Unusual Baltic inflow activity in 2002–2003 and varying deep-water properties*

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Abstract

The unusual sequence of inflow events into the Baltic Sea that occurred in 2002 and 2003 includes the first ever important baroclinic inflow to be described (August 2002), the Major Baltic Inflow (January 2003), which gave rise to the highest oxygen levels in the Gotland Deep since the 1930s, and the baroclinic inflow (August 2003) that elevated the Gotland Basin deep water salinity to values last observed in 1977, and caused the surface salinity to rise again. From these trend changes, salt residence times were estimated at about 20 years in the deep waters and 30 years above the pycnocline. Ventilation of the remote Karlsö Deep took until 2005, two years after the inflow event responsible, at a time when the Bornholm and Eastern Gotland Basins were already returning to stagnation.

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The complete text of the paper is available at http://www.iopan.gda.pl/oceanologia/

1. Introduction

The inflow activity of recent years from the Kattegat into the Baltic Sea was caused by a quite unusual sequence of events: the warm summer inflow of 2002 was followed by a cold gale-forced one in January 2003, and again by a warm inflow in the summer of 2003. The highly contrasting thermal signatures provided natural 'tracers' and allowed a clear insight into the dynamics of deep water propagation through the main basins of the western and central Baltic. Against a background of deep water, which had been stagnating since 1993 and was exceptionally warm after the autumnal inflow of 1997, these inflows brought about dramatic fluctuations in both oxygen and nutrient concentrations. Details of these different inflow processes are reported in several earlier papers (Matthäus et al. 1998, Hagen & Feistel 2001, Feistel et al. 2003a,b,c, 2004a,b, Piechura & Beszczyńska-Möller 2004). These inflows belonged to two qualitatively rather different types.

So-called barotropic inflows are characterised by the following features (Wyrtki 1954, Franck et al. 1987, Matthäus & Franck 1992, Fischer & Matthäus 1996, Feistel et al. 2003b):

– They are driven by barotropic pressure gradients, especially sea level differences.
– They appear during persistent westerly gales (mostly in autumn, winter, spring).
– They import salt (typically 2 Gt) into the Baltic along with the water volume import (typically 200 km^3).
– They import oxygen-saturated waters (typically 1 Mt of O_2) if they happen in winter or spring.
– They pass through the Sound and the Belts.

In contrast, so-called baroclinic inflows have these properties (Knudsen 1900, Thiel 1938, Hela 1944, Wüst et al. 1957, Welander 1974, Jacobsen 1980, Matthäus et al. 1983, Feistel et al. 2003c, 2004a):

– They are driven by baroclinic pressure gradients, especially horizontal salinity differences.
– They appear during persistently calm wind conditions (usually in late summer).
– They import salt into the Baltic along with water volume export.
– They import oxygen-deficient waters, but ventilate the deep Baltic basins by entrainment.
– They pass only through the Great Belt/Darss Sill gateway.

Winter and spring inflows of either type give rise to higher salinities, low temperatures and increased oxygen levels in the deep basins. Summer and autumn inflows increase salinity and raise temperatures but carry only very little oxygen.

The inflow events of 2002–2003 could be tracked, beyond the regular monitoring cruises, by the long-term mooring in the Eastern Gotland Basin (EGB) recording temperature and current at about 200 m depth on the north–eastern flank of that basin. We would like to draw particular attention to the exceptionally warm inflow processes, especially in view of recent monitoring observations indicating an unexpectedly intense warm summer inflow in 2003. In 2004, no additional inflow of comparable effect could be detected at the long-term mooring, either in temperature or in current records, in contrast to the overflow over the Słupsk Sill observed in May 2005 (Piechura, unpublished).

The recent, unusual succession of inflow events in 2002 and 2003, although rather different in their causes and dynamic properties, have together raised the salt content of the Baltic Sea. We use the related salinity change for estimating the deep water and surface water residence time.

Deep water oxygen conditions exhibited a strong temporal improvement after the major inflow of 2003 but have since gradually returned to stagnant conditions. The dramatic changes observed in the levels of oxygen and hydrogen sulphide manifested themselves rather differently and with significant delays in the different deep central basins.

