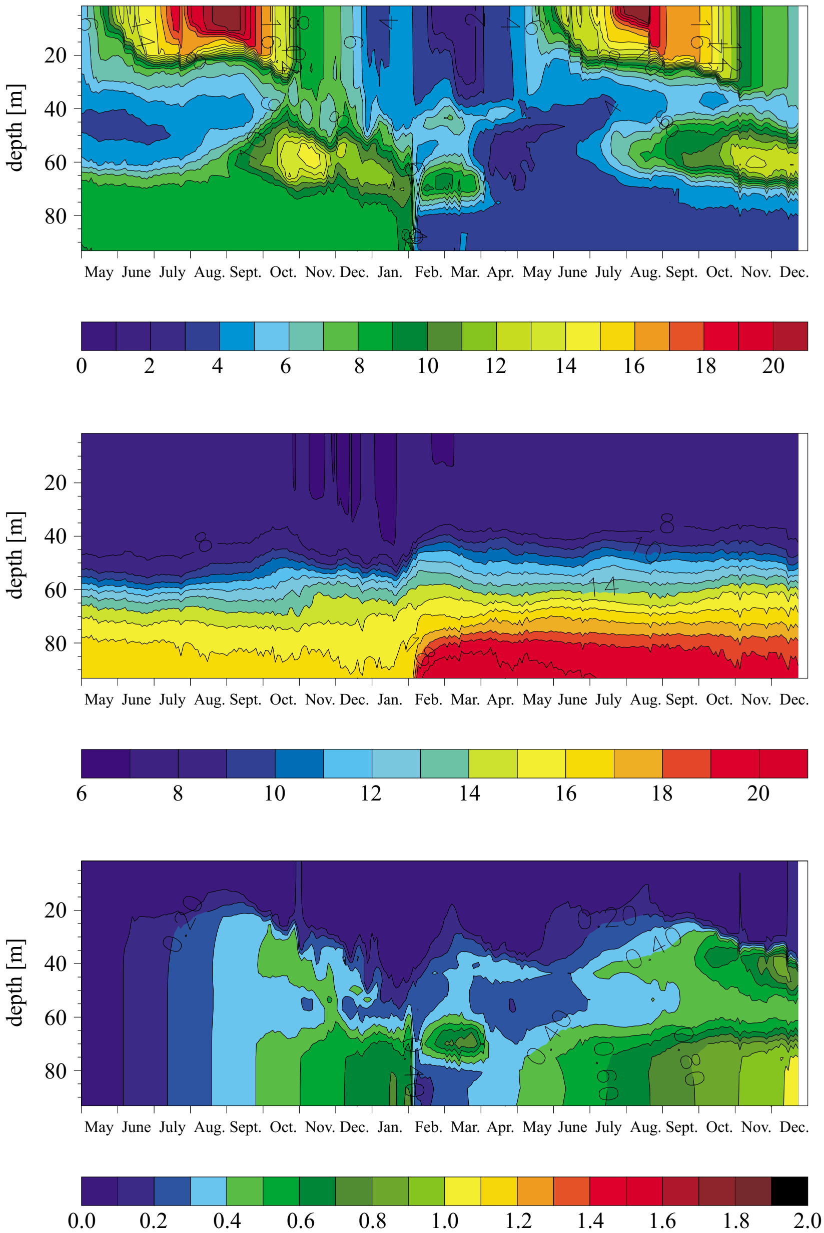

Fig. 10. As Fig. 9 but in the Bornholm Deep (BY5, see Fig. 1; cf. Feistel et al. (2003b), their Fig. 4)

about 12°C) and a decrease in age down to the bottom. In contrast to these warm inflowing water masses, the water of the major inflow in January and later the smaller inflow in March has a temperature of only about

