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Consequences of artificial deepwater ventilation in the Bornholm Basin for oxygen conditions, cod reproduction and benthic biomass – a model study

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Abstract

We develop and use a circulation model to estimate hydrographical and ecological changes in the isolated basin water of the Bornholm Basin. By pumping well oxygenated so-called winter water, residing beneath the level of the summer thermocline, to the greatest depth of the basin, where it is forced to mix with the resident water, the rate of density reduction should increase and thereby the frequency of intrusions of new oxygen-rich deepwater. We show that pumping $1000 \, \text{m}^3 \, \text{s}^{-1}$ should increase the rates of water exchange and oxygen supply by 2.5 and 3 times, respectively. The CRV (Cod Reproduction Volume), the volume of water in the isolated basin meeting the requirements for successful cod reproduction ($S > 11$, $O_2 > 2 \, \text{mL} \, \text{L}^{-1}$), should every year be greater than $54 \, \text{km}^3$, which is an immense improvement since it in certain years is currently much less. Anoxic bottoms should no longer occur in the basin and hypoxic events will become rare. This should permit extensive colonization of fauna on the earlier periodically anoxic bottoms. Increased biomass of benthic fauna should also mean increased food supply to economically valuable demersal fish like cod and flatfish. In addition, the bioturbation activity and re-oxygenation of the sediments should lead to increased phosphorus retention by the sediments.

1 Introduction

The Baltic Sea is the second largest ($373\,000 \, \text{km}^2$) brackish water system in the world with an extensive drainage system that is four times larger than the sea surface area. The salinity varies from about 1 in the northern Bothnian Sea to about 17 in the deep basins of the southwestern Baltic. The Baltic is enclosed, with restricted water exchange with the adjacent seas; the Kattegat, the Skagerrak and the North Sea. Because of the restricted water exchange, an increased input of nutrients since the mid-1900s, and a halocline at 60 to 80 m depth, large volumes of bottom water below the halocline lack oxygen concentrations high enough to support the respiration of

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aquatic organisms. During the last decade, areas of approximately $50\,000\,km^2$ (year 2011) of the seabed have been anoxic and $27\,000\,km^2$ hypoxic ($< 2\,mL\,O_2\,L^{-1}$); thus lacking benthic animals (Karlson et al., 2002). The Baltic is the largest sea area with anthropogenic-related low oxygen concentrations in the world (Diaz and Rosenberg, 2008) because of increased input of organic matter (Bonsdorff et al., 1997) that has changed the ecosystem properties (Elmgren, 1989; Rosenberg et al., 1990). The increased input of organic matter has been boosted by extensive leakage of phosphorus from anoxic bottoms that has increased the phosphorus content in the water column in spite of halved external supply of phosphorus since the 1980s (Stigebrandt et al., 2013). As a consequence, the structure and function of the Baltic ecosystem is affected by two main factors – horizontally by increasing salinity from north to south and vertically by decreasing oxygen concentrations by depth.

The annual fish catches increased in the Baltic in the middle of the last century and the highest landings of cod, the most economically valuable species, peaked around 1990 with about $400\,000$ tons (MacKenzie et al., 2000). Cod is dependent on the deeper bottoms of the Baltic for reproduction and feeding, whereas herring and sprat are pelagic feeders. The cod stocks have declined after 1990 because of too high fishing pressure and unfavorable spawning conditions. Other factors with negative impact on the cod are that herring and sprat are predators on egg and larvae, which deters the recovery of the cod stock.

The anoxic and hypoxic water in the deeper parts of the Baltic has great negative consequences for the ecosystem. Few or no animals are present at depths deeper than $70\,m$, which means that there are vast volumes that are lacking fish and fish food. The reproduction of cod is hampered as the eggs need an oxygen concentration $> 2\,mL\,L^{-1}$ and a salinity of at least 11 to be buoyant but this varies somewhat with egg size and age of the spawning cod (Hinrichsen et al., 2007; Vallin et al., 1999). Such a high salinity is found in the Bornholm Basin and, east of this basin, only close to the bottom and then frequently associated with toxic hydrogen sulfide, which is produced during anoxia. MacKenzie et al. (2002) summarized the volumes of water in the main cod

spawning areas of the eastern Baltic and Bornholm Basins where salinity and oxygen conditions could allow successful fertilization and hatching. From the mid-1980s, this was almost non-existent in the Gdansk and Gotland Basins and favorable conditions were only found in the Bornholm Basin.

5   Defaunated sediments appear laminated, which demonstrates that the biogeochemical processes that occur in oxidized sediments do not occur or are significantly changed (Jonsson et al., 1990). Thus, bioturbation (mixing of sediments) and irrigation (pumping of oxygenated water into the sediment) are lacking in laminated sediments. Lack of benthic animals means reduced nitrification and reduced capacity to bind phosphorus

10 in the top sediments, which may accelerate eutrophication and occasional blooms of cyanobacteria.

     It has been suggested that the eutrophication of the Baltic proper may be reduced by man-made oxygenation of the deepwater, which should decrease the phosphorus content in the water column, and that oxygenation should be done by pumping down

15 well-oxygenated so-called winter water into the deepwater (Stigebrandt and Gustafsson, 2007). Gustafsson et al. (2008) and Conley et al. (2009b) discuss engineering methods that have been suggested to cure hypoxia in the Baltic Sea. Both papers consider below-halocline ventilation, the method applied in the present paper, to be the only realistic alternative.

20   Before considering oxygenation of the whole Baltic proper it may be wise to first look into oxygenation of a smaller part of the Baltic. In this paper we have chosen to consider oxygenation of the deepwater of the Bornholm Basin, beneath the level of the greatest depth (60 m) of Stolpe Channel (Slupsk Furrow) in the southwestern Baltic proper (Fig. 1). Very salty deepwater may stay in this basin for several years whereby

25 hypoxia and anoxia may develop in the lower parts.

     Pumping winter water into the deepwater creates a mixture of winter water and deepwater that is buoyant and rises as an entraining plume until reaching the level where the buoyancy is lost and interleaving occurs. Removal of deepwater by entraining plumes causes compensatory downward vertical motion to replace the removed

water. Pumping thus drives a vertical circulation cell between the levels of interleaving and pump outlet, respectively, with rising motion in the (narrow) buoyant plumes located at the pumps, and sinking motion everywhere outside the plumes. The sinking motion tends to reduce vertical concentration gradients of oxygen and other dissolved

5 substances in the cell.

Stigebrandt and Kalén (2012) implemented pumping in a model of the Bornholm Basin. They showed that pumping $800\,\mathrm{m^3\,s^{-1}}$ of winter water from 40 m to 90 m with a tenfold increase of the flow due to mixing at the outlet, would have prevented the development of hypoxia and anoxia that occurred during a selected period starting in

10 early May 2003 after a complete water exchange. The characteristics of the modeled pumped flow fit well with observations made in a comprehensive, long-term pilot experiment in the By Fjord where $2\,\mathrm{m^3\,s^{-1}}$ of surface water were pumped into the usually anoxic deepwater (Stigebrandt et al., 2014). The pumped water had the speed $2\,\mathrm{m\,s^{-1}}$ through the outlet orifices. The total flux of deepwater carried by the plumes at a cer-

15 tain depth can readily be estimated from the speed of vertical migration of isohaline surfaces multiplied with the horizontal surface area of the fjord at that depth. In the By Fjord, the typical downward speed of isohaline surfaces was $2\,\mathrm{m\,day^{-1}}$, implying an entrained flow of deepwater that was thirty times greater than the pumped flow. Preliminary estimates suggest a tenfold increase due to initial mixing and a further increase

20 by a factor of three due to entrainment in the buoyant plume phase. Furthermore, the pumping in the By Fjord increased the rate of density reduction in the basin water by a factor of 100 and the frequency of water exchanges in the deeper parts of the basin by a factor of 10 (Stigebrandt et al., 2014).

In the present paper we complement the model in Stigebrandt and Kalén (2012) with

25 a time-dependent model of the inflow of new deepwater into the basin, which allows us to estimate how the rate of water exchange and the oxygen conditions in the deeper parts of the basin would be changed by pumping.

The goal of the present paper is to investigate hydrographical and ecological effects of enhanced deepwater circulation by pumping in the Bornholm Basin. First the pump is

turned off (reference run) and the high quality of model predictions is demonstrated by comparisons between model results and hydrographic observations of salinity and oxygen in the deepwater of the Bornholm Basin for the period 1990–2010 (21 years). Then the pump is turned on to simulate how the vertical circulation of the Bornholm Basin

5 should be affected with regard to residence time and changed salinity, temperature and oxygen conditions. Finally we compute how the changed hydrographical and oxygen conditions should affect the water volume available for successful cod reproduction and describe the expected colonization of the oxygenated bottoms. We estimate the possible increase of benthic biomass and discuss the importance of this as fish feed and

10 for biogeochemical sediment processes. We also briefly discuss possible effects on the Baltic Sea east of Bornholm Basin due to changed properties of new deep water entering from Bornholm Basin.