In the results section we describe the lasting impact of the exceptional 2002–2003 inflow series on the deep water properties with respect to the subsequent evolution of temperature (section 3.1), salinity (section 3.2) and oxygen (section 3.4). We introduce a simple dynamic model for the salinity budget (section 3.3) in order to exploit the differential salinity trend change in conjunction with those inflows to estimate residence times.

2. Material and methods

The measurements presented in this paper cover the time period between 1968 and 2005. The observations were part of the Baltic Monitoring Programme (COMBINE) of the Helsinki Commission (HELCOM) carried out by the Leibniz Institute for Baltic Sea Research in Warnemünde (IOW), Germany, and its predecessors. The ship-borne investigations by standard CTD probes were supplemented by continuously recorded currents and temperatures at three levels (174 m, 204 m, 219 m) at the mooring in the Eastern Gotland Basin (EGB) (Hagen & Feistel 2001, 2004). The hydrographic and chemical variables studied and the methods used were based on the standard guidelines for the COMBINE programme of HELCOM (HELCOM 2002). The positions of the IOW monitoring stations 213 (55°15′N, 15°59′E), 222 (55°13′N, 17°04′E) and 271 (57°19.2′N, 20°03′E), which are considered in the Results section in greater detail, and of the Eastern Gotland Basin mooring (EGB, 57°23′N, 20°19.5′E) are shown in Fig. 4 (page 30). Salinities are expressed on the Practical Salinity Scale 1978, indicated by PSU. Digital shorelines are taken from Feistel (1999) and the digital bottom topography is from Seifert & Kayser (1995).

3. Results

The dramatic changes in the central Baltic between 2002 and 2005 as consequences of the inflow sequence in 2002 and 2003 will be discussed here separately for temperature, salinity and dissolved oxygen. We shall compare the current events with corresponding anomalies on longer timescales.

3.1. Temperature

Water temperatures at about 200 m depth in the Eastern Gotland Basin rose to above 6°C after the warm inflow of autumn 1997 (Hagen & Feistel 2001), and were further raised by the inflow of October 2001 (Feistel et al. 2003a), as recorded by the EGB long-term mooring (Fig. 1). Both events,

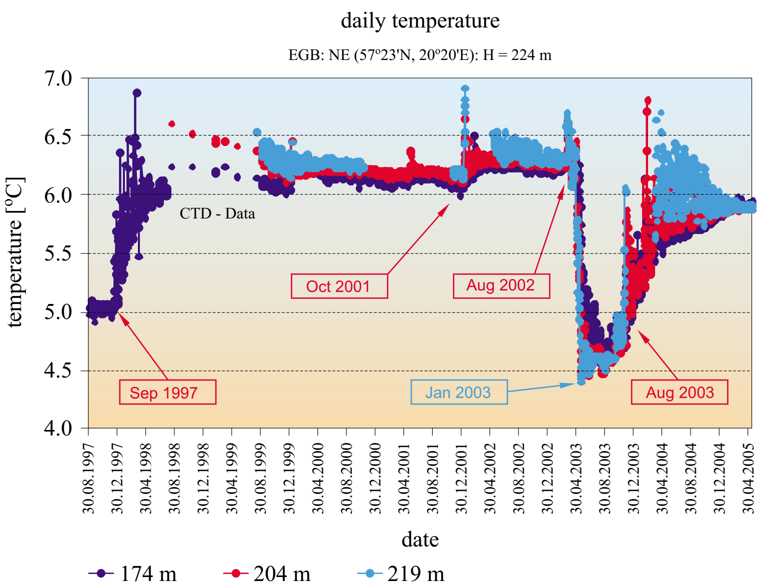

Fig. 1. Temperature series 1997–2005 of the EGB mooring near the Gotland Deep (Fig. 4) at 174, 204 and 219 m depth. Bathymetric depth at the anchor position is 224 m. The temperature signals caused by the latest cold (January 2003) and warm (all others) inflow events are indicated by arrows

attributed to barotropic inflows, were caused by early autumn gales, and only temporarily improved the stagnant oxygen conditions prevailing since the major inflow of 1993. The exceptional baroclinic inflow, which started in August 2002 and lasted, with interruptions, until October 2002, in combination with a small barotropic inflow at the end of October, brought very warm waters into the Gotland Basin (Feistel et al. 2003c, 2004a). Shortly afterwards, it was replaced by the very cold and dense water of the Major Baltic Inflow that occurred in January 2003, and was reinforced by some subsequent smaller events in spring (Feistel et al. 2003b). With temperatures below 4°C, this inflow terminated, as was thought at the time, the rather prolonged warm period in the Gotland Deep, which was exceptional if compared to the past long-term record (Fig. 2).

