Oceanologia (2017) 59, 69—80

Available online at www.sciencedirect.com

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ORIGINAL RESEARCH ARTICLE

Processes and factors influencing the through-flow of new deepwater in the Bornholm Basin

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Received 3 June 2016; accepted 11 September 2016
Available online 23 September 2016

KEYWORDS
Vertical mixing;
Inflow of new deepwater;
Salinity;
Currents;
Baltic Sea

Summary This paper is based on the idea that the hydrographical conditions in the Bornholm Basin, and any other basin, can be understood from knowledge of general hydromechanical principles and basin-specific factors. Published results on the variability of the vertical stratification are shown and discussed. Such analyses demonstrate the residence time of water at different depth levels. Different modes of currents forced by winds and by stratification gradients at open vertical boundaries are presented. Vertical mixing is discussed and published results for the Bornholm Basin are shown. An experiment demonstrates that the diffusive properties of the enclosed basin, i.e. below the sill depth of the Słupsk Furrow, can be computed quite well from the horizontal mean vertical diffusivity obtained from historical hydrographical observations. A published two decades long simulation of the vertical stratification shows that the through flow and modification of new deepwater in the Bornholm Basin can be well described based on existing knowledge regarding crucial hydromechanical processes. It also suggests, indirectly, that there should be a weak anticyclonic circulation above the sill depth, which is supported by current measurements.

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Peer review under the responsibility of Institute of Oceanology of the Polish Academy of Sciences.

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http://dx.doi.org/10.1016/j.oceano.2016.09.001

1. Introduction

When the sea level is higher in Kattegat than in the Arkona Basin, salty water flows across the shallow sills in Fehmarn Belt and the Öresund into the Arkona Basin. Because this water has higher salinity and density than the surface water of the Arkona Basin, it descends along the seabed whereby the potential energy of the dense water is transferred to kinetic energy. Due to friction against the seabed and the overlying water, kinetic energy is transformed to turbulent energy that causes entrainment of the overlying water whereby the volume flow of the inflowing water increases and its salinity decreases. The inflowing saline water is the source of new deepwater in the Baltic proper. In the present paper it is called the new deepwater. Finally, the new deepwater is added to the dense bottom pool in the Arkona Basin. The water from this pool is evacuated through the Bornholm Channel. When this occurs it will undergo further dilution with less dense ambient water in the Bornholm Basin. If the new deepwater has sufficiently high density, it will penetrate into the enclosed part of the Bornholm Basin, below the level of the deepest connection with the Gdańsk Basin (the so-called sill depth) through the Stolpe Channel (Słupsk Furrow). Otherwise it will be interleaved in the halocline above the sill depth from where it may escape the Bornholm Basin through the Stolpe Channel.

The oxygen conditions in the basin water, below the sill level, depend on the rates of supply and consumption of oxygen. Oxygen supply is essentially due to inflowing new deepwater while oxygen consumption primarily depends on decomposition of organic matter sinking down from the surface layer where it is produced. Since the salinity of the new deepwater during an inflow event increases as the inflow progresses (Stigebrandt, 1987a; Stigebrandt et al., 2015), only the later-coming part of very large inflows has high enough salinity to replace the residing deepwater. This means that with the actual rate of vertical mixing, it may take a couple of years before an initially very salty deepwater becomes exchanged. If the long-term oxygen consumption is larger than the oxygen supply, the basin may become exhausted in oxygen and hydrogen sulfide may be added to the water column before the next event of deepwater renewal brings in new oxygen as demonstrated by observations and model simulations in Stigebrandt et al. (2015).

The oxygen conditions are very important for the ecology and the water quality in a basin. Higher forms of life may have difficulties in getting enough of oxygen in hypoxic conditions ($O_2 < 2$ mg L^{-1}). In the Bornholm Basin the bottoms in the deepest part are episodically anoxic (no O_2) and azoic, i.e. they lack animals. This is a large disadvantage for e.g. cod that largely feed on benthic animals. For successful recruitment, cod is depending on the existence of oxic ($O_2 > 2$ mg L^{-1}) water with salinity >11 PSU (e.g. Stigebrandt et al., 2015; and references therein). When oxygen disappears (anoxic), red-ox reactions are reversed and, for instance, the bottom sediment starts to leak phosphorus (e.g. Stigebrandt and Kalén, 2013). This is eventually mixed into the surface layers which increase the biological production which leads to increased oxygen consumption in the deepwater implying expanding bottom areas with anoxic conditions which further increase the phosphorus leakage.

This paper describes and discusses the mixing and through-flow of new deepwater in the Bornholm Basin. The applied approach is to first describe the topography and the hydrographical properties. Thereafter the oceanographic processes influencing the mixing and the through-flow are described. The idea is that the conditions in the Bornholm Basin can be understood from knowledge of general principles and basin-specific factors. Finally some results from a vertical advection-diffusion circulation model are described.

2. Topography, sediment conditions and other external facts

In the central parts, the Bornholm Basin reaches about 100 m depth (Fig. 1). The deepest connection between the Bornholm Basin and the basins east and north of it is 59 m and it goes through the Stolpe Channel. East of the isle of Öland there is an about 46 m deep connection to the West Gotland Basin. Consequently the Bornholm Basin is closed beneath 59 m depth. The horizontal area at this depth equals about 14 150 km^2 and the volume of the closed part beneath this depth equals about 200 km^3 (e.g. Stigebrandt and Kalén, 2013). The vertical circulation of the Bornholm Basin can be thought of as taking place in a so-called diffusive filling-box where new deepwater enters essentially through the Bornholm Strait in the northwest and leaves through the Stolpe Channel in the southeast. During strong inflow events with highly elevated halocline in the Arkona Basin some deepwater may likely enter the Bornholm Basin also across the ~29 m deep sill southwest of Bornholm Island (Lass et al., 2001; Stigebrandt, 1987a).

A map of surficial sediment types shows that bottoms in the deeper parts of the Bornholm Basin usually are muddy and covered by soft material. However, in an area stretching southeastwards from the Bornholm Channel, east of Bornholm Island, bottoms are made up of sand and hard clay (Fig. 2). As further discussed in Section 4.3, it is obvious that the hard bottoms are swept clear of soft matter during occasional events with high-speed dense bottom currents, carrying new deepwater to the Bornholm Basin. Svikov and Sviridov (1994) constructed a map of occasional high bottom speeds based on the properties of the surface sediment.

The Bornholm Basin is the area with the lowest frequency of ice cover in the whole Baltic Sea. The hundred-year wind wave height has been estimated to be 15 m, see Ödalen and Stigebrandt (2013b) for additional information.

