Deep-water oxygen conditions in the Bothnian Sea

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The hydrography of the deep waters of the Bothnian Sea (BS) underwent considerable changes during the most recent two decades. Most importantly, slightly worsened deep-water oxygen (O2) conditions followed a gradual increase in the water-column density gradient. Up to the present time, the lowest measured O2 concentrations fell within the range 3.5–4.5 ml l⁻¹, 1.4 ml l⁻¹ being used in this work as the upper limit for hypoxia. I investigated the long-term deep-water characteristics of the BS to determine how possible it is for an hypoxic event to occur there in the near future. It appears that there is only a remote probability of this, unless profound changes take place in the hydrographic regime of the BS and/or in the anthropogenic load to the BS. The key factor behind this conclusion is the hydrodynamic regime of the Åland Sea and the ridge formations around it. This area acts as a buffering/filtration mechanism that prevents the saline and O2-poor deep waters of the Northern Baltic Proper from entering the BS without considerable mixing and dilution. A discussion is presented of the consequences of this conclusion, taking into account the past adverse development in the Gulf of Finland, with its internal loading of phosphorus and its accelerated eutrophication. With respect to the Finnish coastal area of the BS, the stable O2 conditions and the favourable hydrographic setup in the near-coastal area suggest that any O2 problems will most probably be concentrated in the inner-archipelago areas.

Introduction

The spreading of hypoxia in coastal and estuarine marine environments is currently a ubiquitous problem. Coastal systems that experience episodic hypoxic events or permanent hypoxia have greatly increased in numbers (Selman et al. 2008). The conspicuous factor behind this adverse development is eutrophication, which boosts the flow of organic matter to the benthic system, ultimately degrading the oxygen (O2) conditions there. This development has also taken place in the Baltic Sea.

However, eutrophication is not solely to blame for the current poor O2 status of the Baltic Sea. The Baltic Sea has an intrinsic tendency to experience O2 problems due to its natural vertical salinity stratification, and hence, density gradients that effectively restrict the vertical transport of O2. These salinity anomaly layers, called haloclines, have played a pivotal role in the spreading of a hypoxic environment in the Baltic Sea; indeed, O2 problems have persisted there for as long as the present geographical coverage of the halocline system has prevailed, that is, from the commencement of the Littorina Sea stage at approximately eight millennia BP. (Zillen et al. 2008). From this perspective, anthropogenic eutrophication has had little time to leave its fingerprints in the O2 regime of the
Baltic Sea. Today, hypoxic environments consistently cover large offshore areas in the Baltic Proper, while the western parts of the Gulf of Finland (GOF) can be classified as seasonally hypoxic (HELCOM 2009b); furthermore, the hypoxic environment has recently been spreading in the coastal areas (Conley et al. 2011).

The offshore Bothnian Sea (BS), on the other hand, has not been further eutrophied in terms of dissolved inorganic nutrients since the early 1990s (Fleming-Lehtinen et al. 2008, HELCOM 2009b). Even so, alarming signals have emerged during this period, e.g., the chlorophyll \(a\) levels and cyanobacterial biomasses have steadily increased there since the mid-1990s (Fleming-Lehtinen et al. 2008, Jaanus et al. 2011). The Finnish coastal area of the BS suffers from the allochthonous nutrient load, and the inner coastal area experiences episodic hypoxic events (Lundberg et al. 2009). Lundberg and co-workers stated that “… even if eutrophication in the Gulf of Bothnia is not serious, the increasing trends in the nutrient levels should be seen as warning signs for the future”. A closer look into the BS system is indeed topical.

Eutrophication is a process that, if given a chance, follows the cause-and-effect cascade of increased allochthonous load of nutrients \(\rightarrow\) increased autochthonous load of organic matter \(\rightarrow\) anoxia \(\rightarrow\) increased autochthonous load of nutrients (the internal loading of phosphorus). There exists a strong feedback path from the end-product, the internally-loaded phosphorus (P), adding the autochthonous component to eutrophication. This feedback mechanism has the potential to strongly accelerate the eutrophication process. I wanted to clarify whether this cascade, which has already commenced and is still ongoing in the GOF, could be a potentially realistic scheme for the BS. In this article, I present a long-term hydrographic dataset collected at representative monitoring stations in the BS. I address the following questions: (1) how have the near-bottom \(O_2\) conditions in the BS recently developed? (2) what are the external factors that control the \(O_2\) regime of the BS? and (3) what can we expect to happen in the near future?

The Bothnian Sea

The BS is the southern basin of the Gulf of Bothnia, which is the semi-enclosed northern extension of the Baltic Sea. In the Baltic Sea, the BS is the largest basin after the Gotland Basin, with an area of about one-sixth of the Baltic Sea area and a water volume of about one-fifth of the Baltic Sea volume (Leppäranta and Myrberg 2009). The average depth of the BS is 66 m as compared with 54 m for the Baltic Sea; the BS contains a central plane with large areas deeper than 100, or even 200 m (Fig. 1).