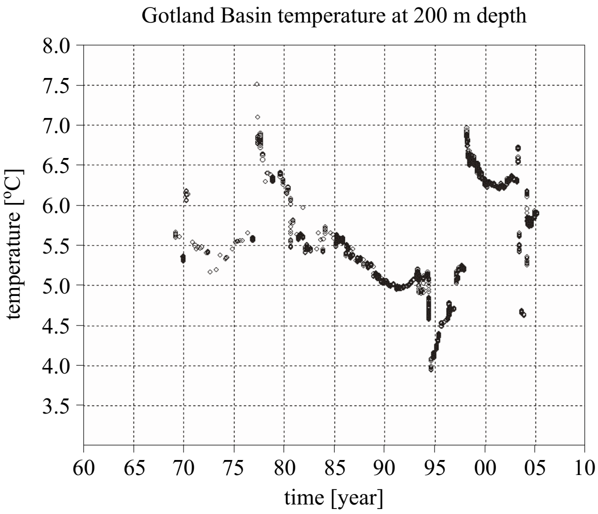

Fig. 2. Temperature series 1968–2005 at the Gotland Deep station 271 at 200 m depth

But, already in December 2003 temperatures started rising steeply again to values at and above 6°C, caused by the short but very intense baroclinic inflow of summer 2003 (Feistel et al. 2004b, Piechura & Beszczyńska-Möller 2004), as recorded by the EGB mooring (Fig. 1). This happened quite unexpectedly, since the small baroclinic inflow of August 2003 had not been expected to almost erase the effect of the preceding major inflow at that depth. In the Gotland Basin (but not in the Bornholm Basin), the water of this baroclinic inflow was denser than that of the barotropic event in January. It was capable of lifting up the ambient deep water and raising the halocline, but also of forcing it to overflow into neighbouring upstream basins.

The considerable temperature fluctuations caused by the baroclinic 2003 inflow took a whole year to fade away (Fig. 1). It is likely that inflow processes in general trigger a dynamic repercussion of comparable duration of various hydrographic parameters, but the existence of a continuously recording device together with the extreme temperature contrast of the successive events made these fluctuations especially obvious in the temperature series. There has been no visible sign of any subsequent inflow signal reaching the Gotland Deep since then. The present temperature of almost 6°C is again exceptionally high compared to the long-term record since 1968 (Fig. 2).

3.2. Salinity

The development of salinity in the surface layer and in the deep water at the central station IOW 271 in the Gotland Basin over the last, almost 4 decades is displayed in Fig. 3. The typical seesaw-like sequence at 200 m, i.e. the steep increase after inflows and the slow decrease in the subsequent stagnation periods, is reflected in the surface water salinity only in a delayed and attenuated manner. It was especially the brief inflow of August 2003 (Feistel et al. 2004b) that pushed the deep-water salinity from 12.5 to 12.96 PSU on 18 July 2004 (Wasmund 2004), a 30-year high after the 13.05 PSU recorded there on 16 May 1977 (Nehring 1990, Nehring & Matthäus 1990). This surprisingly marked impact of the short inflow on the central basins is attributed to its high salinity and very weak vertical mixing in the western Baltic, expressed by an extremely steep halocline and the presence of an intermediate cold water layer in the shallow Darss Sill region at this time (Feistel et al. 2004b). Its effect on the surface salinity has made obvious the trend change in this layer (Fig. 3). This turning point can now be used