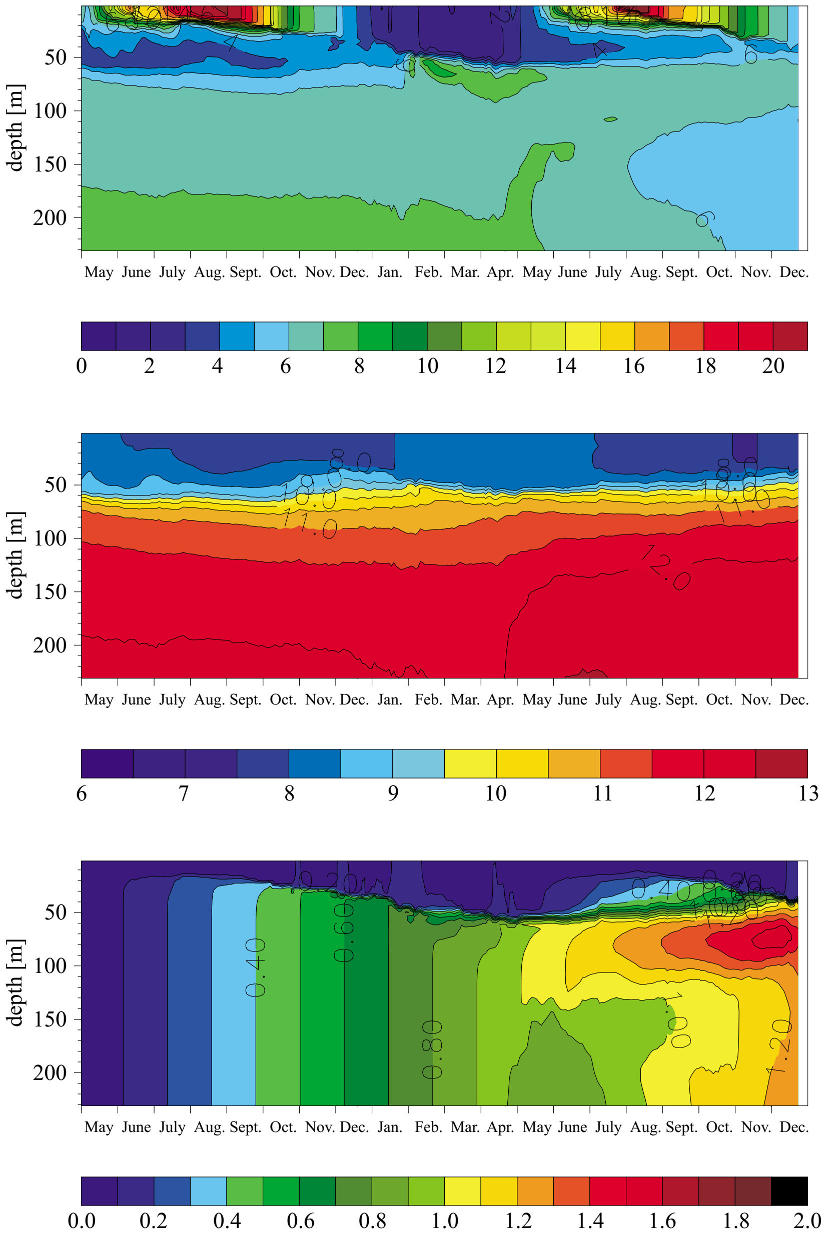

Fig. 11. As Fig. 9 but in the Gotland Deep (BY15, see Fig. 1; cf. Feistel et al. (2003b), their Fig. 6)

3–4°C (observations: 1–2°C (Piechura & Beszczyńska-Möller 2003)). A new summer inflow also occurs in September 2003 with temperatures > 14°C and with salinities > 17 PSU.

Table 2. Periods of warm water inflow across the Darss Sill and their mean temperature and salinity characteristics in observations (Feistel et al. 2004, their Table 4.2.1) and in RCO

Period 2002	Mean temperature [°C]		Mean salinity [PSU]	
	Data	Model	Data	Model
Aug. 5–18	14.7	12.2	16.9	15.2
Aug. 21–Sept. 20	15.9	13.5	17.7	16.8
Sept. 21–29	16.7	14.6	17.1	17.1
Oct. 6–11	15.3	14.4	17.8	15.8