The southern Åland Sill, the southern Quark, and the shallow and scattered Archipelago Sea isolate the BS effectively from the Northern Baltic Proper (NBP). Consequently, the hydrographical forcing of the NBP on the BS is limited, albeit constantly present. The northern Quark, which topographically separates the BS from the Bothnian Bay, does not restrict the
southerly flow of fresher water to the BS. This flow and the direct river runoff into the BS together create a pronounced fresh water impact on the BS. Together these topographic features result in hydrological properties for the BS that are markedly different from those in the NBP. The pronounced river runoff into the Gulf of Bothnia leads to a short water residence time on the Baltic Sea scale, that is, 5 to 6.5 years (Myrberg and Andrejev 2006).

The BS freezes over, at least partly, every winter. The probability of freezing is highest in the Finnish coastal area and the areas next to the northern Quark that freeze every winter, becoming smaller in the southern half of the offshore area; this remains ice-free during mild winters, i.e., about one-third of the winters (Feistel et al. 2008).

Every summer, a thermocline develops in the BS and this acts as a barrier for vertical mixing at a depth ranging from 10 to 20 m depending on the year and the season. It separates the surface water mass with temperatures of 16–18 °C at its maximum from the deep waters at 2–5 °C.

The BS has a cyclonic surface water circulation, which leads to a less saline surface layer on its deeper western side, and a more saline surface layer on its shallower eastern side. This general picture is however strongly subject to wind forcing, and is thus intermittent (Håkansson et al. 1996). The surface salinity of the BS ranges from about 4.5 in the north to about 5.5 PSU in the south. The input of fresher water from the Bothnian Bay produces marked variation in this typical value, especially in the northwestern part of the BS. Salinity increases quite steadily with depth, somewhat faster in the halocline located typically at a depth of 50–75 m. However, no distinct halocline like that in the NBP is found. The salinity near the sea floor is typically about 1 PSU higher than that at the surface. This faint halocline does not create as strong a water-column density gradient as that in the NBP, for example, and consequently the potential of the gradient to restrict vertical mixing varies considerably (Kullenberg 1981). The stratification in the BS is thus governed to a much greater extent by temperature variations than is typical for the southern basins of the Baltic Sea.

Material and methods

Stations

The main stations used in this study (SR5, US2, and US5B) are located in the central plain of the BS at water depths > 100 m (Fig. 1). These stations are situated in the sediment-accumulation areas that are the first to experience O₂ problems. In this deep area, a consistent albeit faint halocline represents a quasi-stable hydrographic barrier for the vertical flow of O₂ to deeper waters, emphasizing the role of the O₂ supply via horizontal advection (Table 1). US2 is located in the northwestern part of the BS, and is hence subject to the impact of the southward surface current of less saline water from the Bothnian Bay. SR5 is located in the southern part of the central plain, and its deep-water hydrographical characteristics are strongly influenced by the hydrographical forcing of the Åland Sea (ÅS).

The near-coastal stations of the BS (SR9, MS11, and US7) are located in the Finnish outer coastal area of the BS. The water depth is < 50 m, and the virtually non-existing vertical density gradient in winter indicates that the halocline does not extend to these shallow areas (Table 1).

The station representative for the ÅS, F64, is located in the northern part of the ÅS in a trench that extends over most of the basin. The sea area is subject to efficient vertical mixing and no hypoxia is observed there, even in deep waters (down to about 300 m).

The station representative for the NBP, LL19, is situated in the midst of the northern Gotland Basin. With a water depth of > 150 m, the sea floor there is typically anoxic.

The station representative for the GOF, LL7, is located in the western part of the gulf in an area that is characterized by episodic and unpredictable anoxic events due to the hydrographical forcing of the NBP; hypoxia is met with there frequently.

Data sources

The marine dataset was compiled within the Baltic monitoring programme (BMP), and later,
Table 1. Stations included in the study, sampling scheme, water-column density ranges ($\Delta\sigma_t$), and near-bottom $O_2$ concentrations.