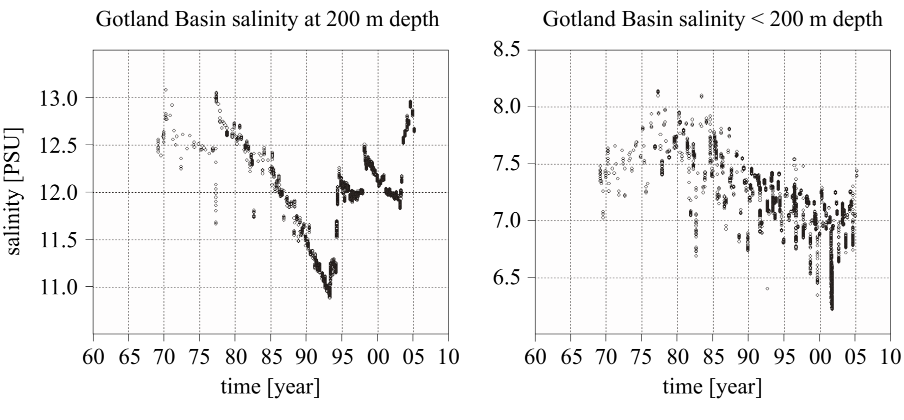

Fig. 3. Salinity at 200 m (left) and in the 20 m surface layer (right) at the Gotland Deep station 271 between 1968 and 2005

to estimate in a simple manner the residence times for salt in the surface and the bottom layer.

In the Baltic Sea, residence time estimates of water or of dissolved substances vary from below 10 to more than 30 years (Folke et al. 1991, Holm 1995, Matthäus & Schinke 1999, Meier & Kauker 2003). Numerical models provide a median water age of 8 years for the Eastern Gotland Basin in the stagnation periods (Meier, unpublished), but the stratification spin-up time scale for salinity is estimated to be about 100 years (Hansson & Omstedt 2005).

From the long-term series of salinities in the surface layer and in the deep water, we shall estimate different residence times for both water bodies, making use of the observed delay between their pronounced salinity turning points in 2002 and 1993 (Fig. 3).

3.3. A salinity budget model

The simplest dynamic salinity budget model we can apply has 2 boxes, with salinity S in the surface layer and $S_B(t)$ in the bottom layer, where the latter is thought to be controlled by external sources (inflows). The transfer of salt from the deep to the surface layer is assumed proportional to the salinity difference between both, and the salt loss of the surface layer proportional to its actual salinity,

$$\frac{\mathrm{d}S}{\mathrm{d}t} = -\alpha S + D \times (S_B(t) - S), \tag{1}$$

or, with $\lambda = \alpha + D$

$$\frac{\mathrm{d}S}{\mathrm{d}t} + \lambda S = D S_B(t). \tag{2}$$

Here, α is the surface salinity flush rate, and D is the vertical exchange rate, both to be estimated in the following. The solution of the differential equation is, after the asymptotic disapperance of the initial conditions,

$$S(t) = D \int_{-\infty}^{t} \exp(\lambda\tau - \lambda t) S_B(\tau) \mathrm{d}\,\tau. \tag{3}$$

In terms of signal transfer, the surface salinity is thus obtained by a convolution of the input signal $S_B(t)$ with a response function in the form of an exponential low pass filter with limiting frequency λ.

We now go on to consider two simplified standard situations for the temporal development of bottom salinity, (i) constant salinity, and (ii) the trend turnover from decreasing to increasing salinity.

Case (i) is given by

$$S_B(t) = S_0 = \mathrm{const.} \tag{4}$$

The result from eq. (3) is

$$S(t) = DS_0 \exp(-\lambda t) \int_{-\infty}^{t} \exp(\lambda\tau) \mathrm{d}\,\tau = \frac{DS_0}{\lambda} = \frac{DS_0}{D+\alpha}. \tag{5}$$