The studies referred to in the present paper have largely used hydrographical data from stations BY4 and BY5 in the Bornholm Basin (Fig. 1) but also from stations in neighboring basins. All data can be obtained from official databases. Additional data are available from HELCOM. Furthermore, Nord Stream collected data on salinity, temperature and currents with high vertical resolution in two locations using moored instruments. These and other data are available from Nord Stream (see https://www.nord-stream.com/environment/data-and-information-fund/dif/).

3. Hydrography

Salinity and oxygen concentrations of the water column in the Bornholm Basin are shown in Fig. 3 for the period

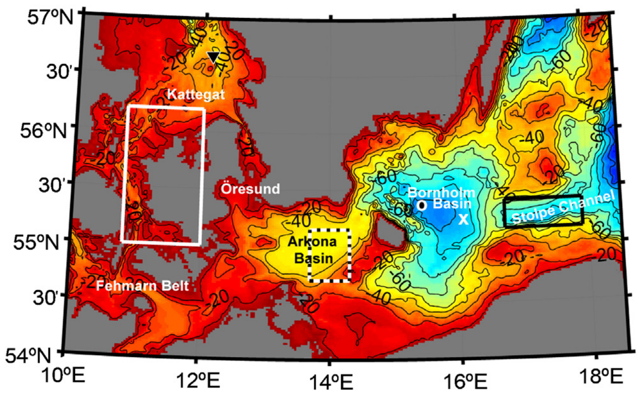

Figure 1 Map of the southern Baltic Sea area (hypsographic data from Seifert et al. (2001)). Reference data for the Arkona Basin originates from the area within the white and black dashed rectangle and reference data station BY4 and BY5 in the Bornholm Basin is marked by a white ○ and × respectively. 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 ▼ (from Stigebrandt et al., 2015).

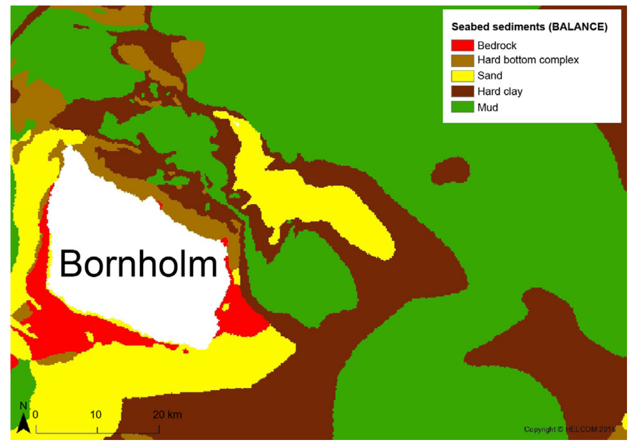

Figure 2 Sediment types in the Bornholm Basin and surrounding waters. Fine grained, muddy sediments (green) and sand (yellow) are dominant in the deep basin, with some areas of more course hard clay (brown). Courser sediments are also found in and around the sand bank (yellow) extending from Christiansø toward the deep basin (Map from HELCOM Map and Data Service). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

1958—2016. As can be seen there are large variations in time, in particular in the deepwater. Variability and long-term averages of salinity, temperature, density and oxygen concentration for the period 1970—2010 were investigated by Ödalen and Stigebrandt (2013a). The long term means and their standard deviations are given in Table 1.

It is interesting to note that the largest variability of temperature (variability equals StD squared) occurs in the surface layer due to the annual heating—cooling cycle while the largest variability of salinity and oxygen occurs around and below the sill depth (Table 1). The large variability below the sill depth is due to long periods of stagnation during which

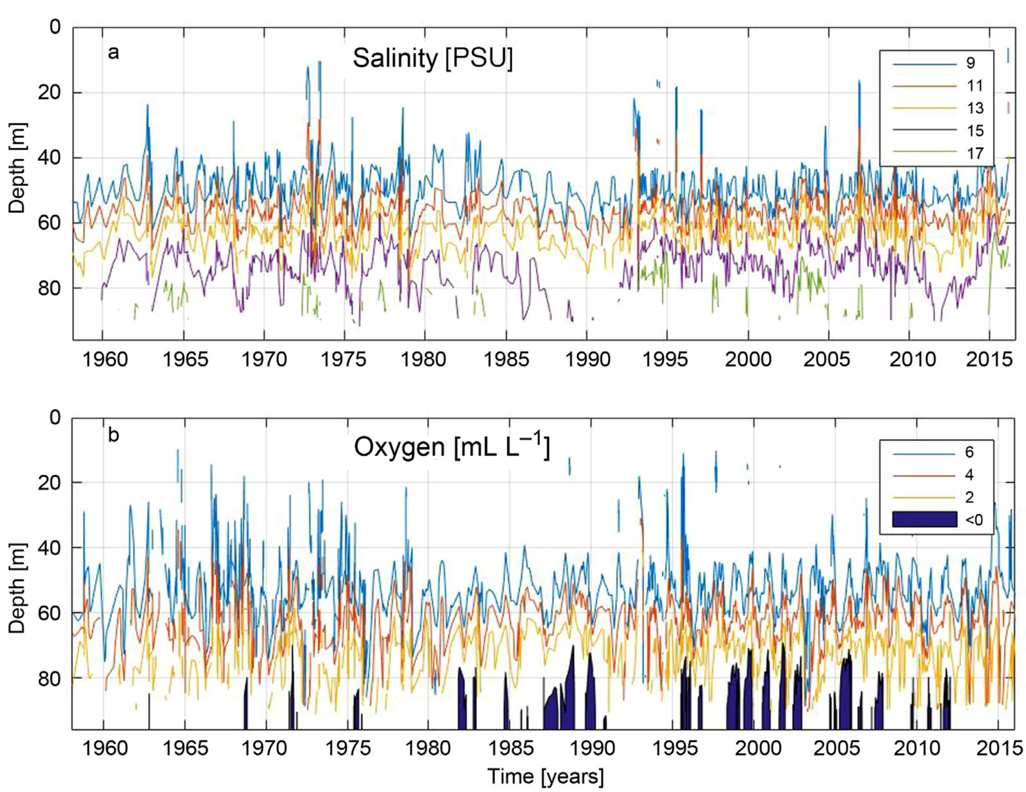

Figure 3 Isopleth diagram of (a) salinity [PSU] and (b) oxygen [mL O_2 L^{-1}] at BY5 from February 12, 1958 to July 22, 2016. Negative oxygen values in (b) denote the presence of anoxia with hydrogen sulfide in the water (data from SMHI).

Table 1 Long-term averages and standard deviations (StD) of temperature, salinity and oxygen (O_2) at selected standard observational depths from 1970 to 2010 inclusive (after Ödalen and Stigebrandt, 2013a).