<table>
<thead>
<tr>
<th>Station</th>
<th>Data range</th>
<th>Number of visits</th>
<th>Sampling scheme</th>
<th>Depth (m)</th>
<th>Mean $\Delta\sigma_t$ (kg m$^{-3}$)</th>
<th>Mean near-bottom $O_2$ (ml l$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>all data</td>
<td>winter data (Jan–Mar)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>years of monitoring</td>
<td>number of years with &lt; 3 visits</td>
<td>number of years with no visits</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Deep stations</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>SR5</td>
<td>1962–2010</td>
<td>245</td>
<td>49</td>
<td>10</td>
<td>2</td>
<td>130</td>
</tr>
<tr>
<td>US2</td>
<td>1962–2011</td>
<td>72</td>
<td>50</td>
<td>43</td>
<td>10</td>
<td>209</td>
</tr>
<tr>
<td>US5B</td>
<td>1966–2011</td>
<td>132</td>
<td>46</td>
<td>17</td>
<td>3</td>
<td>222</td>
</tr>
<tr>
<td>Near-coastal stations</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>SR9</td>
<td>1972–2009</td>
<td>26</td>
<td>38</td>
<td>37</td>
<td>19</td>
<td>41</td>
</tr>
<tr>
<td>MS11</td>
<td>1990–2008</td>
<td>14</td>
<td>19</td>
<td>18</td>
<td>12</td>
<td>36</td>
</tr>
<tr>
<td>US7</td>
<td>1965–2010</td>
<td>48</td>
<td>46</td>
<td>43</td>
<td>18</td>
<td>25</td>
</tr>
<tr>
<td>Stations south of the BS</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>F64</td>
<td>1963–2011</td>
<td>109</td>
<td>49</td>
<td>28</td>
<td>6</td>
<td>286</td>
</tr>
<tr>
<td>LL19</td>
<td>1965–2011</td>
<td>94</td>
<td>47</td>
<td>31</td>
<td>6</td>
<td>169</td>
</tr>
<tr>
<td>Station in the GOF</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>LL7</td>
<td>1962–2011</td>
<td>359</td>
<td>50</td>
<td>12</td>
<td>1</td>
<td>102</td>
</tr>
</tbody>
</table>

Note: Anoxic conditions are indicated by an asterisk (*) in the last column.
the HELCOM/COMBINE programme, carried out by the Finnish Institute of Marine Research (1962–2008) and later by the Marine Research Centre of the Finnish Environment Institute (2009–2010). The sampling scheme at the stations is shown in Table 1. Typically, all existing data sources were used: numerical data collected with CTD sensors, and analytical data determined both from CTD’s rosette samples and from discrete water samples. The riverine dataset was compiled within the HELCOM/PLC programme.

With respect to the water column, this study concentrates on the surface layer (above any density gradient) and the deep layer (below any density gradient). The combined data from the depths of 0 and 5 m describe the surface layer, which is denoted “SURFACE”. The deep layer is described by the data collected as follows: SR5 ≥ 100 m, US2 ≥ 175 m, US5B ≥ 175 m up to 1988 and ≥ 200 m since 1989, SR9 ≥ 35 m, MS11 ≥ 30 m, US7 ≥ 20 m, F64 ≥ 275 m, LL19 ≥ 150 m, and LL7 ≥ 70 m up to 2002 and ≥ 90 m since 2003. The deep layer is denoted “DEEP”.

Parameters

For salinity, the practical salinity scale (PSS78) was employed, and the practical salinity unit (PSU) was used throughout the study (UNESCO 1980). Since 1962, the salinity determined from discrete water samples was measured with various Autosal bench salinometers calibrated against IAPSO standard seawater (Grasshoff et al. 1999). Since 1978, the CTD-based salinity was measured with conductivity cells of various CTD systems whose output was checked against wet analytics on a constant basis.

Since 1962, the temperature (°C) at the sampling depth of the discrete water samples was measured with thermometers installed on the sampling bottles. Since 1978, the CTD-based temperature was measured with the temperature probes of various CTD systems that were calibrated by the manufacturer on a constant basis.

In the shallow waters of the Baltic Sea, these two parameters determine the density \( \sigma_t (= \text{density} - 1000 \text{ kg m}^{-3}) \) (Millero and Poisson 1981). The water-column density range was defined as \( \Delta \sigma_t = \sigma_t(\text{DEEP}) - \sigma_t(\text{SURFACE}) \).

Since 1962, the O\(_2\) concentration (ml l\(^{-1}\)) was analyzed employing the Winkler technique according to Grasshoff et al. (1999) using either visual, or later, potentiometric endpoint recognition. Since 1999, the CTD-based O\(_2\) was measured by Clark-type electrodes of various CTD systems that were calibrated by the manufacturer on a constant basis, and whose output was checked against wet O\(_2\) analytics at every station. The upper limit for hypoxia was defined according to Conley et al. (2009) who suggested a value of 2 mg O\(_2\) l\(^{-1}\) (equivalent to \( \approx 1.4 \text{ ml O}_2 \) l\(^{-1}\)). The O\(_2\) saturation (%) was calculated from the O\(_2\) concentration and temperature.