2   Methods

2.1   The circulation model

15 We will use the vertical advection–diffusion model with high vertical resolution described for the Bornholm Basin by Stigebrandt and Kalén (2012). However, that model lacks a model of the inflow of new deepwater. Here we develop and use a relatively simple mechanistic model of the flow of new deepwater from Kattegat to the Bornholm Basin (for area map, see Fig. 1). Since the model is used only in the hind-cast mode, it

20 may be tuned if necessary to improve the description of the inflow of new deepwater.

A starting point for the construction of our inflow model is that the salinity of the Baltic Sea is controlled essentially by two external factors, namely (i) the freshwater supply to the Baltic Sea and (ii) the sea level variability in Kattegat which drives the water exchange between Kattegat and the Baltic Sea inside the Fehmarn Belts and the

25 Öresund (the Sound) (Stigebrandt, 1983). Freshwater from the Baltic Sea accumulates in the surface layer of the entrance area (Kattegat and the Belt Sea). Due to shallow

sills, the presence of this layer partly blocks inflow of seawater from the deeper part of Kattegat to the Baltic, which strongly influences the salinity of new deepwater flowing into the Baltic. However, freshwater from the Baltic Sea is not only re-circulated back to the Baltic with the new deepwater, it is also permanently lost to Skagerrak by a flow that is driven by the excess pressure caused by the accumulated freshwater in the surface layer of Kattegat. Of course, the long-term freshwater export to Skagerrak equals the long-term supply to the whole Baltic Sea including the entrance area.

The daily water exchange between the Baltic Sea and the entrance area is computed from observed changes of the Baltic Sea volume (sea level) and accounting for freshwater supply inside the entrance area (Sect. 2.1.1). The amount of freshwater in the entrance area, computed by a simple budget model (Sect. 2.1.2), determines the salinity of the new deepwater flowing into the Baltic Sea, which is computed using an empirical function by Stigebrandt and Gustafsson (2003) (Sect. 2.1.3).

The inflowing new deepwater moves along the bottom of the Arkona Basin as a dense bottom current that entrains water from above, whereby the volume flow increases and the salinity decreases. The dense bottom current was observed, described and analyzed by Arneborg et al. (2007). This current replenishes the pool of new deepwater in the Arkona Basin. The flow and dynamics of new deepwater in Arkona Basin are described in Sect. 2.1.4. A current transports water from the deepwater pool towards the Bornholm Basin (Liljebladh and Stigebrandt, 1996, 1987a). After having entered the Bornholm Basin, the current of new deepwater from Arkona Basin transforms to a dense bottom current, which entrains water from above. When the new deepwater in the bottom current has got the same density as the ambient water it is interleaved in Bornholm Basin. Only the densest new deepwater makes it all the way through the halocline and down to the greatest depths, less dense new deepwater is interleaved in the halocline. Water exits the Bornholm Basin through the 59 m deep Stolpe Channel. For a description of the dynamics of the Bornholm Basin, see Sect. 2.1.5.

2.1.1 The flow through the entrance straits

The water exchange through the Öresund and the Fehmarn Belt changes the volume of the Baltic Sea and may thus be quantified from observations of the changing sea level. However, also freshwater supply changes the volume. From conservation of volume, one obtains the following equation for the volume change of the Baltic Sea

$$A_{BS} \frac{dh}{dt} = Q_{BS} + Q_{fBS} \tag{1}$$

Here dh/dt is the rate of change of the horizontal average sea level h, A_{BS} the horizontal surface area of the Baltic Sea, Q_{fBS} the freshwater supply to the Baltic Sea (inside the entrance area) and Q_{BS} the water exchange of the Baltic Sea with the entrance area. For use in the model, Q_{BS} is separated into two parts, Q_{inBS} and Q_{outBS}, describing inflow and outflow, respectively. The salinity and other properties of these flows are specified in Sect. 2.1.3.

The simple description of the inflowing water is complemented with a low-pass filter removing inflows resulting from high frequency fluctuations. This is because the first water flowing towards the Baltic proper after a period of outflow is just surface water from the Baltic. This can be handled by removing certain volumes of the first coming transport, both outgoing and ingoing, using a buffer volume. Such buffer volumes were introduced by Stigebrandt (1983) and discussed by e.g. Stigebrandt and Gustafsson (2003). The presence of a buffer volume delays the onset of flow events and reduces the importance of high-frequency fluctuations. An appropriate value of the buffer volume will be determined by tuning in Sect. 3.2.

An occasional shift in flow, which is too small for the buffer volume to be completely filled (emptied), will only pause the flow event. If the pause is too long, a new flow event starts when the buffer volume is restored. This buffer restoration time limit will be tuned so that inflow event lengths match the observed inflows, see Sect. 3.2. Finally, for each flow event, an average flow rate is calculated from the cumulated sum of water flow during the event.

2.1.2 The freshwater pool in the entrance area

To compute the salinity S_{in} of the inflowing new deepwater, we need to know the mixture of freshwater and seawater. The latter is the deepwater in Kattegat, which is modified Skagerrak water. This is computed from the amount of freshwater in the entrance area, as shown in Sect. 2.1.3 below. Therefore, we construct a model of the freshwater pool in the entrance area. The thickness of the freshwater pool, H_{fK}, is controlled by Q_{BS} and the salinity of this flow, the freshwater supply to the entrance area, Q_{fK}, and freshwater outflow from the entrance area to Skagerrak, Q_{fS}.

The rate of change of freshwater thickness, dH_{fK}/dt, in the entrance area can be calculated from the following continuity equation for freshwater

$$A_K \frac{dH_{fK}}{dt} = Q_{fK} + Q_{outBS} \frac{S_0 - S_B}{S_0} - Q_{inBS} \frac{S_0 - S_{in}}{S_0} - Q_{fS} \tag{2}$$

Here A_K is the horizontal area of the freshwater pool, S_B the salinity of the water flowing out from the Baltic Sea and S_0 the salinity of the seawater in Kattegat. The exact value of A_K is determined by model tuning in Sect. 3.2.

The freshwater outflow from the entrance area to Skagerrak can be described as in Stigebrandt and Gustafsson (2003) by

$$Q_{fS} = \frac{g \beta S_0 H_{fk}^2}{2f} \tag{3}$$

Here g is the acceleration of gravity, f the Coriolis parameter and $\beta \approx 8 \times 10^{-4}$ the salt contraction coefficient, defined by the following approximate equation of state of seawater

$$\rho = \rho_f (1 + \beta S) \tag{4}$$

where ρ_f is the density of freshwater and S is the salinity. This equation is applicable when density variations are essentially due to salinity variations, which is the case in the entrance area.

2.1.3 Properties of new deepwater from the entrance area

The new deepwater flowing into the Baltic Sea is a mixture of freshwater and deepwater from the entrance area. Its salinity S_{in} can be calculated from the following empirical equation (Stigebrandt and Gustafsson, 2003)

$$S_{in} = S_0 \frac{H_K - H_{fK}}{H_K} \tag{5}$$

The thickness H_K is an empirical constant, which is tuned in order to achieve conservation of salt in the Baltic Sea system, see Sect. 3.2. Equation (5) is valid as long as $S_{in} > S_B$ because the inflowing new deepwater must be saltier than the surface water in the Arkona Basin.

Other state variables of the new deepwater, e.g. oxygen concentration and temperature here represented by the dummy c_{in}, are calculated according to

$$c_{in} = c_{surf} \frac{S_0 - S_{in}}{S_0} + c_0 \frac{S_{in}}{S_0} \tag{6}$$

Here, c_0 and c_{surf} are approximated by monthly means of the seawater and surface values in the entrance area as described in Sect. 3.1.

In the model we use the event-average of Q_{inBS} and Q_{outBS} for each time step during a flow event, as described in Sect. 2.1.1. However, the salinity, temperature and oxygen concentration of inflowing new deepwater, computed using Eqs. (5) and (6), change for each time step.

2.1.4 The deepwater pool in Arkona Basin

Because new deepwater from the entrance area is denser than the surface water in the Arkona Basin, it descends along the bottom as a dense bottom current. By mixing with overlying water, the volume flow increases and the salinity (density) decreases.

The entrainment of ambient water into the dense bottom current is computed using the plume model in Stigebrandt and Kalén (2012), but with a reversed, i.e. downward, travelling direction and with a value of the empirical entrainment coefficient, EA, that will be tuned, see Sect. 3.2, in order to achieve a good fit with observations. The properties of the entrained water are described by monthly averages of observations from 15 m depth in the Arkona Basin, see Sect. 3.1.