Depth [m]	Temperature [°C]	±StD T [°C]	Salinity [PSU]	±StD S [PSU]	O_2 [g m^{-3}]	±StD O_2 [g m^{-3}]
0	9.6	5.6	7.5	0.3	11.4	1.5
10	9.2	5.4	7.5	0.3	11.4	1.5
20	8.2	4.6	7.5	0.3	11.3	1.5
30	6.4	3.4	7.6	0.3	11.4	1.4
40	5.0	2.2	7.8	0.4	11.3	1.3
50	5.0	2.0	8.9	1.3	9.8	2.1
60	6.4	2.4	12.1	1.7	6.7	2.3
70	6.8	1.8	14.7	1.1	3.6	2.2
80	6.9	1.7	16.0	0.9	1.9	2.4

salinity decreases steadily owing to vertical diffusion and oxygen decreases due to biological consumption. Seldom occurring water renewals furnish the basin below the sill depth with salt and oxygen. The variabilities of salinity, temperature and oxygen concentration due to the seasonal cycle and periods shorter and longer than this, respectively, are discussed below.

Following Stigebrandt (2012), the variability was partitioned in three period bands viz. periods longer than 1 year, the period 1 year exactly and periods shorter than 1 year (Ödalen and Stigebrandt, 2013a). The result is shown in Fig. 4 where the variability for the three different time-scales is shown. It can be seen that the period 1 year, i.e. the annual cycle, which often is called the seasonal cycle,

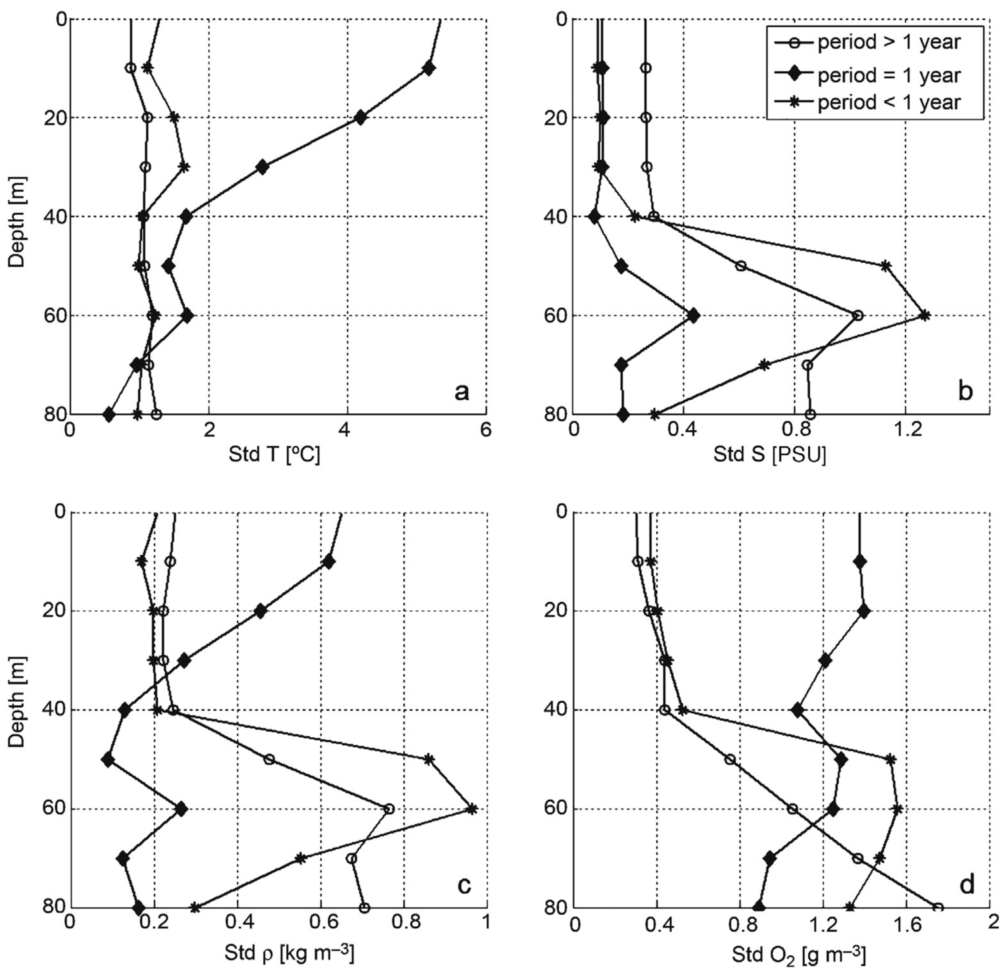

Figure 4 Standard deviation of (a) temperature [°C], (b) salinity [PSU], (c) density [kg m^{-3}] and (d) oxygen [g O_2 m^{-3}] partitioned into three period bands, viz. periods longer than 1 year, the period 1 year and periods shorter than 1 year, respectively (from Ödalen and Stigebrandt, 2013a).

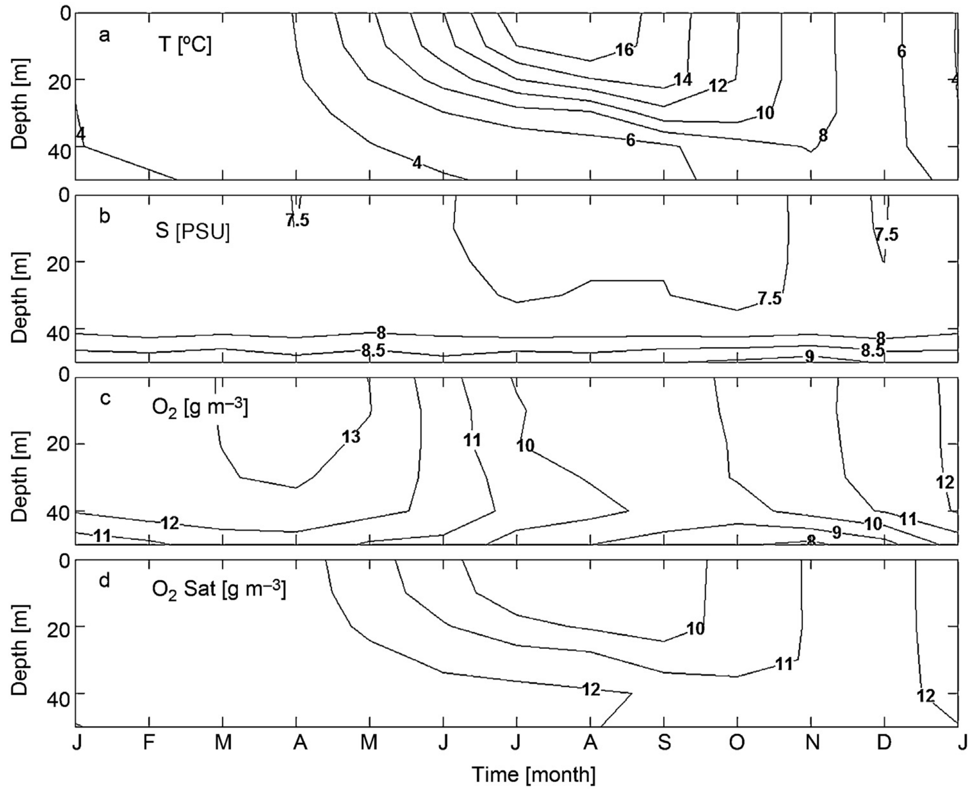

Figure 5 Annual cycles of (a) temperature [°C], (b) salinity [PSU], (c) oxygen concentration [g O_2 m^{-3}] and (d) oxygen saturation concentration [g O_2 m^{-3}] between 0 and 50 m for the period 1970—2010 (from Ödalen and Stigebrandt, 2013a).

dominates the variability of temperature, density and oxygen in the upper layer (about 40 m thick).