We can use this relation to estimate the ratio

$$\frac{\alpha}{D} = \frac{S_0 - S}{S} \tag{6}$$

of the vertical diffusion rate D to the advective dilution rate α, or of the related residence times of both layers. If we read from Fig. 2 the long-term mean values as $S_0 = 12$ PSU and $S = 7.3$ PSU, we get

$$\alpha/D = 0.64. \tag{7}$$

Case (ii): A simple analytical model for a turning point at some arbitrary time $t = 0$ is the 'tent' curve

$$S_B(t) = k\,|t|\,. \tag{8}$$

Application of eq. (3) yields for $t > 0$

$$S(t) = -Dk \int_{-\infty}^{0} \exp(\lambda\tau - \lambda t)\tau \,\mathrm{d}\,\tau + Dk \int_{0}^{t} \exp(\lambda\tau - \lambda t)\tau \,\mathrm{d}\,\tau. \tag{9}$$

After performing the requisite integrations, we obtain the solution

$$S(t) = \frac{Dk}{\lambda^2}[2\exp(-\lambda t) + \lambda t - 1]. \tag{10}$$

Looking for the turning point in the surface salinity, we have to solve $\mathrm{d}S/\mathrm{d}t = 0$:

$$\frac{\mathrm{d}S}{\mathrm{d}t} = 0 = \frac{Dk}{\lambda^2}[-2\lambda\exp(-\lambda t) + \lambda]. \tag{11}$$

Thus, the turning point of the filtered signal is independent of the slope k and appears with a delay of

$$t = \frac{\ln 2}{\lambda}. \tag{12}$$

For the given turning points we read from Fig. 3 $t = 2002 - 1993 = 9$ years, so we obtain a low-pass filter frequency of

$$\lambda = \frac{\ln 2}{9 \text{ yr}} = 1/13 \text{ yr}. \tag{13}$$

With $\lambda = \alpha + D = 1/13$ yr and $\alpha/D = 0.64$ from eq. (7), we finally find as estimates for the particular residence times the values,

$$D \times (1 + 0.64) = 1/13 \text{ yr}, \quad \text{i.e.} \quad 1/D = 21 \text{ yr}, \tag{14}$$

for the deep water, and,

$$1/\alpha = 21 \text{ yr}/0.64 = 33 \text{ yr}, \tag{15}$$

for the surface layer.

The mean vertical transport rate D, corresponding to a deep-water residence time of 21 years as estimated before (eq. (14)), includes the occasional uplift processes of deep water by inflow events. In the absence of the latter, as observed during the long stagnation period visible in Fig. 3, the typical salinity decrease at 200 m is about 1.5 PSU from 1983 to 1993, or slightly more than 1% per year, which corresponds to a residence time of almost a century. These two very different times suggest that deep water uplift during inflow events and the subsequent erosion of the pycnocline is by far the dominating vertical transport mechanism. The flushing rate α in the surface layer (i.e. the layer above the pycnocline) with a residence time of about 33 years (eq. (15)), however, is thought to be controlled mainly by the permanent fresh water excess, i.e. the outflow rate, and should be rather insensitive to the frequency of inflow events.

3.4. Oxygen

The changes in the near-bottom distributions of dissolved oxygen and hydrogen sulphide are displayed for 2002–2005 in Fig. 4. In 2002, at the end of the long stagnation period lasting since 1993, extensive areas of oxygen deficiency or even substantial concentrations of H_2S were encountered in the central Baltic below 70 m depth. One year later, in spring 2003, the Major Baltic Inflow of January 2003 had ventilated the Bornholm, Gdańsk and Eastern Gotland Basins with high oxygen concentrations. Nevertheless, remnants of H_2S are still found in the Gotland Basin area, and the anoxic situation in the Farö Deep (IOW station 286) and beyond is unchanged or has deteriorated further since 2002.

Especially the baroclinic inflow of summer 2003 pushed the oxygenated waters further north, ventilating the Farö Deep, but at the same time displaced more of the old, stagnant waters into the basins west of Gotland. It took until spring 2005, two years after the actual major inflow event, for its oxygenated waters to reach as far as the Karlsö Deep (IOW station 245), south–west of Gotland. At this time, the Bornholm and Eastern Gotland Basins were already gradually returning to stagnation with increasing H_2S concentrations. Some regions, especially around the Landsort Deep (IOW station 284), were never completely ventilated at all throughout the entire water substitution process.