After entrainment, the dense bottom current has the flow rate Q_{inA} and salinity S_{inA} when it reaches the halocline in the Arkona Basin. Below this is a thoroughly mixed pool of dense deepwater that is replenished by the inflowing new deepwater. The pool loses water by the geostrophic flow of deepwater out from the Arkona Basin. The thickness of the Arkona pool, H_A, changes according to the following equation

$$\frac{d[A_A \times H_A]}{dt} = Q_{inA} - Q_{outA} \tag{7}$$

Here A_A is the depth dependent area of the pool and Q_{outA} is the baroclinic geostrophic outflow from the pool which can be computed as in Stigebrandt (1987a), thus

$$Q_{outA} = \frac{g' H_A{}^2}{2f} \tag{8}$$

In this equation, $g' = g\Delta\rho/\rho_f$ where $\Delta\rho$ is the density difference between the pool water, salinity S_A, and the water overlying it, salinity S_B. Using Eq. (4) gives $\Delta\rho/\rho_f = \beta(S_A - S_B)$. The salinity S_A of the pool water changes according to the following equation

$$\frac{d[S_A \times A_A \times H_A]}{dt} = Q_{inA}S_{inA} - Q_{outA}S_A \tag{9}$$

Temperature and oxygen concentrations are computed analogously. Q_{outA} may empty the pool within a few weeks. This allows us to assume that the residence time of the pool water in Arkona Basin is so short that oxygen consumption in the pool may be neglected for simplicity. The validity of this assumption is discussed in Sect. 4.2 below.

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2.1.5 The Bornholm Basin model

The water flowing out of the Arkona Basin is feeding a dense bottom current in the Bornholm Basin that, after entrainment of ambient water, is interleaved in the stratified basin. It then has the volume flow Q_{inB} and the salinity S_{inB}. The computations of the flow in the dense bottom current are performed in the same way as described in Sect. 2.1.4 above. As described in Sect. 3.2 below, the empirical entrainment coefficient for the Bornholm Basin, EB, is tuned in order to achieve a good fit between model results and observations.

The outflow from the Bornholm Basin is geostrophic, driven by the horizontal west-east baroclinic pressure gradient between Bornholm Basin and Stolpe Channel. To resolve the flow vertically, we assume that the current speed, $v(z)$, is proportional to the horizontal baroclinic pressure gradient $dP(z)$ at each depth z, thus

$$v(z) \sim dP(z) \tag{10}$$

The difference in baroclinic pressure between the two basins, $dP(z)$ equals $P_{BB}(z) - P_{SC}(z)$. It is calculated using daily time series, interpolated from observed profiles, from the Stolpe Channel (pressure $P_{SC}(z)$) and modeled profiles from the Bornholm Basin (pressure $P_{BB}(z)$).

We then normalize $v(z)$ so that the total outflow from Bornholm Basin, i.e. the sum of the flows in all flowing layers, equals the flow we obtain from Eq. (8) modified for use in the Bornholm Basin. We use the salinity S_{SC} in the topmost 30 m in the Stolpe Channel to compute the reference salinity corresponding to S_B used in Sect. 2.1.4 above.

Our goal is to achieve a good model of hydrographic and oxygen conditions below 60 m in the Bornholm Basin. Since the new deepwater is not in contact with the sea surface after having entered the Arkona Basin we choose not to model the upper layers driven by processes at the sea surface. Instead we use interpolated observational data to describe the upper layers.

The rate of oxygen consumption in the basin, OXC, should vary with the export production from the euphotic zone, which should be proportional to the total phosphorus

content, TP, in the 0–40 m layer in winter. The reference oxygen consumption rate from Stigebrandt and Kalén (2012) (OXC_{SK}) was accordingly multiplied by the factor TP/TP_n where TP_n is the average phosphorus content in the period 1990–2010.

$$OXC = \begin{cases} OXC_{SK}\frac{TP}{TP_n}, & O_2 \geq 1.5\,mL\,L^{-1} \\ B \times OXC_{SK}\frac{TP}{TP_n}, & 1.5\,mL\,L^{-1} > O_2 > 0\,mL\,L^{-1} \\ OXC_{SK}\frac{TP}{TP_n}, & O_2 < 0\,mL\,L^{-1} \end{cases} \tag{11}$$

To adapt consumption to observations, we need to decrease the consumption rate by a factor B in the interval $1.5 > O_2 > 0\,mL\,L^{-1}$, see Sect. 3.2. The reason for this is oxidation due to reduction of nitrate, nitrite and iron and manganese oxides/hydroxides. Equation (11) is used during stagnant conditions without advective water exchange. However, when the bottom water is oxidized by inflowing new deepwater, it is assumed that both hydrogen sulfide and ammonia, if present in the water column, are oxidized as described in Sect. 3.1. To oxidize 1 mole of hydrogen sulfide to sulfate, 2 moles of oxygen are needed and to oxidize 1 mole of ammonium to nitrate, 1.5 moles of oxygen are needed.

2.2 Biological issues

2.2.1 Critical oxygen concentrations and re-colonization

In their reviews of marine benthic hypoxia, Diaz and Rosenberg (1995) and Karlson et al. (2002) concluded that critical oxygen concentrations for marine benthic animals varied between species. Mass mortality occurred when the oxygen concentration dropped below the range 0.5 to $1\,mL\,L^{-1}$. Rosenberg et al. (2002) studied the reactions of benthic animals in the field during successively declining oxygen concentrations and found that mass mortality occurred at $0.7\,mL\,L^{-1}$. However, behavioral responses and effects on reproduction occur in higher concentrations – commonly around $2\,mL\,L^{-1}$.

The value of $2\,\mathrm{mL\,L^{-1}}$ has been used as the limit below which hypoxia is defined and hypoxic waters are considered to harbor only a reduced benthic macrofauna under stress with adverse behavior and lack of reproduction ability (Conley et al., 2009a; Diaz and Rosenberg, 1995). In hypoxic areas where the sediments also contain toxic hydrogen sulfide close to the sediment surface, the conditions will be particularly unsuitable for benthic animals. Where oxygen minimum zones occur naturally, like in the eastern Pacific, animals are adapted to oxygen deficiency and tolerate oxygen levels close to zero (Levin, 2001). Such adaptations have not occurred in the Baltic, which is a young sea established after the last glaciation (Elmgren, 2003). Fish species are generally less tolerant to low oxygen concentrations and show avoidance reactions and reproductive failure in oxygen saturations $< 50\,\%$ ($\approx 4\,\mathrm{mL\,L^{-1}}$) (Breitburg, 2002).

The recovery process of a formerly hypoxic/anoxic system that is re-oxygenated depends on the severity and extent to which the system was disturbed. The succession of benthic animals will most probably follow the Pearson and Rosenberg (1978) paradigm, which shows that the colonization begins with small, opportunistic species in high numbers, which will facilitate the sedimentary habitat for the later successional stages. Transitory immigrants will follow the pioneering stages and, finally, a more mature benthic community will establish. In areas where the sediments have a high organic content because of heavy loadings, the time of recovery to pre-pollution times could be six to eight years (Rosenberg, 1976). In a Swedish fjord, where the low oxygen concentrations caused mass mortality, but the near-bottom water was soon re-oxygenated, the recovery of the benthic fauna to pre-hypoxic compositions in this case took about two years (Rosenberg et al., 2002), the reason being that the sediment was not over-enriched by organic matter. In an experiment with oxygenation of the basin water in the usually anoxic By Fjord, on the west coast of Sweden, colonization of the bottoms started after about one year when the sediment surface had been oxidized (Stigebrandt et al., 2014).

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2.2.2 The cod reproduction volume (CRV)

The cod reproduction volume (CRV) is defined as the water volume where the physical properties allow the survival of cod eggs, a prerequisite of successful cod spawning. The criteria that need to be fulfilled are salinity $S > 11$, oxygen concentration $O_2 > 2\,\mathrm{mL\,L^{-1}}$ and temperature $T > 1.5\,°\mathrm{C}$. Vallin et al. (1999) conclude in their review that the combination of high salinity and high availability of oxygen is necessary for large recruitment of cod in the Baltic Sea. Thus, large recruitment is only possible after major inflows of new deepwater. Since inflow of new deepwater in general occurs during the winter months (Matthäus and Schinke, 1994) and oxygen consumption causes reduction of the oxygen concentration over time, the CRV will be smaller later in the spawning season (Vallin et al., 1999). During present environmental conditions, the Bornholm Basin has been shown to be the most favorable area for successful cod reproduction in the Baltic Sea (MacKenzie et al., 2000). The season for cod spawning has shifted from the spring months towards the summer months (e.g. Hinrichsen et al., 2007; Vallin et al., 1999). To reveal seasonal effects, CRV in the second and third quarters of the year should be presented separately.

3 Data and tuning

3.1 Data

To calculate the inflow Q_{BS} we use Eq. (1), which needs data on the horizontal mean sea level h and the fresh water supply. For a first estimate of Q_{BS} with the temporal resolution 1 day, we use five days running averages of the observed sea level in Stockholm. Strictly, one should use the horizontal mean sea level of the Baltic Sea h for the computations of Q_{BS} but since Stockholm is situated in the nodal line of the first mode of sea surface oscillations, we use the sea level in Stockholm as a proxy for horizontal mean sea level (e.g. Samuelsson and Stigebrandt, 1996). For Q_{fBS} we use a time

series of monthly values of river runoff to the Baltic Sea, with the Arkona Basin as the western limit. To this, we add the average annual cycle for precipitation on minus evaporation from the sea surface (Rutgersson et al., 2002). The average of Q_{fBS} in the period 1990–2010 was calculated to $15\,500\,\mathrm{m^3\,s^{-1}}$.