3.3.3. Bornholm Deep

In the central Bornholm Basin the summer 2002 inflow appears within the halocline at about 60 m depth between September and November with maximum temperatures exceeding 14°C (Fig. 10). This is in agreement with observations (Feistel et al. 2003b). A corresponding local age minimum is found on November 1 at the same depth. Although the simulated inflow in November has the correct temperature characteristic (maximum temperature exceeding 11°C), its salinity (and respective density) is too low. Thus, the corresponding temperature and age signals are centered during December at about 60 m depth. By contrast, observations show well-pronounced temperature and oxygen maxima at 90 m depth (Feistel et al. 2003b). In the model, either the November inflow into the Arkona Basin is too weak or the water is diluted too much on its way from the Arkona Basin into the central Bornholm Basin. On January 28 the water mass of the major inflow with temperatures of 3–4°C and with maximum salinity exceeding 20 PSU replaces the old bottom water below 75 m. The old bottom water is thereby lifted above the western sill of the Słupsk Furrow, which has a depth of about 60 m. Some of the older water remains at 60 to 75 m depth until the end of March.

3.3.4. Gotland Deep

Around January 21, 2003 the water of the summer inflow 2002 is sandwiched into the halocline of the Gotland Deep at 50–80 m depth (Fig. 11). This is somewhat higher in the water column than in the observations (between c. 80–130 m, see Feistel et al. 2003b) mainly because the halocline changes in the model during the 24-year long integration (cf. Meier & Kauker 2003). This water is characterized by temperatures still larger than 7°C, which is in agreement with observations. In observations of oxygen (or hydrogen sulphide) the first splash of the summer inflow appeared in the bottom layer already at the beginning of December 2002, reducing the hydrogen sulphide concentration there (Feistel et al. 2003b). The age tracer did not provide any evidence in our model simulation for the arrival of the water of the summer inflow before mid-January 2003. The smaller inflow from the beginning of November 2002 also appears at the depth of the halocline during March 2003 spreading the temperature maximum in time, whereas oxygen observations indicate that this inflow even ventilates the bottom layer. As the simulated salinities of the November inflow in Bornholm Basin were already too low (section 3.3.3), we cannot expect a ventilation of the bottom layer in the Gotland Basin. The subsequent, much more intense inflow from January 2003 ventilates the whole water column below 130 m after April 16, 2003, which is in agreement with observations (Feistel et al. 2003b). The deepwater in the Gotland Basin, which was unusually warm since the inflow in September 1997 (Nausch et al. 2003), is cooled down to less than 6°C by the inflows from January and March 2003. The latter arrived at Gotland Deep in August 2003 and ventilated the deepwater in RCO mainly between 130 and 200 m depth.

4. Discussion

RCO is a regional climate model which is frequently utilized in multi-year simulations to study past (e.g. Meier & Kauker 2003) and future climate (e.g. Meier 2002a, b). Therefore, it is important that saltwater inflows be well simulated. In this study we have shown that the deepwater ventilation in the case of larger inflows worked well. However, artificial entrainment of the saltwater flow between the Bornholm Basin through the Słupsk Furrow into the eastern Gotland Basin might be the reason that interleaving at 100 m in the Gotland Deep could have been overestimated, whereas ventilation of the bottom water is underestimated. A small bias of each simulated saltwater inflow might have caused the artificial change in the halocline depth and in the surface salinity in the eastern Gotland Basin in multi-decade-long simulations. However, further investigations, inter alia in the form of process studies, are still necessary to elucidate the mechanisms involved, because it might be necessary to improve not only the mixing parameterization but also the atmospheric forcing and the horizontal resolution. Therefore, we plan to use the atmospheric forcing of the mesoscale analysis project MESAN (Häggmark et al. 2000) with a horizontal resolution of 10 km in future. In addition, RCO will be used with a horizontal resolution of 1 nautical mile. Although such a resolution might still be only eddy-permitting, we hope to find that the simulated mesoscale variability, e.g. in the Słupsk Furrow, will become more realistic. Observations in the Słupsk Furrow show that the transport of the highly saline water through the channel is very variable (Piechura & Beszczyńska-Möller 2003). After saltwater inflows, intensive mesoscale cyclonic eddies carrying the inflow water were generated just east of the Słupsk Furrow and in the permanent halocline along the saline intrusion pathway into the Gotland Deep (Zhurbas et al. 2003). Zhurbas et al. (2003) showed that this mesoscale variability is realistically simulated using a horizontal resolution of 0.5–1 nautical mile.