The concentration of the dissolved inorganic phosphorus (PO\(_4\)-P, \( \mu \text{mol l}^{-1} \)) was determined colorimetrically, first using manual methods employing the photometric determination of the substance (Grasshoff et al. 1999), and later using the FIA approach with a SKALAR 5101 segmented flow analyzer (1993 to 1999) and a Lachat QuickChem 8000 flow injection analyzer (1999 to 2011). The outputs of the instruments were internally validated against each other. Currently, the measurement accuracies of PO\(_4\)-P, salinity and dissolved O\(_2\) are \( \pm 22\% \), \( \pm 0.3\% \) and \( \pm 7\% \), respectively.

Quality assurance

For salinity and O\(_2\), the chemical foundations of the measuring techniques as well as the basic technical principles of the instruments did not change during the data collection period presented in this article. For PO\(_4\)-P, the change from the manual procedure to the FIA technique did not change the optical setup, as the chemical foundation remained the same. Thus, with regard to these parameters, the only plausible source of error related to quality control stems from the inevitable changes in the instrumentation. This source of error has been minimized with the instrument validation and inter-comparison procedures.

All the methods above employed by the Finnish Environment Institute and the Finnish Meteorological Institute are currently internally accredited by the Finnish Accreditation Service (FINAS) following the SFS-EN ISO/IEC 17025 standard.
Results

DEEP hydrography in the BS

The DEEP salinity in the BS underwent prominent changes during the past five decades (Fig. 2). Various phases emerged: (i) a phase of increasing salinity in the 1960s and 1970s ending up with the peak in the late 1970s, (ii) a phase of decreasing salinity throughout the 1980s ending up with the low-point in the early 1990s, and (iii) a phase of increasing salinity ever since. This pattern also reflects the DEEP density because, in the BS, density variation is almost solely governed by salinity variation, except in the surface layer in summer (Table 2A). The DEEP temperature does not present such a distinct fluctuating pattern but showed a consistent, albeit faint, increase during the studied period. Any temporal trend in the DEEP $O_2$ saturation is somewhat obscured by the intrinsic seasonal variation linked to $O_2$ dynamics. Nevertheless, on the large scale, the DEEP $O_2$ saturation decreased whenever the DEEP density increased.
Table 2. (A) Linear dependence of water density ($\sigma$) on water temperature and salinity at the offshore stations in the BS since the early 1960s. (B) Linear dependences between the DEEP densities ($\sigma$) at SR5, F64, and LL19. Annual means are used since the early 1960s. CTD are the data from the CTD probes and/or rosette samples, DIS are the data from discrete water samples, for SURFACE and DEEP see the exact depth distributions in Material and methods. (C) Linear regressions describing the decrease in the O$_2$ concentration (ml l$^{-1}$) as a function of time from 1990 to 2011. Target year is the extrapolated year when the O$_2$ concentration reaches the upper limit of hypoxia ($\approx$ 1.4 ml l$^{-1}$). Only data collected from the discrete samples taken ≤ 5 m above the seafloor were included. In addition to being a consequence of natural variation in O$_2$ concentrations, the low $r^2$ values resulted from $a \approx 0$ in temporal O$_2$ trends.

<table>
<thead>
<tr>
<th>A</th>
<th>Data</th>
<th>Time</th>
<th>Salinity $r^2 (F, n, p)$</th>
<th>Temperature $r^2 (F, n, p)$</th>
</tr>
</thead>
<tbody>
<tr>
<td>SURFACE</td>
<td>CTD, DIS</td>
<td>Jul–Sep</td>
<td>0.413 (173, 250, &lt; 0.001)</td>
<td>0.677 (522, 250, &lt; 0.001)</td>
</tr>
<tr>
<td>DEEP</td>
<td>CTD, DIS</td>
<td>Jul–Sep</td>
<td>0.999 (182826, 174, &lt; 0.001)</td>
<td>0.011 (19, 172, 0.17)</td>
</tr>
<tr>
<td>SURFACE</td>
<td>CTD, DIS</td>
<td>Jan–Mar</td>
<td>0.987 (13491, 179, &lt; 0.001)</td>
<td>0.067 (12.8, 179, &lt; 0.001)</td>
</tr>
<tr>
<td>DEEP</td>
<td>CTD, DIS</td>
<td>Jan–Mar</td>
<td>0.998 (59274, 149, &lt; 0.001)</td>
<td>0.001 (0.09, 148, 0.76)</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>B</th>
<th>Data</th>
<th>Time</th>
<th>SR5 $r^2 (F, n, p)$</th>
<th>F64 $r^2 (F, n, p)$</th>
</tr>
</thead>
<tbody>
<tr>
<td>F64</td>
<td>CTD, DIS</td>
<td>Jan–Mar</td>
<td>0.854 (187, 33, &lt; 0.001)</td>
<td>–</td>
</tr>
<tr>
<td>LL19</td>
<td>CTD, DIS</td>
<td>Jan–Mar</td>
<td>0.420 (23.8, 34, &lt; 0.001)</td>
<td>0.583 (41.9, 31, &lt; 0.001)</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>C</th>
<th>Equation ($O_2$ conc. = $a \times$ year + $b$)</th>
<th>Target year</th>
<th>$r^2 (F, n, p)$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Case 1</td>
<td>$O_2$ conc. = −0.0714 × year + 149</td>
<td>2062</td>
<td>0.175 (38.3, 176, &lt; 0.001)</td>
</tr>
<tr>
<td>Case 2</td>
<td>$O_2$ conc. = −0.0411 × year + 87.0</td>
<td>2082</td>
<td>0.177 (9.80, 41, &lt; 0.01)</td>
</tr>
<tr>
<td>Case 3</td>
<td>$O_2$ conc. = −0.0359 × year + 75.8</td>
<td>2071</td>
<td>0.353 (4.27, 6, 0.094)</td>
</tr>
</tbody>
</table>
Hydrographic variation between DEEP and SURFACE in the BS