The Słupsk Channel (IOW station 222) appears unaffected by the currently increasing oxygen depletion in progress in the central basins, and especially in the adjoining Bornholm Basin. The Słupsk Channel bottom water is usually formed by waters flowing over the Słupsk Sill at a depth of

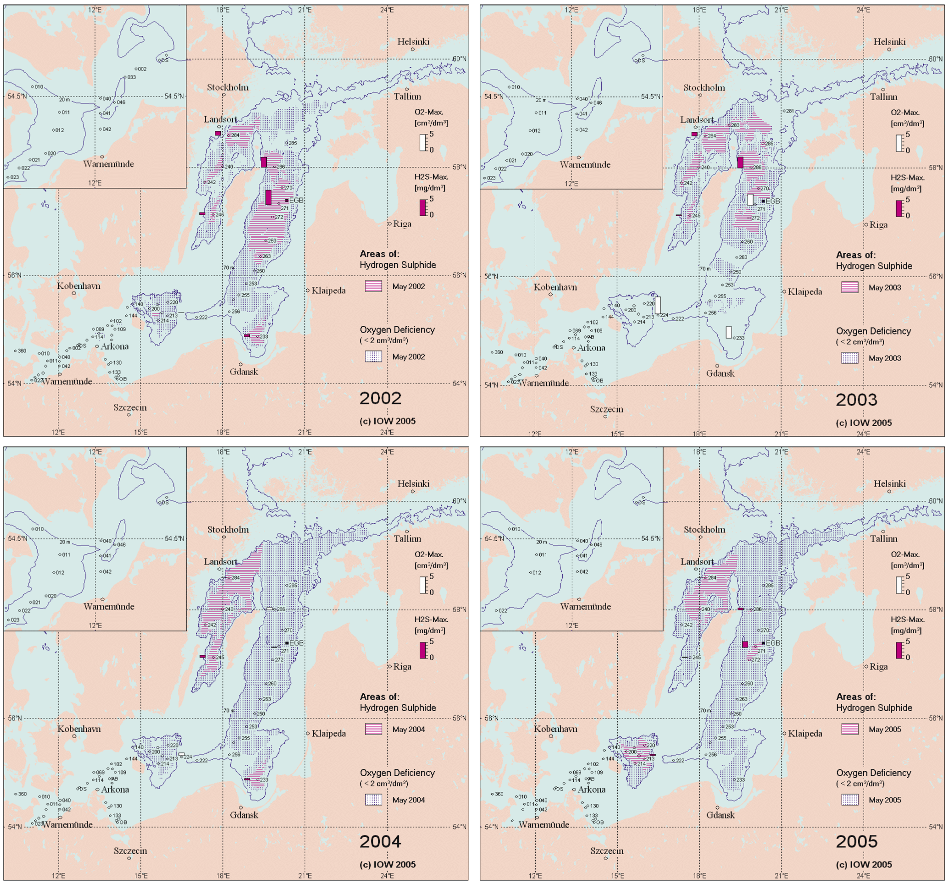

Fig. 4. Near-bottom distributions of oxygen and hydrogen sulphide concentrations for the months of May in 2002–2005. The inset in the top left-hand corner magnifies the western Baltic

about 60 m. Such overflow processes have turned out to be very complex, as observed in great detail in October 2003 during the ongoing summer inflow of August and recently classified by Paka (unpublished). The threshold is at about the same depth as the halocline, above which a persistent layer of cold, oxygen-rich winter water is located. In this way, water flowing over the sill can exchange heat, salt and oxygen with that layer and carry it into the near-bottom levels of the Słupsk Channel and the basins beyond. This mechanism was already considered responsible for the surprising ventilation of the Gdańsk Basin in autumn 2002 by the preceding baroclinic summer inflow of oxygen-poor Kattegat surface waters (Feistel et al. 2003c, 2004a).