The start value for H_{fK}, the thickness of the freshwater pool in the entrance area, is estimated from salinity observations from Anholt E ($56°40'$ N, $12°7'$ E, ▼ in Fig. 1) from the end of November 1989 to the end of January 1990. Of 1636 available vertical hydrographic profiles from southern Kattegat ($55°0'$–$56°12'$ N, $10°48'$–$12°0'$ E, white box in Fig. 1), all temperature and oxygen data from 5 m depth (1635 and 782 respectively) are used to compute monthly averages to represent the annual cycle of the surface layer in the entrance area. All temperature and oxygen data measured in deep water of salinity ≥ 32 are used similarly to compute monthly averages representing the seawater annual cycle. The properties of the inflowing new deepwater are then given by the monthly averages and the proportions of surface and deep water in the specific inflow, as described by Eq. (6).

We use all available data (historic vertical hydrographic profiles) at 15 m depth from 1990 to 2010 from the eastern half of the Arkona Basin ($54°42.0'$–$55°9.0'$ N, $13°41.4'$–$14°18.0'$ E, dashed box in Fig. 1) (455 observations) to compute monthly averages of salinity, temperature and oxygen describing the properties of the upper layer in the Arkona Basin. Water with these properties is entrained into the dense bottom current with new deepwater. 421 observed vertical salinity profiles from the same area, going down to at least 35 m depth and with bottom salinity > 11.6, were used to calculate the baroclinic outflow from the dense bottom pool according to Stigebrandt (1987a). The observed outflow is compared to model estimates for different values of the entrainment coefficient EA, see Sect. 3.2 below.

In the Bornholm Basin, we use 361 profiles from the station BY4 ($55°23.0'$ N, $15°20.0'$ E, white o in Fig. 1) and 247 observed vertical hydrographic profiles from the station BY5 ($55°15.0'$ N, $15°59.0'$ E, white x in Fig. 1) from 1990 to 2010 for the following computations. An average of BY5 profiles from 17 November 1989 and

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24 January 1990 is used for the initial water column properties of the model on the start date (3 January 1990). The observed salinity and oxygen concentration profiles from BY4 and BY5 are interpolated to form daily time series which are then compared to the corresponding time series resulting from the model. These vertical profiles are also used to compute the "observed" cod reproduction volume during the second and third quarter of each year from 1990 to 2010. This is then compared to the modeled cod reproduction volume. The interpolated daily time series from BY4 are chosen to represent the state variables (salinity, temperature and oxygen) above 30 m depth in the Bornholm Basin.

From observations obtained at BY5 we construct time series of annual winter content of total phosphorus in the uppermost 40 m, TP, to estimate the year-to-year variation of the export production supplying organic matter below 60 m depth in the Bornholm Basin as described in connection to Eq. (11).

From observations from depths $\geq$ 85 m at BY5 (43 observations) we have estimated that changes of the mole ratio between ammonium and hydrogen sulfide equals 1/5 under anoxic conditions. Furthermore, we estimated that the concentration of ammonium when the water turns from oxic to anoxic equals 4 mmol m^{-3}. Ammonium and hydrogen sulfide are oxidized when water exchange mixes old oxygen-free deepwater and new deepwater as described in Sect. 2.1.5.

219 vertical salinity and temperature profiles from the Stolpe Channel area (55°9′– 55°24′ N, 16°42′–17°54′ E, black box in Fig. 1) are used to form the interpolated daily time series used to calculate P_{SC} for the calculation of the geostrophic outflow from the Bornholm Basin into the Baltic Sea. The same data set is used to calculate a time series of S_{SC}, the mean salinity of the upper 30 m in Stolpe Channel.

Finally, hypsographic data for the studied sub-basins of the Baltic Sea were obtained from Seifert et al. (2001). The horizontal surface area of the Baltic Sea, A_{BS}, used in Eq. (1) equals 3.7×10^{11} m^2. For other data sources, see Acknowledgements.

3.2 Tuning

Model coefficients have been tuned in order to improve the match between computed and observed data. Accordingly the buffer volume located at the entrance sills equals 30 km^3 with a restoration time limit of seven days. When the model run begins, the buffer volume is empty. This approach gives long-term average flow rates to/from the Baltic of 16 400/32 100 m^3 s^{-1}, respectively, for the investigated period 1990–2010. Observations from 15 m depth in the Arkona Basin for the same period give an average outgoing salinity of 7.9. If salt content and volume of the Baltic are assumed to be preserved over the period 1990–2010, new deepwater needs to have an average salinity of 15.5. This is achieved in the model by setting $S_0 = 33$ and H_K in Eq. (5) equal to 6.7 m. The freshwater pool area, A_K, was adjusted to avoid too salty inflows. The value 42 000 km^2, proposed by Stigebrandt (2001), gives good correspondence between the model and observations. The thickness H_A of the dense Arkona pool is measured upwards from 41 m depth, which is the sill depth of the Arkona Basin (Stigebrandt, 1987a).

The empirical entrainment coefficient of dense bottom currents, which determines the rate of increase of the volume transport per unit depth, was tuned to give a good fit between model results and observations. For the Arkona Basin, EA = 0.02 m^{-1} gives a satisfactory relationship between salinity and volume flowing out from the Arkona Basin as seen in Fig. 2. For the Bornholm Basin, EB = 0.01 m^{-1} gave a good fit between observations and modeled values of salinity, oxygen and temperature. A wide and thin bottom gravity current, flowing on a sloping plane, entrains more ambient water than a thicker and narrower current that runs in a canyon (Stigebrandt, 1987b). This qualitatively explains the difference between EA (plane) and EB (canyon).

The salinities S_{in} of seven large modeled inflows into the Bornholm Basin were adjusted manually in order to get a better fit between modeled and observed salinity in the Bornholm Basin. The salinity was slightly increased in the inflows that occurred in 1994, 2003 and 2010 and slightly decreased for the inflows in 1997 and 2007. The salinity in two subsequent inflows during 1998 was increased and decreased, respectively.

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To conform to observations in the Bornholm Basin, we halved the oxygen consumption rate for the interval $1.5 > O_2 > 0\,\mathrm{mL\,L^{-1}}$, thus using $B = 0.5$ in Eq. (11).

4   Results and discussion

The modeled flows of new deepwater in different salinity intervals from the Baltic entrance area, out from the Arkona Basin and out from the Bornholm Basin, respectively, are shown in Fig. 3. As expected, the salinity decreases and the volume flow increases downstream due to mixing in the basins.

4.1   Arkona basin – deepwater flow estimates using historic vertical hydrographic profiles vs. pool model

The distributions of average flow rates among different salinities according to the Arkona Basin pool model and computed from hydrographical observations are shown in Fig. 2. We find that hydrographical observations for the period 1990–2010 give the ensemble average flow rate from the dense pool of $22\,300\,\mathrm{m^3\,s^{-1}}$ with an average salinity of 13.7, which is similar to the results estimated from hydrographical observations for an earlier period in Stigebrandt (1987a). The corresponding values according to our pool model are $19\,900\,\mathrm{m^3\,s^{-1}}$ and 14.4, respectively. The modeled outflow of deepwater from the Arkona Basin thus follows the observations rather well.

4.2   Bornholm Basin – observed and modeled vertical hydrographic profiles

The observed salinity at station BY4 (Fig. 4a) can be compared to the modeled salinity (Fig. 4b). The modeled salinity is well correlated to observations below 40 m. The model also describes the temperature variations in the Bornholm Basin rather well during the modeled period (not displayed). The evolution of the salinity is clearly governed by three different physical processes that are included in the model, (i) during deepwater inflows salinity increases and the halocline rises, (ii) between inflows vertical diffusion

dominates whereby salinity decreases with time, (iii) outflows through Stolpe Channel lower the halocline.

The observed oxygen concentration at BY4 (Fig. 5a) is compared to modeled levels in Fig. 5b. In the plots, hydrogen sulfide is expressed as negative oxygen such that 2 moles of negative oxygen correspond to 1 mole of hydrogen sulfide. The actions of the three advection/diffusion processes seen in Fig. 4 can also be seen in Fig. 5. However, in addition oxygen is influenced by oxygen consumption (also included in the model) that decreases the oxygen concentration. Vertical diffusion of oxygen from above tends to compensate for oxygen consumption in the enclosed basin. The modeled oxygen concentration largely follows the observations from station BY4 but also fits data from station BY5. In Fig. 5a we do not see any traces of small volumes of water with relatively low oxygen concentration that might come from the Arkona Basin in the autumn. This is further discussed below.

The comparison between observations and the reference model run (Figs. 4a, b and 5a, b) shows that the timing of the modeled inflows corresponds well to the observations in the Bornholm Basin. However, the observed vertical hydrographic profiles used for validating the model generally have a monthly time resolution, usually with still lower resolution during winter months. The resolution of observations is thus very low compared to the daily resolution of the model results. It is therefore possible that the observations may miss the precise timing of the inflow and the exact water properties just after a deepwater inflow. Sometimes the observed bottom salinities appear to be higher than the modeled bottom salinities. This suggests that the model produces a more homogenous water column in the deepest parts of the basin, but with similar salt content as observed in the whole water column.