The comparison between the temporal salinity patterns in the DEEP and the SURFACE makes it obvious that the water-column salinity gradient has become larger since the low-point of the DEEP salinity in the early 1990s (Fig. 3). The dataset is not long enough to reveal any possible long-term fluctuations. In the shorter time interval, however: (i) the gradient in the 1970s was about as large as that in the 2000s, (ii) the gradient became smaller from the late 1970s up to the low-point, mainly because of the decrease in the DEEP salinity, and (iii) the gradient has gradually increased since the low-point, accompanying the increasing DEEP salinity. The SURFACE salinity seems thus to have followed the DEEP salinity pattern only remotely and with a time lag. Again, the pattern above also describes the behaviour of the SURFACE and DEEP density. With respect to temperature, the strong interannual variation in the SURFACE — caused mainly by the severity of the early winter before the time of the station visits but also by the variable timing of the station visits — makes it difficult to detect any definite temporal patterns.

Near-bottom O$_2$ in the BS

During the period since the low-point of the DEEP salinity in the early 1990s, both the near-bottom density and the near-bottom O$_2$ saturation contain fluctuations superimposed on the general increasing and decreasing trends, respectively (Fig. 4A). Here, the density variation explains much of the observed variation in the O$_2$ saturation. If we postpone the density data in time by two years, the density pattern closely resembles a mirror image of the O$_2$ saturation pattern; i.e., whenever the density increased, the O$_2$ conditions worsened with about a two-year time lag, and vice versa.
The near-bottom $O_2$ concentration in the BS has gradually decreased since the low-point of the DEEP salinity in the early 1990s (Fig. 4B). During the studied period, the lowest $O_2$ concentrations fell within the range of 4 to 5 ml $l^{-1}$, with episodic events values of 3.5 to 4.5 ml $l^{-1}$ being recorded. The average value for the entire dataset (Case 1) decreased from 6.6 at the low-point to 5.1 ml $l^{-1}$ at the end of the study period. Regardless of the approach used when choosing the dataset for the linear extrapolation (Cases 1–3), the predicted point in time for the emergence of hypoxia is point-blank as occurring more than fifty years in the future (Table 2C). Most interestingly, the Case 3 estimate for the emergence of hypoxia for the first time ever, produced a predicted point in time that was a decade later than that produced by the Case 1 estimate for the emergence of hypoxia on a frequent basis. This seemingly impossible result indicates that, while the overall near-bottom $O_2$ conditions have slowly worsened, the lowest recorded values of $O_2$ concentration have decreased at a slower rate than the values in general, thus obscuring the linear extrapolations that reach out into the future.

Impact of the NBP on the BS

As the NBP waters drift into the ÅS and further enter the BS as a deep current, the water density gradually decreases through mixing processes. The density of the surface layer of the NBP is much closer to the density of the deep layer of the ÅS than to the density of the deep layer of
the BS (Fig. 5). The NBP water thus flows into the ÅS relatively unchanged, and most of the mixing occurs in the ÅS. The close temporal resemblance of the SURFACE density of the NBP and the DEEP densities of the ÅS and the BS underlines the close hydrodynamic connection between the basins (Fig. 5B and Table 2B).

Discussion

External factors govern the O₂ dynamics of the BS

The environmental status of the BS has received fairly little public attention, mostly because the BS has not experienced such drastic environmental problems like those that have emerged elsewhere, e.g., in the GOF. With regard to O₂ dynamics, the problems in the GOF originate partly from the catchment (increased organic load, both allochthonous and autochthonous, consuming more O₂) but stem partly from the direct connection with the NBP (incoming highly saline waters poor in O₂). Here, I describe those hydrographical and/or topographical features that have helped the BS to maintain the favourable O₂ conditions in its deep layers. Those features are external and, in this respect, the BS is in a fortunate position in that it is not so directly connected to the NBP like the GOF is.