The mean seasonal cycles of temperature, salinity, oxygen concentration and oxygen saturation concentration between 0 and 50 m for the period 1970—2010 are shown in Fig. 5. The surface layer shows effects of seasonal local heating/cooling cycles, of precipitation/runoff and of oxygen production and use coupled to production and decomposition of organic matter. A comparison between oxygen concentration (Fig. 5c) and oxygen saturation concentration (Fig. 5d) shows that surface water is super-saturated in spring and under-saturated in autumn. There is also an effect of an annual cycle of new deepwater inflows, which shows up clearly in the salinity increase at 50 m depth in autumn. The oxygen concentration is simultaneously depressed (Fig. 5c), suggesting that the salinity increase at 50 m is due to upward advection of saltier and oxygen-depleted water.

The vertical salt stratification in the Bornholm Basin can be described roughly by two layers on top of each other that are separated by a 5—20 m thick halocline. The top of the halocline is usually at about 50 m depth but it is up-lifted in periods of larger inflows of new deepwater from the Arkona Basin. The halocline top deepens by outflow of halocline water through the Stolpe Channel and also by entrainment of halocline water into the surface layer during events with strong winds. Compared to the situation in the basins east and north of Bornholm Basin, the top of the halocline is elevated in the Bornholm Basin. The rather shallow sill of the Stolpe Channel (59 m) is responsible for the elevated halocline by blocking outflow from the density stratified basin water below the sill depth. However, the salinity of the lower layer, below the halocline, is quite variable on time-scales longer than 1 year due to relatively strong vertical mixing, which reduces the salinity, in combination with large long-time variability of the salinity of the inflowing new deepwater, which occasionally increases the salinity, as can be seen in Fig. 3a.

4. Currents

Currents are generated in several ways, which are briefly summarized below. The direct action of the atmosphere on the sea surface, in particular through the wind stress, generates quasi-steady and time dependent wind-drift currents in the surface layer. The response to the wind stress depends on the vertical stratification, in particular on the thickness of the well-mixed surface layer, and on the buoyancy fluxes through the sea surface caused by local heating/cooling and precipitation/evaporation. Convergence and divergence of drift currents in the surface layer at coasts and density fronts may generate horizontal pressure gradients and thereby horizontal circulation in a basin. It may also generate internal waves in the stratified part of the water column, beneath the well-mixed surface layer. Internal waves may also be generated by interaction between barotropic currents varying in time and bottom topography. The internal waves generated this way have the same frequencies as the barotropic currents generating them. In addition, currents may be generated by horizontal pressure gradients at the open vertical boundaries of a basin. In the Bornholm Basin such pressure gradients may be caused by e.g. inflow of new deepwater through the Bornholm Channel. Since the new deepwater has relatively high density, the currents carrying this water takes, initially, place along the seabed in so-called dense bottom currents. The water in the dense bottom current is mixed with ambient water, that is why the negative buoyancy decreases and eventually the dense bottom current loses its buoyancy and contact with the seabed. The water is then interleaved in the water column at the level where the density equals that of the current. This process sustains stratification above the sill depth in the Stolpe Channel, which drives currents out of, and circulation within, the Bornholm Basin. In this section various types of currents occurring in the Bornholm Basin are discussed together with examples of observations of currents.

4.1. Wind-forced drift currents and inertial currents in the surface layer

In deep non-stratified water, a steady wind in the Northern Hemisphere will transport surface water perpendicular to the right of the wind direction. The Coriolis force acting on the vertically integrated transport (the so-called Ekman transport) is balanced by the wind stress on the sea surface. If there are buoyancy fluxes through the sea surface, due to fluxes of heat and water, and if the water column is stably stratified, the dynamics of the wind drift becomes more complicated and the transport will also obtain a component parallel to the wind direction as shown by Nerheim and Stigebrandt (2006) who analyzed current data from moored rigs in the Baltic Proper. When the wind drift converges, either against coasts or at density fronts in the open sea, barotropic and baroclinic pressure fields are created due to sloping sea surface and density surfaces, respectively. These force large-scale barotropic and baroclinic currents. On large scales in the open ocean, wind drift due to steady wind systems, like the trade winds and the west wind belt, create permanent large-scale current systems. In the Baltic Sea, the winds are often shifting direction and speed and it seems that no permanent large-scale circulation is sustained by the wind drift.

The response to the unsteady component of the wind stress takes the form of inertial currents that manifest themselves as damped clockwise (Northern Hemisphere) circulation as shown theoretically by Ekman (1905). Observations of damped inertial currents from the Baltic Sea were presented by Gustafsson and Kullenberg (1935). The period of inertial currents equals $T_f = 12/\sin(\varphi)$ h where φ is the latitude; for e.g. $\varphi = 55°$N the formula gives $T_f \approx 14.6$ h. The inertial period is the upper limit for periods of free internal waves. Inertial currents in the surface layer may transfer energy to free internal waves (of slightly shorter period than T_f) in the stratified deepwater, see Section 4.2, and to internal waves of periods longer than T_f that are bound to move along topography like e.g. internal Kelvin waves. In the surface layer, the typical amplitude of inertial currents is 0.1 m s^{-1} and the damping time scale is 1 day. It was assumed that the damping of inertial currents in the surface layer of the Baltic Proper is due to vertical transfer of power to internal waves in the stratified water column beneath the well-mixed surface layer (Liljebladh and Stigebrandt, 2000). From observations obtained during the DIAMIX experiment, these authors estimated that the average vertical transfer of power from inertial currents in the surface layer to internal waves in the deepwater is about 0.3 mW m^{-2}. This power is believed to eventually be transferred to deepwater turbulence as further discussed in Section 5.