Observations show that the halocline uplift after larger inflows, such as the one in early 2003, can persist for up to one year. In the model, such uplifts seem to decay more quickly suggesting that the modeled outflow might be too fast. However, another explanation may be that the vertical resolution of observations is typically 10 m while the model resolution equals one meter. Observations would then give the impression

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of a persistent halocline in the depth interval 50 to 60 m. Since the modeled conditions in the basin water below sill depth (60 m) appear to follow observations quite well, the possibly slightly too rapid outflow (above 60 m depth) should not be critical for the results in the basin water (below 60 m depth).

Our model assumes that there is no oxygen consumption in the new deepwater when passing the Arkona Basin. We will investigate whether or not this assumption may be violated in periods. The volume of the basin water, below the sill level at 41 m depth, in Arkona Basin is about 12 km^3, estimated using the topographic database in Seifert et al. (2001), which is only 8 % of the volume of the basin water in Bornholm Basin. If initially of high salinity, water may stay in the basin from spring to autumn whereby it may obtain low oxygen concentration due to local consumption. The lowest record of oxygen concentration in the Arkona Basin, according to the data available for this study, equals 0.47 mg L^{-1} (December 1997) (1 mL L^{-1} ≈ 1.4 mg L^{-1}).

In Table 1 are given volume and mean salinity and mean oxygen concentration of all 10 cases of hypoxic water with volumes > 3 km^3 estimated from hydrographical observations in the Arkona Basin during the period 1990–2011 (22 years). When the basin water resting in the Arkona Basin eventually is flushed by a new denser deepwater, parts of the flushed basin water will be entrained into the new deepwater while the remainder is lifted above the sill level, cf. the process of water exchange in the By Fjord (Stigebrandt et al., 2014). After the uplift, the flushed deepwater will be contained in a rather thin layer due to topographical reasons. For example, a layer of volume 10 km^3 centered at 35 m depth will be only 2 m thick. This layer is surrounded by water of higher oxygen concentrations why oxygen efficiently may diffuse into the hypoxic layer from both above and below. The oxygen content of the layer of flushed basin water is thus expected to increase before it flows through the Bornholm Channel.

In the Bornholm Basin, the flushed basin water from the Arkona Basin will entrain ambient water of low salinity (average 9) and high oxygen content (average 9.8 mg L^{-1}) at the rate EB, meaning that the volume has increased by 30 % when reaching 60 m depth. The estimated salinity and oxygen concentration of the water when reaching the

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depth 60 m in the Bornholm Basin are given in Table 1. As can be seen, the water is in none of the cases hypoxic when reaching 60 m and the salinity is usually too low to permit penetration down into the deep basin. The figures for oxygen concentration given in Table 1 should be conservative, because the probably substantial diffusive supply of oxygen by vertical mixing in the Arkona Basin after the uplift discussed above is not accounted for.

4.3 Hydrographical changes of pumping in Bornholm Basin

The model was run with the pumping turned on with an initial (jet) mixing rate $\alpha = 10$, meaning an eleven-fold increase of the volume flow, and plume entrainment $E = 0.05$, meaning a 5 % increase of the volume flow when the buoyant plume rises 1 m. This fits well with the observed mixing of the pumped flow in the By Fjord experiment (Stigebrandt et al., 2014). The natural vertical mixing in the basin is modeled as specified in Stigebrandt and Kalén (2012). The results with pumping turned on are compared to the model results with pumping turned off presented in Sect. 4.2 above. We are primarily interested in finding out the volume transport by pumping required to maintain oxic conditions in the deep water. The water intake depth was set to 30 m but also intake depths at 0 and 40 m were tested for comparison. In all model runs the horizontal pump outlet is at 90 m depth. The investigated pumping rates were 800, 1000, 1200, 1500 and 4000 $m^3 s^{-1}$. At the present stage, it is not necessary to specify the number of pumps and the outlet velocity needed in practice to achieve the properties of the modeled pumping.

The salinity and oxygen content in the water column with a pumped volume flow of 1000 $m^3 s^{-1}$ can be seen in Figs. 4c and 5c, respectively. Notice that pumping leads to an increased number of inflow events and that the salinity never falls below 11 beneath 65 m depth. Notice also that the oxygen content for this pumping rate only rarely drops below 1.5 mL L^{-1} and never below 1 mL L^{-1}. It is also to be noted that, at all occasions when oxygen content in the pumping scenario (Fig. 5c) drops below 2 mL L^{-1}, the oxygen content in the model reference run (Fig. 5b) is slightly too low compared to

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observations (Fig. 5a). The oxygen conditions produced by the model in the basin water are further discussed below.

The effects of different pumping rates on the mean flow rate, the number and properties of inflow events and on the basin salinity, temperature and oxygen levels, represented by values from 75 m depth, are shown together with the effects on the CRV in Table 2. It can be seen that an increased pumping rate leads to an increased rate of water exchange at greatest depths. For the highest pumping rates we get the highest average inflow rates but notice fewer inflow events, since inflow events merge, compare Table 2 for pumping rates 1500 and 4000 $m^3 s^{-1}$. Like in the By Fjord experiment (Stigebrandt et al., 2014), the average salinity in the basin is decreased by pumping. This explains why pumping allows for increased basin ventilation by inflows that previously were not able to penetrate to the bottom of the basin. The minimum basin salinity decreases approximately linearly with increased pumping rate (approximately −0.1 per 100 $m^3 s^{-1}$), thus if the pumping rate exceeds 2200 $m^3 s^{-1}$ the minimum salinity may occasionally drop below 11, which affects e.g. the minimum CRV for the highest investigated pumping rate 4000 $m^3 s^{-1}$ (see Table 2).

Inflows contribute three to five times more oxygen than the pumped water, dependent of the magnitude of the pumped volume flow (Table 2). By pumping 1000 $m^3 s^{-1}$, the inflow of oxygen by new deepwater is increased by 144 %. An additional increase of the oxygen inflow by about 56 % is contributed by the pumped water. Both average and minimum oxygen levels in the basin water increase significantly with increased pumping. A pumped volume flow of 1000 $m^3 s^{-1}$ increases the average oxygen level at 75 m from 1.9 to 3.6 $mL L^{-1}$. The minimum (average of the 2 % lowest values) level increases from −2.4 (anoxic), to 1.4 $mL L^{-1}$ respectively. The pumping makes particularly large difference in the deepest parts of the basin because the vertical downward motion in the whole basin, induced by the pumping, transports oxygen and other substances from higher levels much more efficiently than the natural vertical diffusion does, cf. Fig. 5c and an example in Stigebrandt and Kalén (2102).

In the autumn, hypoxic basin water from the Arkona Basin may be exchanged and flow into the Bornholm Basin as discussed at the end of Sect. 4.2. The oxygen deficit of a package of $20\,km^3$ of water with O_2 concentration 3.9 instead of $9.8\,mg\,L^{-1}$ (oxygen deficit 6 mg per kg water) equals 120 ktonnes of oxygen. The total oxygen supply in the reference case without pumping equals 647 ktonnes yr^{-1} (Table 2). For pumping $1000\,m^3\,s^{-1}$ the supply is estimated to 1940 ktonnes (360 ktonnes from pumping and 1580 ktonnes from inflows). The oxygen supply has thus increased by 200%. If we include the package with low oxygen content the total supply decreases to 1720 ktonnes. This gives an increase of oxygen supply by 181%. This should be a conservative figure, because most of the packages in Table 1 should not reach the greatest depth of the Bornholm Basin with the suggested rate of pumping, as their salinity is too low. In addition, the packages from the Arkona Basin are released in September or later, which is too late in the year to affect the CRV because the spawning occurs in the time period April–September. It would be easy to complement the oxygenation system in Bornholm Basin by a system in the Arkona Basin to keep the small deepwater volume there oxygenated and in this way eliminate the possibility that low oxygen concentrations develop in the Arkona Basin. This would actually be quite beneficiary for the bottom fauna in the deepest part of the Arkona Basin.

The average basin temperature is 8.1 °C in the reference case and does not change significantly with any of the tested pumping rates. Hence, the long term basin average temperature does not appear to be affected by pumping when the pump intake depth is in the interval 30–40 m. However, the average and maximum temperatures at 75 m depth increase by approximately 0.2 °C if the water intake is located at the sea surface. The salinity changes very little if the intake depth is changed to 0 or 40 m.

4.4 The effect on cod reproduction volume by artificial mixing

The CRV was calculated as the volume below 61 m depth that fulfills the definition in Sect. 2.2.2. These conditions may also be met above 61 m depth, but cod eggs present above this depth might be transported out from the Bornholm Basin by the outflow

and, during that process, salinity will likely decrease due to entrainment. Therefore, the water volume above this depth is not included in the calculation of CRV. The total water volume below 61 m depth is 142 km^3, which is thus the upper limit for the CRV in our approach.