Hydrodynamic cascade of NPB–ÅS–BS

The hydrodynamic forcing of the NPB on the deep waters of the BS is mediated, and somewhat obscured, by the ÅS and the ridge formations surrounding it to the north and south. The water flow from the NBP to the BS is most probably continuous, with the main mechanism being the winter cooling of the NBP water, its subsequent sinking over the Southern Åland Sill, mixing in the ÅS, flowing over the Understen–Märket sill of the southern Quark to eventually form the deep BS water mass (Marmefelt and Omstedt 1993, Leppäranta and Myrberg 2009). The first key aspect here is that the sill depth of the Southern Åland Sill between the NPB and the ÅS is 70 m (Marmefelt and Omstedt 1993), i.e., either above or at the typical halocline depth of the NBP. The main bulk of the water entering the ÅS, therefore, originates from above the halocline of the NBP and not from the deep waters, which are highly saline and poor in O₂ (Kullenberg 1981, Leppäranta and Myrberg 2009). This aspect in itself has a drastic influence on the BS deep water characteristics.
The Southern Åland Sill largely determines the characteristics of the waters flowing into the BS. Within this framework, the subsequent mixing occurring in the ÅS alters the characteristics of the waters ultimately entering the BS. The second key aspect here is that the inflow to the BS takes place as a deep current, due to the strong riverine-induced southward surface current through the ÅS. As the density difference between the deep ÅS water and the water at the sill depth of the southern Quark is still significant (Fig. 5A), the deep waters have to get less dense, i.e., less saline, to be able to cross the sill. Indeed, the deep water salinity changes at SR5 have been reported to considerable resemble those occurring at the sill depth of the Understen-Märket trench (about 90 m) in the northern ÅS (Hietala et al. 2007).

Taken together, the northward water flow over the ÅS can be seen as a two-step filtration mechanism: first, a clear cut-off mechanism where the densest waters are prevented from entering the ÅS, and then a mixing/dilution mechanism which further shapes the water characteristics before the water’s actual entry into the BS. This mechanism is not expected to change unless there occurs an alteration in the water budget of the Baltic Sea that leads to a permanent rise in the halocline in the NBP. If this scenario were ever to occur, the deep water characteristics of the ÅS and BS would be reshaped towards a state that would to some extent resemble the current state of the NBP below the halocline, that is, more saline and poorer in O2. Alterations on this scale are only caused by widespread climatological changes over the northern hemisphere. The scenarios linked to the on-going climate change predict that, in the future, wintertime precipitation, and hence, river runoff, will increase in quantity, and the inflow of saline water from the North Sea will decrease in quantity (BACC 2008, ICES 2008). Both of these tendencies work against the rise of the halocline in the NBP. To conclude, even though the salinity of the deep inflow entering the BS has been increasing since the low-point of the DEEP salinity in the early 1990s, this trend is not expected to continue throughout the effective time-span of climate change.

Long-term changes in the BS salinity

After its gradual increase spanning several decades and reaching its highest observed level of the last century in the late 1970s (Fonselius and Valderrama 2003), the DEEP salinity in the BS took a different path and decreased throughout the 1980s. A significantly decreased deep water inflow from the NBP (ICES 2008) as well as a sudden and distinctive increase in the river inflow into the Gulf of Bothnia (Fig. 6) in the early 1980s probably had a synergistic effect that enabled this alteration. Naturally, the change in precipitation did not concern only the catchment of the Gulf of Bothnia: in the same period, river inflows increased in the Baltic Sea area in general (Winsor et al. 2001). The progressively decreasing salinity of the BS was, along with increasing temperature, the main driver for the regime shift taking place in the BS in 1988–1989 (ICES 2008). This shift has been proposed to have been driven by climate change (ICES 2008): the global warming period started in around 1980 (Lehmann et al. 2011). It has also been explained by climatological fluctuations: the NAO index increased to its highest positive observed level of the last century from 1988 onwards (Kraberg et al. 2011, Lehmann et al. 2011). Interestingly, the current dataset could not clearly reveal the increase in temperature that evidently had taken place in the Baltic Proper (Alheit et al. 2005). The Baltic Sea renews most of its heat annually (Stigebrandt and Gustafsson 2003), which is a challenge for the conventional monitoring approach. This approach may not be able to reveal trends in temperature, which is a less conservative parameter than, e.g., salinity with a corresponding annual turnover proportion of only 3%.