4.2. Internal waves

Free internal waves have frequencies in the interval between the inertial frequency ($\approx$0.00012 s^{-1}) and the buoyancy frequency N defined by $N^2 = g/\rho \cdot (d\rho/dz)$ (symbols are defined in connection to Eq. (1)) that typically is about 0.03 s^{-1} in the strongly stratified Bornholm Basin. In this frequency interval, the orbital motion of water particles caused by the internal waves goes from vertical at the buoyancy frequency to horizontal at the inertial frequency. The vertical structure of internal waves varies with the shape of the vertical stratification of the water column. If the vertical stratification is constant with depth, the oscillation may have evident vertical modes (similar to oscillations of e.g. a violin string). In other cases the stratification can be described by two-layers on top of each other and the orbital motions in the two layers are horizontal and in opposite directions. A common property of all types of internal waves is that the vertically integrated horizontal transport vanishes, i.e. there is no horizontal net transport by internal waves.

The speed c_i of internal waves in two-layer stratification, with density difference $\Delta\rho$ and layer thicknesses h and H, respectively, and with h much less than H, equals

$$c_i = (g'h)^{1/2}. \quad (1)$$

Here $g' = g\Delta\rho/\rho_0$, where g is the acceleration of gravity and ρ_0 a reference density. Example: with $\Delta\rho/\rho_0 = 0.01$, $h = 20$ m, and $g = 10$ m s^{-2} one obtains $c_i = 1.4$ m s^{-1}.

Oscillating barotropic currents flowing across sloping bottoms in stratified waters will generate baroclinic (internal) currents that get their energy from the barotropic currents. The most well-known example is the generation of internal tides by the flow of surface tides over ridges and sills in the bottom. Internal waves are not very stable why they lose energy to turbulence by different mechanisms (e.g. Stigebrandt, 1976, 2012). Most of the power driving the turbulence responsible for the mixing of the deepwater of the oceans (and fjords) comes from tides via internal tides generated at sloping bottoms.

In the Baltic, tides are quite small and therefore the energy transfer from tides to deepwater turbulence should be negligible. However, time-dependent wind-forced barotropic currents occur frequently and these may lose power to internal waves and deep water turbulence. This was investigated by Nohr and Gustafsson (2009) who estimated the mean energy transfer from time-dependent barotropic currents for the Baltic proper but they did not discuss the Bornholm Basin specifically.

4.3. Dense bottom currents

The density of new deepwater, emanating from Kattegat and the Belt Sea, is higher than the density of water above the halocline. For small bottom slopes, the speed c of dense bottom current increases with the bottom slope up to the so-called critical slope when the speed equals the speed of c_i of the internal wave in the actual vertical stratification, see e.g. Stigebrandt (1987b). However, if the bottom slope is supercritical, c will still be close to c_i because in this case hydrodynamic instabilities grow rapidly at the interface and cause energy losses and strong apparent flow resistance by entrainment of overlying water. The two-layer stratification applied in the example of internal wave speed, just below Eq. (1), should also be typical of dense bottom currents during major inflows. Dense bottom currents during major inflows should thus in extreme cases be expected to reach velocities up to about 2 m s^{-1}, depending on the layer thickness and the density difference. Piechura and Beszczyńska-Möller (2003) observed current speeds up to 0.75 m s^{-1} during an inflow they characterized as medium-sized.

The new deepwater that moves as a dense bottom current along the seabed entrains lighter water from above. It has been estimated that entrainment of ambient water into the dense bottom current in the Bornholm Basin increases the volume flow of the bottom current by about 1% for every meter it descends but in the Arkona Basin the corresponding figure is rather 2% (Stigebrandt et al., 2015). These authors suggest that the difference might be due to differences between the two basins with respect to the large-scale bottom topography where the dense bottoms currents run. Slightly simplified, the bottom currents run on a sloping plane in the Arkona Basin while they run in a valley, or canyon, in the Bornholm Basin, see also Stigebrandt et al. (2006).

An ambitious observational program to estimate the supply of new deepwater to the Bornholm Basin/Baltic Proper was performed in the Bornholm Channel during the period 1973—1977 (Walin, 1981) The flow through a fixed vertical cross-section was observed on 32 different days using gelatin pendulum current meters and CTD. Walin reported that the salinity of the new deepwater is typically 10—20 PSU and the speed 10—50 cm s^{-1}. A short time after the major inflow to the Baltic Sea in January 1993, the flow in several vertical cross-sections in both the Bornholm Channel and the Arkona Basin was observed using ship ADCP and CTD (Liljebladh and Stigebrandt, 1996). These authors reported transports of new deepwater in the interval 60 000—120 000 m^3 s^{-1} but did not explicitly report currents speeds.

4.4. Circulation due to inflow/outflow from the basin

In the Bornholm Basin, new deepwater entering from the Arkona Basin contributes to an upward motion above its level of interleaving. Much of the new deepwater is not dense enough to penetrate into the closed basin, below the sill depth of the Stolpe Channel, that is why the through-flow of this water occurs entirely above the sill depth. The outflow of halocline water takes place through the Stolpe Channel, and probably occasionally also through the shallower channel east of the island of Öland. In model computations by Stigebrandt et al. (2015), further discussed in Section 6, it was assumed that a geostrophic baroclinic current, driven by the horizontal pressure difference between the Bornholm Basin and Stolpe Channel, exits through the Stolpe Channel. However, comparison with hydrographical observations showed that the modeled pool of halocline water in the Bornholm Basin was drained too fast.

One reason for too fast draining of the modeled halocline water could be that just a fraction of the computed baroclinic transport exits through the Stolpe Channel while the residual is re-circulated within the Bornholm Basin. One would then expect that the outflow sustains a dome-structure of the halocline water in the Bornholm Basin, with the isopycnal surfaces sloping downwards from the central parts of the basin toward the periphery, implying a radial horizontal baroclinic pressure gradient in the basin sustaining an anticyclonic, clockwise, circulation of halocline water. ADCP observations performed by the Polish research vessel Oceania, reported by Bulczak et al. (2016), support the expectation of a clockwise circulation of halocline water in the Bornholm Basin. Rak (2016) documented effects on hydrography and currents of the major inflow in the period December 2014—January 2015.

5. Vertical mixing

Different mechanisms of vertical mixing are very important for the modification and through-flow of new deepwater in the Bornholm Basin. The horizontal mean rate of vertical mixing in the deepwater can be estimated using the budget method described by e.g. Stigebrandt and Kalén (2013) that is applicable under so-called stagnation periods, i.e. periods lacking water exchange by advection. During such periods, storage changes of conservative entities below a horizontal surface at a specified depth z are due to transport by vertical diffusion through that surface. The horizontally averaged value of the vertical diffusivity κ at the specified depth can be estimated using the following equation:

$$\kappa_{z=u} = \left(A\frac{\partial S}{\partial z}\right)^{-1}_{z=u}\int_{d}^{u}\frac{\partial S}{\partial t}A\,dz. \tag{2}$$

Here $A = A(z)$ equals the horizontal area of the basin at the depth z and $S(z,t)$ the horizontally averaged value of S at depth z and time t obtained from observations. The value of the first factor on the right hand side of Eq. (2) is the value at the upper level of integration $z = u$, which must be below the sill level. The greatest depth in the basin equals $z = d$. This equation may be used to compute κ at $n - 1$ levels if there are n levels of measurements of a conservative property like sea salt $S = S(z,t)$. Concentrations of non-conservative dissolved substances like oxygen and nutrients depend in addition on internal sources and sinks due to e.g. decomposition of organic compounds in the water column or the seabed. Table 2 shows results reported by Stigebrandt and Kalén (2013) who used virtually all available salinity observations from the hydrographical stations BY4 and BY5 for the period 1957—2011 for their computations.