As the summer 2002 inflow was apparently not driven by westerly gales, there have been speculations about the mechanism driving them (Feistel et al. 2003b). We have shown that the volume budget of our model is quite realistic (cf. Figs 2 and 3). The volume balance in Fig. 4 indicates a net inflow into the Baltic of about 50 km^3 between August 28 and October 8. This coincides with the main inflow period of highly saline water in the model simulation. We found an outflow through the Sound but a significant inflow across the Darss Sill. Thus, barotropic forcing by sea level differences caused by remote atmospheric forcing as a driving mechanism cannot be excluded. Very likely, calm weather is a precondition for the inflow event with increased salinities in the bottom layer of the Belt Sea caused by baroclinic estuarian inflow.

Another interesting question is what causes the uphill flow of dense bottom water from the eastern Bornholm Basin towards the top of the western sill of the Słupsk Furrow (Fig. 6). We suggest that selective withdrawal is the dominant process. Flow into a sink in a stratified medium leads, at small Froude numbers, to the selective withdrawal of fluid from a well-defined layer centered around the level of the sink (e.g. Imberger 1980). According to Stigebrandt (1978b) the thickness of the flowing layer is in general controlled by the discharge, the geometrical dimensions of the outlet, and the upstream stratification. The approach was successfully applied to explain the dynamics of an ice-covered lake with through-flow

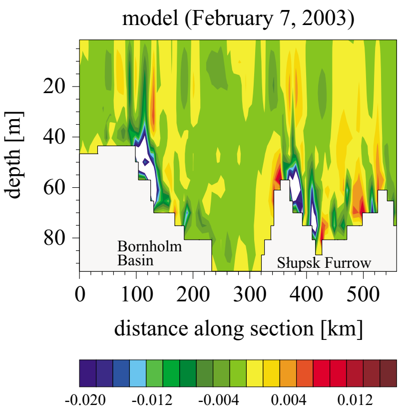

Fig. 12. Cross-section of simulated vertical velocity (in cm s^{-1}) from the Bornholm Channel to the Słupsk Furrow on February 7, 2003. The position of the section is depicted in Fig. 1. Velocities smaller than -0.02 cm s^{-1} are shown in white

(Stigebrandt 1978a). During February we found in our results significant upward velocities at the eastern slope of the Bornholm Basin between about 75 m and the depth of the sill (Fig. 12). Our hypothesis is supported by the fact that in the model neither rotation (of a cyclonic current following the 60 m isobath) nor the impact of wind (Krauss & Brügge 1991) seem to be important during February. However, as in our case, the approach by Stigebrandt (1978a) is not applicable, and further investigations are still necessary.

5. Conclusions

The main conclusions are as follows:

(1) During inflow periods the overall performance of the RCO climate model is good. Simulated volumes of inflowing high saline water are close to the figures estimated from observations. In principle, the ventilation of the deepwater in the Bornholm and Gotland Basin is in agreement with observations. Even smaller inflows like the ones in August/September and November 2002 are treated satisfactorily. Traces of these inflows were identified even in the halocline of the western Gotland Basin.

(2) The implementation of the diffusive part of the BBL improved the ventilation of the Bornholm deepwater significantly.

(3) RCO is a useful tool for studying climate variability on long time scales. With the aid of model results, reliable volume and salt budgets of saltwater inflows have been calculated. As measurements are sparse in time and space, budget calculations of saltwater inflows and analysis of processes are important applications of model results.

(4) Selective withdrawal is a possible explanation for the frequently observed process where dense water from below sill level rises up to the sill and spills over into Słupsk Furrow.

Appendix: Description of the BBL

5.1. General remarks

Beckmann & Döscher (1997) coupled a terrain-following bottom boundary layer to a standard OGCM. Bottom boxes communicate by advection and diffusion as indicated in Fig. 13.

Bottom boxes are altered according to

$$\rho_t = \rho_t^{adv} + \rho_t^{diff} \tag{2}$$

with the potential density ρ and its time tendency ρ_t. The advective velocities are inferred from the OGCM. This approach combines some of

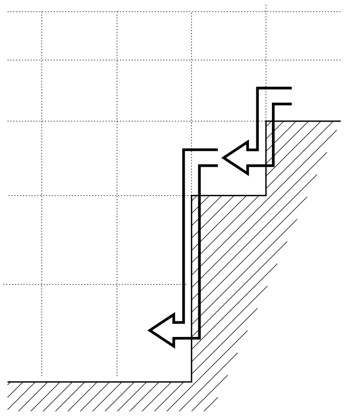

Fig. 13. BBL connecting bottom boxes of the OGCM, enabling advective or diffusive transports along the bottom. (Figure taken from Döscher & Beckmann 2000)

the benefits of a σ-coordinate model (bottom following) with the simplicity of a geopotential coordinate model. A separate BBL momentum equation is not used in this approach.