The trend in the DEEP salinity in the BS again took a different path in the early 1990s, and has increased ever since. As the remnants of the major Baltic inflows can be traced all the way to the Gulf of Bothnia (Fonselius and Valderrama 2003), it is tempting to speculate that the upward step in the DEEP salinity in 1995–1996 was ultimately driven by the major Baltic inflow in 1993, which was the 5th most intensive inflow during the past 130 years (Leppäranta and Myrberg 2009).

There are not many candidates for the driving force behind the increasing trend in the DEEP
salinity since the low-point in the early 1990s. As the river inflow does not show any apparent trend at this time (Fig. 6), the driver has to be either the increased salinity of the deep inflow entering the BS or the increased volume of the deep water inflow. In this case, the former seems to be the more plausible option; the increasing DEEP density in the NBP and ÅS shown in Fig. 5B was a manifestation of the increasing salinity (data not shown).

**Past and future of the O₂ conditions in the BS**

The linkage between the BS and the NBP is obscured by the topographic and hydrographic features of the ÅS. Here, I describe what the existence of the topographic and hydrographic features in the ÅS area actually means to the BS in terms of the O₂ dynamics.

This study relied on the data collected by conventional monitoring programs. Thus, it is right to ask whether this approach produced trends that could describe the true natural fluctuations in adequate detail for drawing reliable conclusions regarding the state of the BS. In my opinion, the dataset was of sufficiently high-quality and dense to describe the O₂ regime of a sea area that is less dynamic with respect to O₂ than, e.g., the GOF. Particularly, the absence of direct hydrodynamic forcing of the NBP, that take place in the GOF, is the key for the more stable and predictable O₂ regime in the BS.

**Deep-water O₂ conditions**

The worsening of the DEEP O₂ conditions in the BS is not a novel observation; to upscale, this has been going on more or less constantly for the past century (Fonselius and Valderrama 2003), and an adverse change of at least a similar magnitude to the one presented here occurred between 1910 and 1930. The interpretation of a long environmental dataset frequently ignores the fluctuations found on a time-scale shorter than the actual span of the data. In this article, I concentrate on recently-occurring fluctuations over the period of a decade or two.

Hypoxia does not seem to pose any immediate threat to the offshore BS environment. When the available data since the low-point in the DEEP salinity in the early 1990s are extrapolated to the future, hypoxic conditions — frequently defined as < 2 mg O₂ l⁻¹ or < 1.4 ml O₂ l⁻¹ (Conley et al. 2009) — would not be observed near the bottom before the 2060s, regardless of the type of approach. By that time, climate change will affect the BS system, e.g., the water balance of the BS, in ways that are difficult to predict at this stage (BACC 2008). Thus, the predictions presented here have to be regarded only as a first-order estimate based on current knowledge. However, they still suggest that profound changes to the hydrographic regime of the BS and/or to the anthropogenic load on the BS will have to take place if any hypoxic events are to occur in the offshore BS in the near future.

If, however, hypoxic events do in fact emerge in the BS at some level, the hypoxic conditions at the sediment–water interface will have adverse impacts on the benthic fauna that is currently thriving in the BS (HELCOM 2009a). Hypoxic events would not have any effect on the phosphorus cycle of the BS per se. For this to happen, the near-bottom O₂ conditions should reach anoxic levels. The pivotal geochemical consequence of anoxia is liberation of the bio-available phosphorus in the benthic system, i.e., the internal phosphorus loading (Conley et al. 2009). In the GOF, this process has led to a situation where the sediment processes largely control the water-column phosphorus stock, and where phosphorus efflux from the sediments has been suggested to have triggered the extensive cyanobacterial blooms in the late 1990s and early 2000s despite marked decreases in the external phosphorus loading to the GOF in the 1990s (Pitkänen et
The accelerated internal phosphorus loading in the GOF was a consequence of both the high allochthonous nutrient load leading to a high autochthonous load of organic matter to the benthic system, as well as the GOF’s tendency to display strong vertical density gradients. The BS has got neither of these. The share of organic matter in the BS’s sediments is less than half of that in the GOF’s sediments (Lehtoranta et al. 2008). The halocline in the BS is sufficiently strong to restrict vertical mixing only in the central basin, and even there periods of mixing right down to the sea floor may occasionally take place (Kullenberg 1981, Leppäranta and Myrberg 2009). As a result, internal P loading has not occurred in the BS, and PO₄-P concentrations in the near-bottom layers of the BS have not increased recently, which sets the BS apart from the GOF (Fig. 7). To conclude, there is no reason to expect that a similar kind of cause-and-effect cascade of eutrophication — increased autochthonous load of organic matter — anoxia — internal loading of phosphorus that has progressed in the GOF could take place in the BS. Considering the utmost importance of this phosphorus source to the past adverse development of the GOF, it is apparent that the offshore BS will not follow the path of the GOF in this respect.