For modeling purposes, Stigebrandt (1987b) parameterized κ using the stratification parameter N, defined in Section 4.2, in the following way

$$\kappa = a_0 N^{-1}. \tag{3}$$

Here a_0 is an empirical intensity factor (velocity squared) accounting for the horizontal mean mixing activity of turbulence. Using a vertical advection-diffusion filling-box model with an entraining dense bottom current carrying new

Table 2 Average and standard deviation of vertical diffusivity κ [m^2 s^{-1}] at depth z [m], and work W [W m^{-2}] against the buoyancy forces and a_0 [m^2 s^{-2}] below the depth z in the Bornholm Basin based on hydrographical data from BY4 and BY5. 'No κ' is the number of estimates of vertical diffusivity.

z [m]	κ [$\times 10^{-6}$ m^2 s^{-1}]	No κ [—]	W [$\times 10^{-4}$ W m^{-2}]	a_0 [$\times 10^{-7}$ m^2 s^{-2}]
65	2.5 ± 1.7	12	1.0 ± 0.5	1.1 ± 0.8
75	4.5 ± 3.6	35	0.7 ± 0.5	1.3 ± 0.9
85	8.0 ± 7.2	88	0.4 ± 0.4	1.4 ± 1.3

deepwater into the basin, Stigebrandt (1987b) estimated $a_0 = 2.0(\pm 0.7) \times 10^{-7}$ m^2 s^{-2} for the Baltic Proper. Later circulation models for the Baltic Proper (Gustafsson, 2003; Meier, 2001; Omstedt, 2011) use this parameterization with $a_0 = 1.5 \times 10^{-7}$, which incidentally is quite close to the values presented in Table 2 for the Bornholm Basin.

The total rate of work against the buoyancy forces P by mixing processes below the level $z = u$ in a basin is obtained by integrating the buoyancy flux, $b = \kappa N^2$, from the greatest depth $z = d$ to the depth $z = u$:

$$P = \int_d^u \rho_0 \kappa(z) N^2(z) A(z)\, dz. \tag{4}$$

To compare the power used by buoyancy fluxes in various basins, one may divide P by the area of the basin at the upper integration limit u. This gives the normalized power $W(u) = PA^{-1}(u)$ spent to buoyancy fluxes in the water column of the basin water beneath the depth u. The results for the Bornholm Basin obtained by Stigebrandt and Kalén (2013) are shown in Table 2. The estimated rate of work beneath 65 m is thus 100 μW m^{-2}. It should be underlined that if κ is estimated using the budget method, the value of P is only dependent on the hydrographical observations; no empirical coefficient is involved in the estimate. The error in P should thus only depend on the quality and the number of hydrographical observations used.

Only a minor fraction Rf of the supplied energy E is used for work against the buoyancy forces W. Thus $W = Rf \cdot E$. Most of the supplied turbulent energy is dissipated to heat $D = (1 - Rf)E$. With $Rf = 0.07$, as discussed below, the result from Stigebrandt and Kalén (2013) gives $E \approx 1.5$ mW m^{-2}. The horizontal and time mean dissipation D in the deepwater of the Bornholm Basin should then be about 1.4 mW m^{-2}. Since tides are very small in the Baltic, it should be obvious that the power E ultimately is derived essentially from the wind. Axell (1998) found that the annual variation in mixing is well correlated with the variation in wind stress. This result is supported by Stigebrandt and Kalén (2013) who found that the ratio between the powers W used against buoyancy forces in summer and winter, respectively, equals 0.65.

It is however not clarified by which mechanisms the wind energy is transferred to the deepwater. Liljebladh and Stigebrandt (2000) estimated that decaying inertial currents in the surface layer east of Gotland might deliver about 0.3 mW m^{-2} to deepwater turbulence, see Section 4.1. Nohr and Gustafsson (2009) found that oscillating barotropic currents may supply substantial power to deepwater turbulence via internal waves generated at sills and sloping bottoms, see Section 4.2.

The budget method accounts for the integrated action of all vertical mixing taking place in a stagnant basin but it does not tell when and where in the basin that mixing was carried out. Probably the first model of boundary mixing in a closed basin was presented by Stigebrandt (1976) who applied it to the inner part of the Oslo Fjord. A few years later, a tracer cloud experiment in the deepwater of the Oslo Fjord (Bjerkeng et al., 1978) showed that vertical mixing estimated from the evolution of the tracer cloud that had no contact with the side boundaries, was only about 10% of the vertical mixing estimated from the budget method applied on sea salt (S), which proves the dominance of boundary mixing (Stigebrandt, 1979). Boundary mixing versus interior mixing in the eastern Gotland Basin was one of the major research questions of the DIAMIX project, another was the overall energy budget of deepwater turbulence (Stigebrandt et al., 2002). Holtermann and Umlauf (2012) report results from a large-scale tracer experiment in the East Gotland Basin that verified the dominance of boundary mixing. Van der Lee and Umlauf (2011) estimated dissipation in the basin water of Bornholm Basin from microstructure profiles. It was an order of magnitude smaller than that estimated above from the estimate of W. This strongly suggests that the observed mixing was interior mixing, which should be of less importance for basin mixing, cf. also Section 6.

It follows from the dense bottom current model in Stigebrandt (1987b) that the efficiency Rf of turbulence generated by combined bottom and interfacial drag in dense bottom currents equals approximately 0.04 (Stigebrandt et al., 2006). This is rather low compared to e.g. the mixing efficiency in fjord basins where numerous estimates typically give an (overall) *system Rf* in the range of 0.06—0.07, see Stigebrandt (2012). This is less than half of the value 0.2 which is often used in oceanography. However, this value seems to be obtained from laboratory process studies. Arneborg (2002) explained that the *process Rf* includes both the work done against the buoyancy forces on the system level and short lived potential energy in turbulent patches. The latter dissipates when patches collapse. He concluded that the *system Rf* should equal about half the *process Rf*.