5.2. BBL advection

The original scheme by Beckmann & Döscher (1997) has been modified by a conditional advection (see Döscher & Beckmann 2000), i.e. bottom-following advection is allowed only if dense water overlies less dense water on the slope (i.e. $\nabla\rho\nabla h < 0$ with depth h) and if the velocity u is directed towards greater depth ($u\nabla h > 0$, 'dense downward condition'). Under these conditions, the cross–isobath bottom layer flow of the level model is rotated completely into the terrain–following system:

$$\begin{aligned}\rho_t^{adv} = &-\frac{1}{h}[(\alpha^x h u_{BBL}\rho)_x + (\alpha^y h v_{BBL}\rho)_y + (\alpha^z h \Omega_I \rho)_\sigma]^\sigma \\ &-[((1-\alpha^x)u\rho)_x + ((1-\alpha^y)v\rho)_y + ((1-\alpha^z)w\rho)_z]^z\end{aligned} \tag{3}$$

(Ω_I = vertical velocity in the σ coordinate system of the bottom layer, u,v, and w = zonal, meridional and vertical velocities). Tracer fluxes are calculated in the topography-following coordinate only if the dense downward condition is fulfilled. Formally, this can be expressed as:

$$\begin{pmatrix}\alpha^x \\ \alpha^y\end{pmatrix} = \begin{cases}1 & \text{if} \quad \nabla\rho\nabla h < 0, u\nabla h > 0 \\ 0 & \text{else}\end{cases}, \tag{4}$$

$$\alpha^z = \max(\alpha^x, \alpha^y). \tag{5}$$

5.3. BBL diffusion

The bottom-following diffusion equation of the BBL

$$\rho_t^{diff} = [(A^\sigma \rho_x^{bottom})_x + (A^\sigma \rho_y^{bottom})_y] \tag{6}$$

is constrained by the Laplacian diffusion coefficient A^σ:

$$A^\sigma(x,y,t) = \begin{cases} A^\sigma_{\max} & \text{if} \quad \nabla\rho^{bottom}\nabla h < 0 \\ A^\sigma_{\min} & \text{else} \end{cases} \tag{7}$$

(ρ^{bottom} symbolizes potential density in the bottom tracer box). The BBL diffusion affects the tracer fields in addition to the diffusion of the GCM. A typical BBL diffusivity for the Baltic Sea application is 2×10^8 cm^2 s^{-1}. A small minimum diffusivity (10^4 cm^2 s^{-1}) is necessary to prevent numerical instabilities.

5.4. Features of the Baltic Sea configuration

Tests of the BBL for a Baltic Sea configuration revealed the advection to be inefficient. Vertical columns receiving advective inflow at their base from the BBL, are homogenized rapidly owing to a necessary compensating upward flow. The time scale τ of that vertical flow is approximated by

$$\tau = \frac{N\Delta x}{U}, \tag{8}$$

with N the number of vertical boxes per step, Δx the model grid size and U the bottom velocity. The homogenization time rises for increasing horizontal resolution. For 2 nautical miles (as in our study) and typical velocities (e.g. 0.1 m s^{-1}), τ is just a few hours to days. Thus, the simplifications in the concept of an advective BBL component are not appropriate to keep a thin bottom layer in our high-resolution setup of RCO. Consequently, the BBL advection is disabled in this study. Instead, the effect of the BBL arises from the diffusive part.

In general, the effect of the BBL is very much dependent on the interplay with other mixing algorithms in the model. In some situations, the pure addition of a BBL scheme to the current RCO version with a k-mixing scheme in combination with deepwater mixing due to breaking internal waves does not add much, either because the existing schemes do part of the job of the BBL scheme, or because the vertical mixing destroys additional BBL effects during certain events. However, overall, the BBL leads to significantly improved salinities in the deepwater of the Bornholm Basin.

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