The quarterly CRV from the model (no pumping) is compared to the CRV calculated from observations from BY4 and BY5 (Fig. 6). The second (April to June) and third (July to September) quarters are the periods when cod spawning occurs and these periods are highlighted in Fig. 6. The minimum and average CRV given by the model are 16 and 88 km^3, respectively. The corresponding values from observations were calculated to 0 and 78 km^3 for BY4 and 29 and 89 km^3 for BY5. It is apparent from these values and from Fig. 6 that the modeled CRV generally follows the observed CRV well.

There are apparently observations during the studied period when the CRV is small and even vanishes (Fig. 6). We have partitioned the CRV data by quartiles, Q1–Q4, and calculated the average for each partition. The average of the lowest 25 % of the CRV values (Q1) based on spring and summer model values equals approximately 40 km^3 (Fig. 7). By pumping, 1000 m^3 s^{-1}, this value would increase to > 65 km^3 (see also Table 2 for results of other tested pumping rates). Figure 7 shows that it is in particular during years with small CRV's that pumping might make a big difference. Thus, by pumping there will be a large CRV each year, and cod recruitment should not fail in any single year due to lacking CRV.

The model runs with the pump intake at 0 and 40 m depth showed a similar variation as the runs in Table 2, but with somewhat smaller CRV for the 40 m intake. For example, the Q1 and minimum CRV with a pump volume flow of 1000 m^3 s^{-1} was 66 and 54 km^3 for the 30 m intake depth. For an intake placed at 40 m these values were 60 and 46 km^3. The corresponding values with the intake located at the sea surface are 70 and 54 km^3.

Gustafsson et al. (2008) performed computations showing that halocline ventilation might not increase CRV. This is opposite to the conclusion reached in the present paper. The reason for the diverging opinions is that Gustafsson et al. (2008) reached

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their result without studying the conditions in the Bornholm Basin explicitly, the only basin with successful cod recruitment during the last decades.

4.5 Outflow of deepwater from the Bornholm Basin

The effects of artificial mixing on the water flowing from the Bornholm Basin to the Stolpe Channel were investigated for the pumping rate $1000\,\mathrm{m^3\,s^{-1}}$. The total geostrophic transport, reference flow ca. $30\,000\,\mathrm{m^3\,s^{-1}}$, increases by approximately 2 %. Transports with the highest salinities (ref ca. $9000\,\mathrm{m^3\,s^{-1}}$; $S > 13$) decrease by 17 %. Flows with oxygen content $< 2\,\mathrm{mL\,L^{-1}}$ disappear completely while flows in general are shifted towards higher oxygen content. For instance, volume transports with very high oxygen content (ref. ca. $6300\,\mathrm{m^3\,s^{-1}}$; $O_2 > 7\,\mathrm{mL\,L^{-1}}$) increase by 10 %. The model thus shows that the salinity distribution of the outflow to Stolpe Channel becomes narrower and the oxygen concentration becomes higher when artificial mixing is applied. This should decrease the vertical stratification, as shown by a numerical experiment in Stigebrandt (1987b), and improve the oxygen conditions in the deeper parts of the Baltic proper east of Bornholm Basin.

4.6 Benthic fauna, potential new biomass and fish feed

The benthic fauna in the Bornholm Basin was studied in the 1950s and 1960s by Demel and Mulicki (1954), and Leppäkoski (1969), respectively, in relation to changes in salinity (13.5–17) and oxygen concentrations (0–5 $\mathrm{mL\,L^{-1}}$). Up to seven species were recorded in the area deeper than 70 m, and in the 1950s, the bivalve *Astarte borealis* was the most common and with the highest biomass. Among potential colonizers in the deeper Bornholm Basin, should the oxygen be high over several years, are the bivalves *Astarte borealis, Macoma balthica* and *Macoma calcarea*; the polychaetes *Halicryptus spinulosus, Scoloplos armiger, Capitella capitata* and *Harmothoe sarsi* and the crustaceans *Diasylis rathkei, Pontoporeia affinis* and *Pontoporeia femorata.* In the 1960s, bivalves were rare and instead polychaetes, particularly *Scoloplos armiger,* colonized

the bottoms during periods of re-oxygenation. However, the oxygen could drop from 4–5 mL L^{-1} to < 1 mL L^{-1} during a few months.

The biomasses of the benthic fauna in the 1960s were much lower at depths > 70 m than in the 1950s, probably as a result of too short periods of suitable oxygen conditions to let a healthy community to establish. The average biomass at seven stations > 70 m depth in the 1950s was 17.5 g m^{-2} wet weight. Karlson et al. (2002) estimated that about 50 000 tons of benthic biomass could be missing in the Bornholm Basin during hypoxic/anoxic conditions in the area. Should re-oxygenation of the Bornholm Basin also have positive secondary effects on the oxygen conditions in the Baltic proper an additional biomass in the order of 1 400 000 tons could be the result, as was estimated by Karlson et al. (2002).

A local example of re-colonization of benthic fauna of formerly dead bottoms is from the rather enclosed inner Stockholm archipelago. After reducing the input of phosphorus from the sewage of the city, the oxygen concentrations in the deeper waters improved. In 1996 to 1998 all sediment samples at 20 to 50 m were black and reduced (Rosenberg and Diaz, 1993), but in 2008 seven of these eight stations had an oxidized sediment surface (Karlsson et al., 2010). The authors recorded abundances of between 2300 and 5600 ind. m^{-2} and biomasses of between 6 and 65 g m^{-2} (wet weight, excluding the reduced station). The invasive polychaete *Marenzelleria neglecta* was totally dominant at all stations and is a fast colonizer in great numbers. *Marenzelleria* spp. could possibly also colonize the deeper parts of the today anoxic/hypoxic Bornholm Basin and the Baltic proper if these areas were oxidized. Should this worm establish in the present defaunated area of ca. 80 000 km^2 in the Baltic proper with biomasses of the same weights as in the Stockholm archipelago, the total biomass could be in the order of 0.5 to 5.0 million tons, i.e. in the same order as calculated above. As a comparison, Cederwall et al. (1999) found *Marenzelleria viridis* invading the Gulf of Riga in 1993 to 1996 in average biomasses of 11.8 g m^{-2}. The species *Marenzelleria* spp. is described in more detail in Sect. 4.7.

4.7 Bioturbation and related benthic biogeochemical processes

Benthic faunal activities have significant impact on biogeochemical processes in the sediment globally through bioturbation, i.e. reworking the sediment by digging and burrowing; oxygenation of burrows and voids; and biodeposition of particles on the sediment surface and into the water column. Different species have different activity patterns, and in general larger and more active species have a larger impact on sediment and irrigation processes. As the number of species in the brackish Baltic Sea is reduced compared to in more saline areas, the number of functions are also comparatively lower. Based on if animals are mobile or sessile, and where and how they feed, the number of "functional groups" can be calculated (Pearson and Rosenberg, 1987). Based on such a scheme the number of functional groups in the Skagerrak are above 20 but only 8 in south Baltic (Bonsdorff and Pearson, 1999). Species missing in the Baltic (until recently) were those that were digging deep and feeding within the sediment (sub-surface deposit feeders) and animals using tentacles as feeding appendages. Thus commonly only the upper few centimeters were bioturbated in the Baltic compared to several tens of centimeters in oceanic waters.

The spionid polychaete *Marenzelleria* spp., a recent invasive species first recorded in 1985, is a deep burrowing worm with consequences for the biogeochemical processes. *Marenzelleria* spp. consists of three different species that look alike and are difficult to separate into species. These species are nowadays common from shallow to deep waters and often dominant in many oxygenated areas of the Baltic. *Marenzelleria* spp. are both surface and subsurface deposit feeders and burrow deep (10–30 cm) into the sediment. They can locally reach numbers as high as 30 000 ind. m^{-2} (Zettler et al., 2002).

By its intrusion into the Baltic, *Marenzelleria* spp. have added a new dimension to the Baltic sedimentary system. This genus constructs galleries in the sediment down to more than 10 cm and adds new niches and complex functions to the ecosystem services. The three species-complex of *Marenzelleria* are spread from shallow waters

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down to deeper oxygenated bottoms from the southern Baltic up to the Bothnian Sea and the Gulf of Finland. The genera are dominant in many areas and occur, with great variations, frequently in numbers of between 500 and 5000 $ind. \, m^{-2}$, but with a somewhat heterogeneous distribution. The *Marenzelleria* spp. are potential invaders of the deeper Baltic sediments if the oxygenated areas are extended. It can colonize new areas by settlement of pelagic larvae or by actively swimming or crawling adults (Norkko et al., 2011). However, to what extent they will colonize these today defaunated areas and in which densities remain unknown. Nonetheless, this is most likely to happen and the new invader will, in addition to other colonizers, have a significant impact on the fluxes of nutrient in the sediment through their activities in the deep galleries.