Possible hydrographical effects upon inflowing deepwater of a pipeline crossing the route of the dense bottom current carrying new deepwater was investigated by Borenäs and Stigebrandt (2009). Referring to Stigebrandt et al. (2006), they assumed that the total dissipation of the dense bottom current, i.e. the dissipation integrated from the entrance sills in Fehmarn Belt and Öresund to Słupsk Furrow, is essentially determined by the potential energy of the dense water when passing the entrance sills. They estimated that a pipeline that rises 1.5 m above the seabed is capable of dissipating up to 0.5% of the total potential energy of the dense bottom current, depending on the speed of the dense bottom current in the crossing section. The mixing of the new deepwater might increase if the mixing efficiency Rf_{pipe} of pipeline-generated turbulence is greater than 0.04, which is the efficiency of turbulence generated by combined bottom and interfacial friction as mentioned above. It is not an easy task to estimate the system efficiency of mixing induced by pipes using in situ observations. Ingenious methods must probably be invented.

6. Modeling of mixing and through-flow in the Bornholm Basin

A vertical advection-diffusion circulation model was applied to the closed, lower parts of the Bornholm Basin (Stigebrandt and Kalén, 2013). Empirical values of the vertical diffusion were applied using the diffusivity described by Eq. (3), with a_0 changing with depth according to Table 2. The model was run for a year-long stagnation period in the basin water that started with high salinity. Fig. 6 shows the development of the salinity as observed (upper) and according to the model run (lower). There is a large oscillation in observed salinity in November 2003 but there are no signs of water exchange, implying that the oscillation might be real and caused by large horizontal water displacements within the basin or false and due to erroneous observational data. The salinity computed by the model declines at about the same rate as observed showing that the vertical diffusivity, derived from historic salinity data, describes the vertical diffusion quite well during the modeled period.

To be able to compute the water exchange of the Bornholm Basin during even longer periods, the model had to be supplemented with a model that computes the inflow of new deepwater from the Arkona Basin and the outflow of water through the Stolpe Channel. A simple mechanistic model computing inflow from and outflow to the entrance area, i.e. the Belt Sea and Kattegat, was constructed by Stigebrandt et al. (2015). Flow rates were estimated from volume changes of the Baltic Sea inside the entrance area corrected for volume changes due to the freshwater supply. In practice they used 5 days moving averages of the sea level in Stockholm and the monthly supply of freshwater inside the entrance area. The surface water flowing from the Arkona Basin to the entrance area has low salinity and sustains a usually several meters thick fictive freshwater layer in the entrance area. In reality the freshwater is mixed into the surface layer. Changes of the freshwater layer thickness were computed in response to inflows and outflows, for details see Stigebrandt et al. (2015). The presence of a fresher surface layer together with relatively shallow sills, restricts (partially blocks) the inflow of saltier sea water from the entrance area to the Arkona Basin. As long as there is much freshwater in the entrance area surface layer, the salinity of inflowing new deepwater will be low. However, during events with inflow to the Baltic Sea, the salinity of inflowing water increases with the duration of the inflow (Stigebrandt, 1987a) because the thickness of the freshwater layer in the entrance area diminishes during such episodes due to export to both the Arkona Basin and Skagerrak.

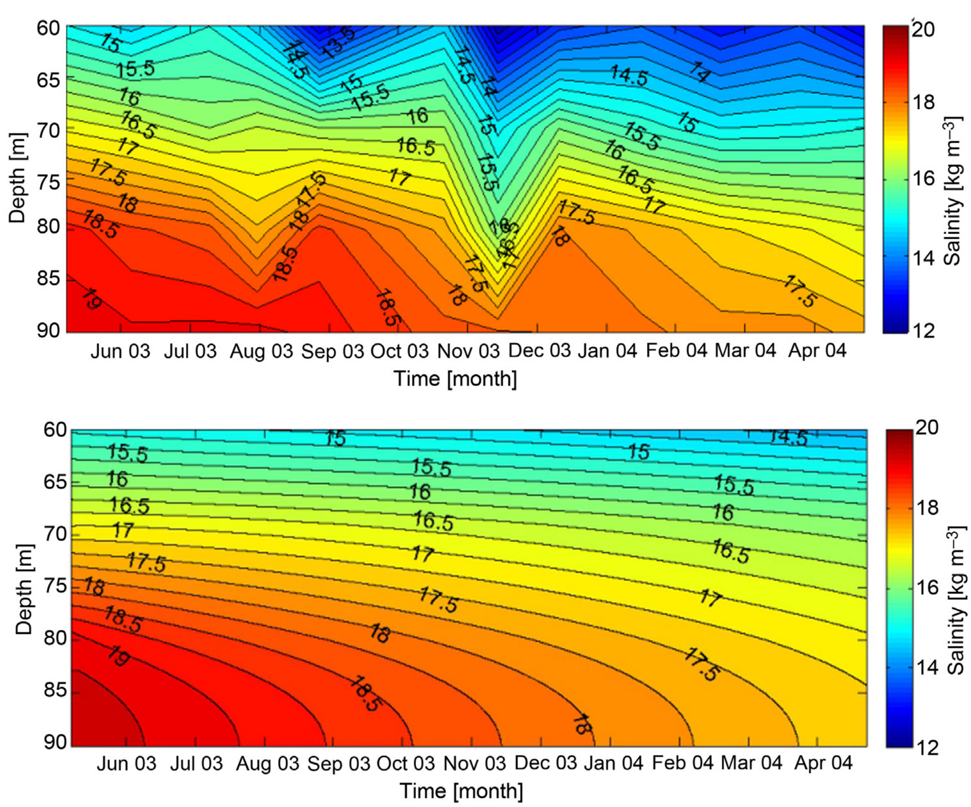

Figure 6 Observed salinity [PSU] at BY5 (upper panel) and modeled salinity [PSU] (lower panel) in the basin water of the Bornholm Basin during the period 7 May 2003 to 21 April 2004 (from Stigebrandt and Kalén, 2013).