Vertical and lateral O₂ flows shaping the deep O₂ conditions

In the Baltic Sea, water density variation is largely dictated by salinity variation (Kullenberg 1981); this fact has also been verified by this study. In the Baltic Sea, the salinity-driven vertical density gradient is strong and relatively independent of the time of year, and generally serves as a stable and effective barrier to the vertical flow of O₂. In the offshore BS, however, the salinity-driven vertical density gradient of 0.5 to 1.5 kg m⁻³ is much weaker than in the NBP (Table 1), which leaves room for a vertical O₂ flow to be a more important O₂ supply for the deep layer in the BS than in the more southerly basins of the Baltic Sea (Table 1). The lateral deep water flow, originating from the ÅS and ending up to form the deep water mass of the BS, also carries O₂ to the deep waters. The drawback to this O₂ supply is that it is mediated by the flow of water of higher salinity than the BS water, which strengthens the density gradient in the deep layer and restricts the vertical O₂ flow.

When the near-bottom density data are postponed in time by two years, the near-bottom density pattern closely resembles a mirror image of the near-bottom O₂ saturation pattern (Fig. 4A). The hypotheses to explain the observed relation are as follows: (i) the organic matter — descending to the sea floor and gradually being decomposed by O₂-consuming microbial processes — is a constant sink of O₂. Both a lateral and a vertical supply of O₂ are required to compensate this consumption in order to retain the good O₂ conditions that are typically met with in the deep layer of the BS. (ii) Of the supplies of O₂, the lateral one can be considered a constant source. Thus, it retains the “baseline” O₂ conditions in the deep layers. The vertical one, in turn, gets to some extent diminished whenever the water-column density gradient gets stronger. Whenever the vertical O₂ supply decreases, the O₂ conditions worsen, only not to the brink of hypoxia. This could explain the temporally opposite behavior of the near-bottom density and O₂ saturation patterns. (iii) The O₂ conditions do not worsen abruptly after the decrease of the vertical O₂ supply. The remaining lateral O₂ flow compensates to some extent for the O₂ demand of the biological decomposition process, which leads to the observed temporal lag between the near-bottom density and O₂ saturation patterns.
The coastal area

The coastal area of the GOF is often a mosaic of water and land, and in terms of water circulation, the coastal basins can be quite isolated from the general offshore circulation pattern. In these coastal basins, O₂ problems have occurred for a long time, and their extent has lately increased, most probably due to eutrophication (Wallius 2006). Some scattered areas of the Finnish coastal zone of the BS also suffer from a restricted water exchange, and sporadic hypoxic events have also been observed there (Lundberg et al. 2009). However, this zone is rather narrow and the area subject to hypoxia is rather limited; even there, hypoxic events only have been observed in the deepest of the locally-restricted basins (Conley et al. 2011). If assessed by the level of degradation of the benthic system, the width of the coastal area in the BS chronically influenced by river runoff and/or direct anthropogenic load extends only to about 10 km at most, even off the largest towns or industrial facilities (Sarvala and Sarvala 2005; my interpretation based on the data). In the outer and exposed archipelago areas, wind-driven mixing and the cyclonic water circulation pattern of the BS (Myrberg and Andrejev 2006), and particularly vertical mixing that is not suppressed by any salinity-driven density gradient (Table 1) ensure that the water exchange remains at a level high enough to prevent the emergence of even episodic hypoxic events. Consequently, since the 1990s, the observed winter O₂ concentration in the DEEP at the coastal stations studied has consistently fallen within the range of 7 to 9 ml l⁻¹ (data not shown), as compared with the 4 to 5 ml l⁻¹ of the DEEP at the offshore stations. To conclude, in the Finnish coastal zone of the BS, any O₂ problems will most probably continue to be concentrated in the inner coastal areas.

Conclusions

Even though the oxygen conditions in the deep layers of the offshore Bothnian Sea have somewhat deteriorated during the most recent two decades, there is little evidence to suggest that hypoxic events could emerge there in the near future. For consistent hypoxic events to emerge, the climate of the northern hemisphere would have to force profound changes in the hydrographic regime of the BS, and some of these changes are not predicted to happen in the time span of the on-going climate change. Nevertheless, the decreasing O₂ trends in the offshore Bothnian Sea warrant paying of close attention to the future development of the BS environment. As the emergence of hypoxic events does not seem likely in the near future, there is no reason to expect that the internal loading of phosphorus will occur in the offshore Bothnian Sea. Considering the utmost importance of this phosphorus source to the past adverse development of the Gulf of Finland, it is apparent that the Bothnian Sea will not follow the example of the Gulf of Finland with regard to eutrophication.

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