The pool of inorganic phosphate in the sediment in the Baltic is clearly related to the oxygen concentration in the bottom-near water; phosphorus is bound to the sediment during oxic conditions and released at anoxic conditions (e.g. Conley et al., 2009a; Stigebrandt et al., 2013). Thus, when animals are present in the sediment their bioturbation and ventilation of the sediment will increase the amount of phosphorus bound in the sediment. When *Marenzelleria* spp. occur in high densities they can, through their deep burrowing activity, have a significant impact on these biogeochemical cycles. Quintana et al. (2011) found that *Marenzelleria viridis* had two types of ventilation – a muscular pumping of water out of the burrow and a ciliar pumping of water into the burrow. Significant amounts of water percolate upwards to the sediment surface, which will have a significant effect also on the reduction of sulfate in the sediment (Kristensen et al., 2011). Norkko et al. (2011) demonstrated in a model that *Marenzelleria* spp. had a significant density-dependent impact on the phosphorus cycling in the Baltic, where increasingly more phosphorus are bound to the sediment when the density increases beyond 3000 $ind. \, m^{-2}$. The great capacity of *Marenzelleria*-bioturbated sediments to store phosphorus was demonstrated by the fact that twice as much was bound annually to the nowadays oxic sediments in the Stockholm archipelago as annually could be removed by the sewage treatment plants in Stockholm.

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Another possibility is that changed redox conditions in the sediment and activity of *Marenzelleria* spp. could increase the flux of contaminants such as PCBs compared to if the sediment remains anoxic (Granberg et al., 2008). Hedman et al. (2008) showed that contaminants and cadmium could be either buried or remobilized in the sediment because of the activity of different infaunal species. The authors emphasized the importance of understanding the complex interactions between ecological and physiochemical processes when assessing the fate of contaminants in aquatic ecosystems. Thus, more research is needed before any general and firm conclusions could be made regarding the fate of contaminants and metals in the Baltic if the seabed is re-oxygenated. However, the By Fjord oxygenation experiment did not lead to increased leakage of toxic organic compounds or toxic metals from the sediments when these became oxidized (Stigebrandt et al., 2014).

Oxygenation of the deepwater should tend to decrease the oxygen consumption through the coupling between primary production and decreased leakage of phosphorus from deep bottoms and the following decreased phosphorus content in the surface layers. As in eutrophic lakes, a substantial fraction of the oxygen consumption might be due to fluxes of reduced substances from the sediments (Müller et al., 2012) why the decrease of oxygen consumption initially might be less than expected. However, provided the eutrophication of the Baltic proper does not increase, there is no reason to expect that oxygen consumption should increase when artificial oxygenation is turned on at the suggested pumping rate.

5 Concluding remarks

From simulations of the vertical circulation in the Bornholm Basin in the period 1990 to 2010 we have shown that the deep Bornholm Basin may be kept oxic by pumping down about $1000 \, \mathrm{m^3 \, s^{-1}}$ of water from around 30 m depth. The average oxygen concentration in the basin will then be $3.6 \, \mathrm{mL \, L^{-1}}$ and the minimum concentration is estimated to $1.3 \, \mathrm{mL \, L^{-1}}$, compared to 1.9 and $-2.4 \, \mathrm{mL \, L^{-1}}$ without pumping. The rate

of water exchange increases from 66 to 160 km^3 year^{-1}. Like in the By Fjord experiment (Stigebrandt et al., 2014), the increased water exchange is the main reason for why the supply of oxygen increases. The average salinity of the basin decreases from 15.4 to 14.4 while the average temperature of 8.1 °C remains almost unchanged.

The small CRV in certain years, endangering the recruitment of cod, would increase with oxygenation. The average of the smallest 25 % of CRV should, by pumping 1000 m^3 s^{-1}, increase from about 40 to more than 65 km^3. Thus, oxygenation by increased deepwater ventilation might be a powerful method to strengthen cod reproduction in the Baltic Sea in years when the CRV otherwise would be small and even vanishing.

Benthic biomass will be established by colonization of sea bottoms, which means a new food source for demersal fish like cod and flatfish. Re-oxygenation of the deeper basins in the Baltic will undoubtedly have a positive effect on the recruitment and food availability for cod, which would be beneficial for the fisheries in the long run. Today the total annual catches of cod are only in the order of 50 000 tons, whereas the potential might be 400 000 tons as it was in the 1990's (MacKenzie et al., 2002).

Reduced leakage of phosphorus from the sea bed in the Bornholm Basin by about 7500 tonnes per year (Stigebrandt et al., 2013) will reduce the phosphorus content in the surface layer and thereby contribute to decreasing eutrophication of the Baltic Sea. It will also reduce the magnitude and spatial extent of toxic cyanobacteria blooms in summer with large socioeconomic and economic benefits for tourism. The present investigation also indicates that an isolated effort in the Bornholm Basin might give rise to positive effects in the East Gotland Basin because the deepwater of the Bornholm Basin, which is exported to the East Gotland Basin, would be of lower salinity and contain less phosphorus and more oxygen. This should have positive impacts for the functioning of the ecosystem of the Baltic proper.

The present paper contributes an important part of an analysis of the possible consequences of artificial oxygenation of the deepwater in the Bornholm Basin. Since the consequences are suggested to be positive, artificial oxygenation by pumping

is recommended to be undertaken in the Bornholm Basin. It would be positive for the deepest bottom fauna to also oxygenate the Arkona Basin, which only requires a smaller effort. A carefully constructed observational program, like the one in the By Fjord experiment (Stigebrandt et al., 2014) should be run to study various ecological and biogeochemical effects of such an oxygenation.

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Table 1. Volume $V(A)$, mean salinity, $S(A)$, and mean oxygen concentration, O_2 (A) of hypoxic water observed in the Arkona Basin during the period 1990–2011. Also shown are estimated salinity, $S(B)$ and oxygen concentration, O_2 (B), when the water reaches 60 m depth in Bornholm Basin.

Day	Month	Year	$V(A)$[a]	$S(A)$	$O_2(A)$[b]	$S(B)$	$O_2(B)$[b]
31	08	1992	23.35	14.33	1.29	13.1	3.25
1	09	1994	15.13	15.07	2.24	13.82	3.99
15	12	1997	15.13	16.62	1.55	14.86	3.46
23	09	1998	19.09	14.12	2.14	12.94	3.91
25	09	2002	15.13	16.66	2.41	14.89	4.12
30	10	2007	19.09	14.53	1.71	13.25	3.57
17	09	2008	19.09	14.42	2.09	13.17	3.86
16	09	2009	11.50	15.30	2.61	13.85	4.27
15	09	2010	15.13	13.64	2.18	12.57	3.95
14	09	2011	8.26	13.01	1.90	12.08	3.72

[a] Volume is given in km^3.
[b] Oxygen values are given in $mg\,L^{-1}$.

Table 2. Model results for 1990–2010 with the pump intake placed at 30 m depth and for different pumping rates.

Pumped volume flow ($m^3 s^{-1}$)	Ref.[a]	800	1000	1200	1500	4000
Average inflow rate ($km^3 year^{-1}$)[b]	66.8	141.0	160.0	177.7	205.2	587.5
Number of inflows[c]	17	38	45	50	47	37
Average inflow salinity[d]	15.89	15.05	14.90	14.79	14.69	14.08
Average inflow of O_2 ($kton year^{-1}$)[b]	647	1385	1579	1769	2051	6076
Flow of O_2 via pump ($kton year^{-1}$)	0	288	360	431	536	1438
Average basin O_2 ($mL L^{-1}$)[e]	1.92	3.44	3.64	3.81	4.04	5.10
Minimum basin O_2 ($mL L^{-1}$)[f]	−2.41	1.26	1.38	1.45	1.68	3.78
Average basin salinity[e]	15.41	14.52	14.38	14.25	14.07	13.11
Minimum basin salinity[f]	13.34	12.57	12.46	12.32	12.09	10.80
Average basin temp (°C)[e]	8.16	8.12	8.11	8.11	8.09	7.92
Q1 of CRV (km^3)[g]	41.9	57.2	65.7	81.1	100.6	112.2
Minimum CRV (km^3)[h]	15.2	54.0	54.0	54.0	54.0	0

[a] Model reference values (pumping rate set to 0).
[b] Volumes annually reaching 90 m depth.
[c] The number of inflows is defined as the number of new inflow events reaching 90 m and lasting at least 24 h after a minimum of 1 week of stagnation.
[d] Volume weighted average for salinities of the inflow of new deepwater.
[e] The average of values from 75 m depth.
[f] The average of the 2 % lowest values at 75 m depth.
[g] The average of the 25 % smallest CRVs calculated for the cod spawning period, April to June and June to September.
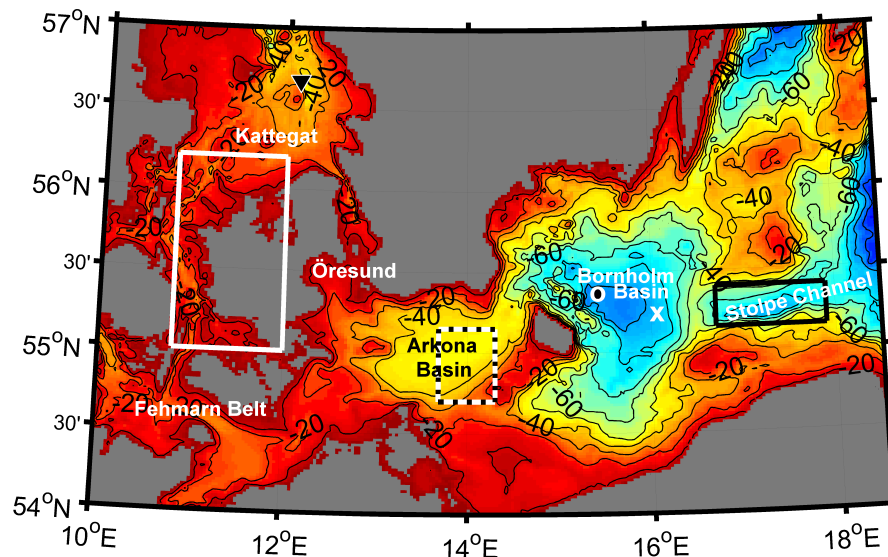[h] Calculated for the cod spawning period, April to June and June to September.