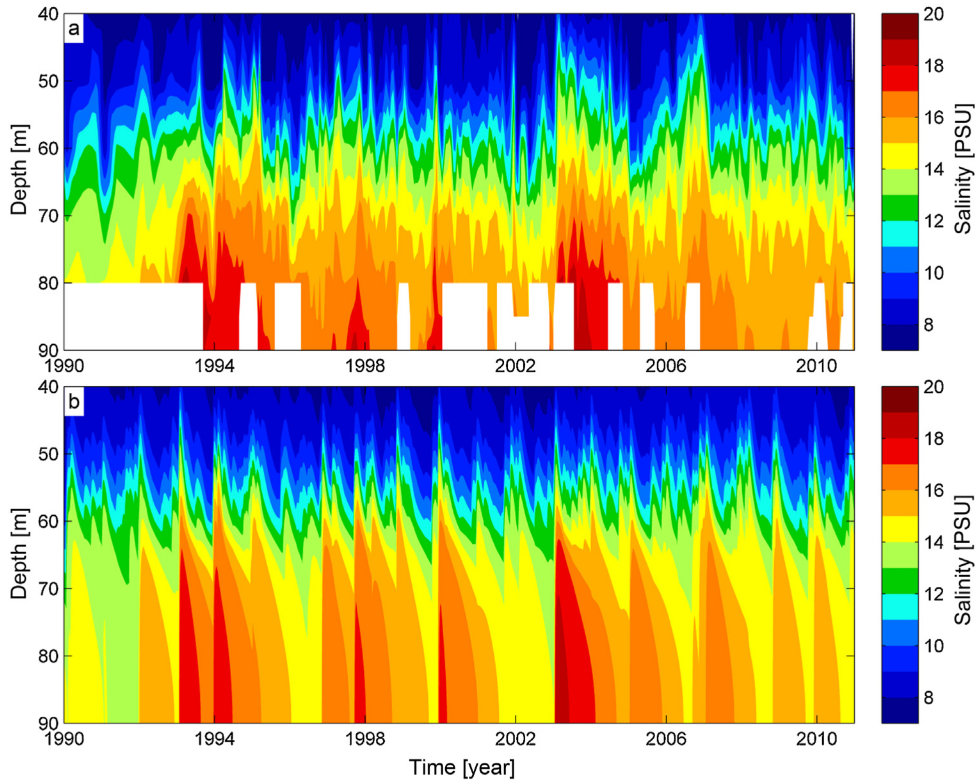

Figure 7 The observed salinity [PSU] from BY4 (a) and the modeled salinity [PSU] (b). White areas in (a) indicate lack of data (from Stigebrandt et al., 2015).

The vertical advection-diffusion model supplemented with the through-flow model was run for the lower part of the Bornholm Basin for the period 1990—2011 (21 years) (Stigebrandt et al., 2015). Fig. 7 shows the observed salinity (a) and the modeled salinity (b). As can be seen, the model quite well describes the evolution of salinity in the lower part of the basin. The modeled salinity shows seemingly abrupt increases of salinity due to inflows of new deepwater. The observed salinity increases are more diffuse because the observational data grid is coarse and because horizontal displacements within the basin (cf. Fig. 7a) as well as internal waves create a large variability that influences the observations. A closer inspection shows that the heightened salinity above 60 m depth due to halocline uplift after larger inflows, such as the one in early 2003, can persist for up to 1 year. In the model, the effect of such uplifts seems to decay more quickly, suggesting that the baroclinic outflow through the Stolpe Channel might be too fast in the model as discussed in Section 4.4. In the model by Stigebrandt et al. (2015) the upper part of the water column changes as a result of data

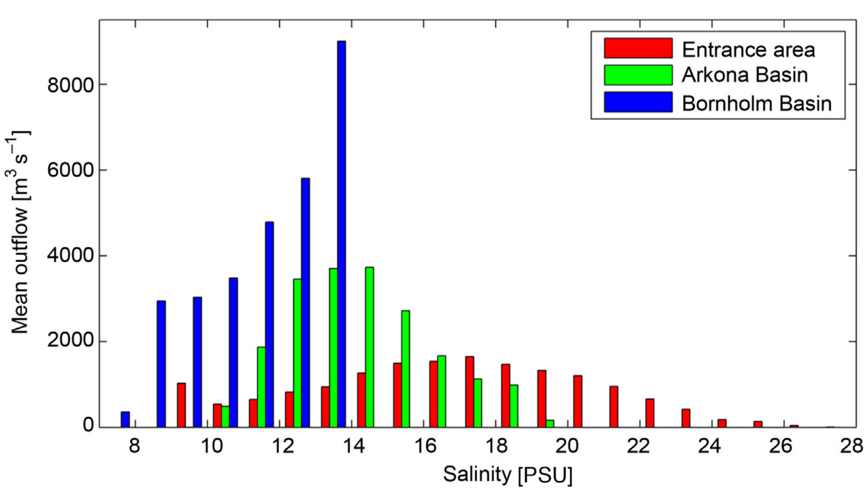

Figure 8 Distribution of average volume flow among different salinities for the modeled deepwater flows from the entrance area, the Arkona Basin and the Bornholm Basin (from Stigebrandt et al., 2015).

assimilation. Thus, there is no model that computes e.g. the dynamics of the mixed surface layer because this was considered of less importance for the evolution of the stratification in the lower part of Bornholm Basin. In the simulations, the temperature field shows that the model misses some instances of strong mixing in the surface layer (see Stigebrandt et al., 2015).

The salinity of the deepwater in the Baltic Proper east of the Słupsk Furrow is only about 1/3rd of the salinity of Kattegat deepwater. The very large decrease of the salinity of new deepwater is taking place in the Arkona and Bornholm Basins. Due to efficient mixing between inflowing water and residing Baltic Sea surface water, the salinity range of inflowing new deepwater decreases the further into the Baltic Sea the deepwater intrudes. Accordingly, the high salinity end of new deepwater from the entrance area has salinities in the range of 20—28 PSU; the high salinity end of new deepwater from the Arkona Basin is in the range of 15—20 while the highest salinity is less than 15 from the Bornholm Basin (Fig. 8). By looking at e.g. the upper panel of Fig. 6 it can be understood that the mixing of residing water in the Bornholm Basin is paramount to reduce the high salinity end of new deepwater imported from the Arkona Basin. The salinity range of new deepwater is a very important factor for the vertical stratification and the length of stagnation periods in the Baltic Proper as discussed in Stigebrandt (1987b). Increased mixing in the Bornholm Basin by pipelines and other man-made devices should thus contribute to shorter stagnation periods, and thereby decrease the occurrence of anoxia in the Baltic Proper, see also Stigebrandt et al. (2015).

7. Concluding remarks

Scientific knowledge about the hydrography of the Bornholm Basin has increased drastically during the last decade. One reason for this is the increased interest in using the Bornholm Basin for various purposes that has required analyses to quantify and understand the dynamics of this sea. Thus, investigations of environmental effects of various possible man-made operations and building of physical constructions have forced interest into this particular sea which has led to increased scientific knowledge. After thorough investigations of possible environmental effects, the company Nord Stream has now laid down two gas pipelines on the seabed crossing the Bornholm Basin in the SSW—NNE direction. There have been also theoretical investigations of possible man-made oxygenation of the nowadays often anoxic deepwater and its effects, in order to reduce the leakage of phosphorus from anoxic bottoms. Oxygenation would improve the water quality at greater depths so that the now azoic bottoms may be colonized which would imply more food for the cod, for instance. Improved pelagic water quality would also lead to better conditions for cod recruitment.

Acknowledgement

Josefina Almén adapted the figures borrowed from other publications to *Oceanologia* standards and drew Fig. 3.

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