Figure 1. Map of the southern Baltic Sea area (hypsographic data from Seifert et al., 2001). Reference data for the Arkona Basin originates from the area within the white and black dashed rectangle and reference data station BY4 and BY5 in the Bornholm Basin is marked by a white o and x respectively. Data from the Stolpe Channel used for forcing of the model originate from the area within the black rectangle. The forcing data for the freshwater pool model were obtained from within the white rectangle and from the Anholt E station marked by.

Figure 2. Distribution of the average volume flow among different salinities for the modeled deep water flows from the Entrance Area, the Arkona Basin and the Bornholm Basin.

Figure 3. Distribution of observed and modeled average flow rates among different salinities for the baroclinic outflow from the Arkona Basin.
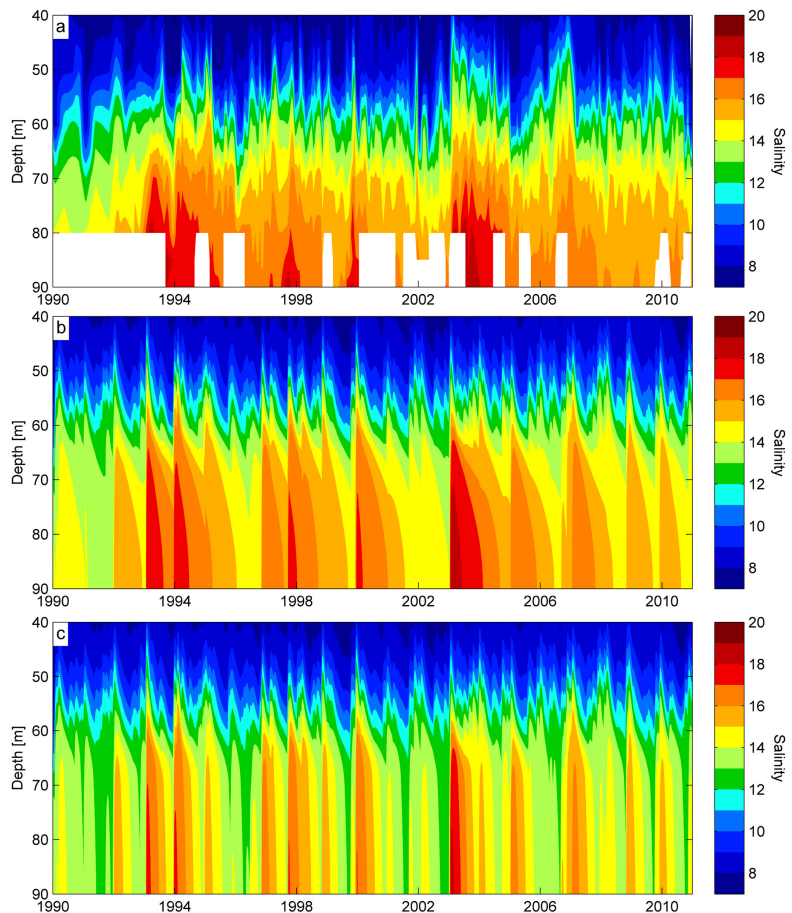
Figure 4. (a) The observed salinity from BY4, **(b)** the modeled salinity (reference run) and **(c)** the modeled salinity when pumping $1000 \, \mathrm{m^3 \, s^{-1}}$ from 30 m depth down to 90 m.
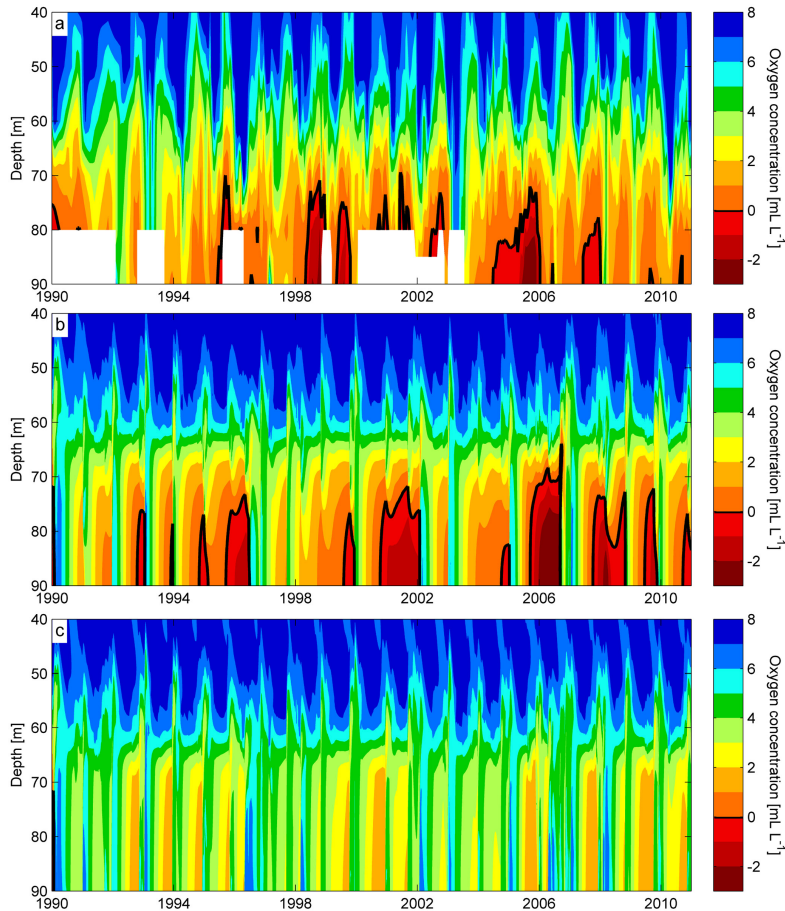
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Figure 5. (a) The measured oxygen content from BY4, **(b)** the modeled oxygen content (reference run) and **(c)** the modeled oxygen content when pumping $1000\,m^3\,s^{-1}$ from 30 m depth down to 90 m.
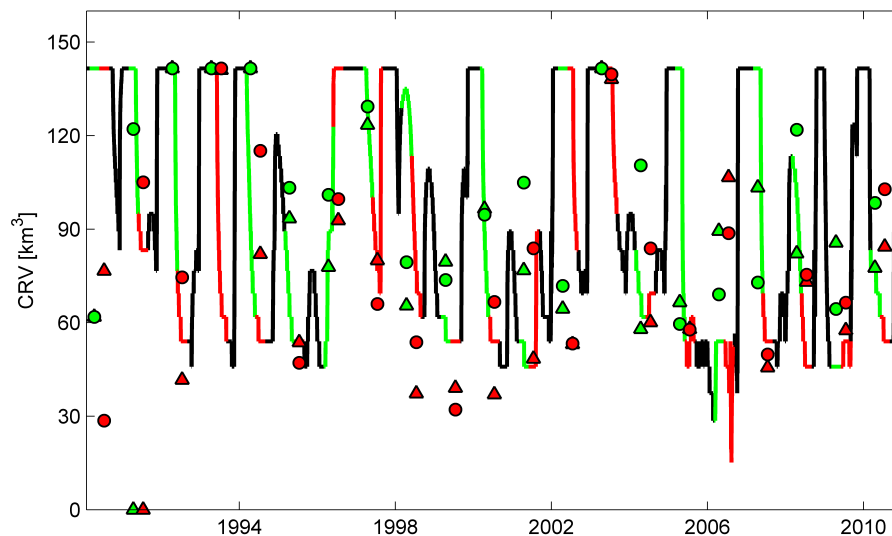
Figure 6. The modeled cod reproduction volume (CRV) (black line). The second and third quarter periods are highlighted by green and red, respectively. The second and third quarterly average cod reproduction volumes, calculated from BY4 and BY5 data, are represented by triangles and circles respectively.

Figure 7. Model average cod reproduction volumes (CRV) from the second and third quarters of the year separated by quartiles and averaged, here presented in rising order (*Q1–Q4).* Dark grey bars represent the model reference run (no pump) and the light grey bars represent the model with a pump volume flow of $1000\,\text{m}^3\,\text{s}^{-1}$.