The temporal development of oxygen and hydrogen sulphide concentrations after 1990 is displayed in Fig. 5 for the Bornholm Basin and the Słupsk

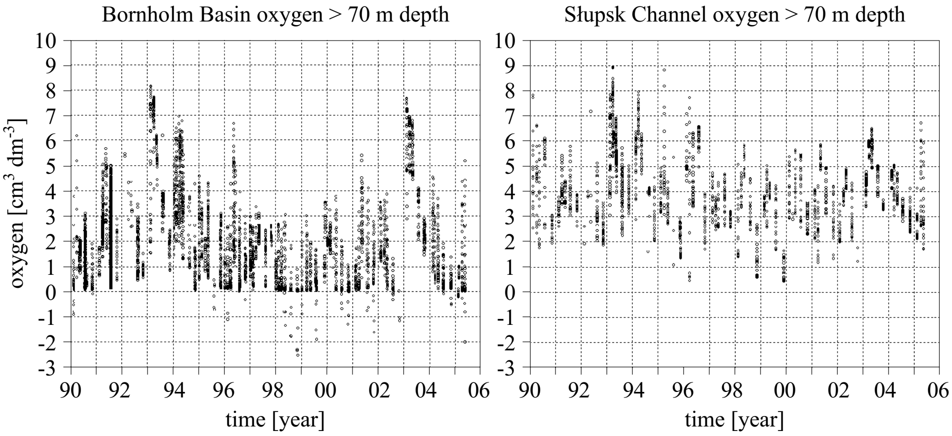

Fig. 5. Time series (1990–2005) of dissolved oxygen and hydrogen sulphide, shown as negative oxygen equivalents, at depths greater than 70 m in the Bornholm Basin (IOW station 213, left panel) and, separated only by the Słupsk Sill and the Słupsk Channel (IOW station 222, right panel)

Channel at depths greater than 70 m. Even though the Słupsk Channel is further east, its average oxygen level of 4.1 $cm^3\ dm^{-3}$ for the period shown is significantly higher than the average of 2.0 $cm^3\ dm^{-3}$ in the Bornholm Basin. Anoxic conditions were not observed in the former. While the major inflows of 1993 and 2003 look rather similar in the Bornholm Basin, the 2003 event carried much less oxygen over the Słupsk Sill, but in conjunction with the accompanying baroclinic inflows, it kept the minimum oxygen level there above 3 $cm^3\ dm^{-3}$ for almost two years. The stagnation period from 1995 to 2002 is characterised by the repeated annual occurrence of anoxic conditions in the Bornholm Basin, interrupted only by the 1997 autumn inflow with a low oxygen content. Since 2004, these conditions have re-established themselves: indeed, the highest value of hydrogen sulphide ever measured in the Bornholm Basin, 70 μmol dm^{-3} or $-3.14\ cm^3\ dm^{-3}$ oxygen equivalent, was reported in the late summer of 2005 (Aneer & Höglander 2005).

4. Discussion

The data gathered from regular CTD monitoring stations in conjunction with the continuous temperature record from the long-term mooring in the Eastern Gotland Basin provide a fairly complete and consistent picture of the complex and dramatic events that unfolded in the Baltic deep waters in 2002 and 2003, with after-effects lingering even till 2005. Three successive inflow events changed temperatures from high to low and back to high in a short period of time, high hydrogen sulphide to high oxygen concentrations, decreasing again towards stagnation, and raising salinity to values not seen for almost 30 years. Old, stagnant waters were in part shifted into more remote basins, and in part lifted up with the release of enormous amounts of salt (and nutrients) into the surface water. Compared to past inflow processes, the current series was not as significant in terms of its water volumes, salinities and oxygen levels. But fresh light was shed on the dynamic details of the water renewal processes by its pronounced temporal and spatial temperature contrasts, and by the important role baroclinic inflows may play, as emphasised by the present salinity high in the deep water and its effects on the surface layer. Although the salinity turning points appeared in a very striking manner in the deep water already in 1993, they became visible in the surface layer of the central Baltic only recently. Their average salinity difference, together with the delay between the long-term minima discernible at both depths, could be used for estimating residence times: about 20 years for passive substances remaining below the halocline, and another 30 years for them to be washed out into the North Sea by the excess fresh water in the Baltic. As a result of the joint action of both processes, the surface salinity responds to deep water changes with a memory of 13 years. We have compared the actual events with interannual salinity and temperature anomalies on a longer timescale. With respect to certain climate indices, like the one proposed by Hagen & Feistel (2005) for the strengths of Baltic winters, these anomalies virtually lag behind by several years for reasons not yet understood thoroughly. Nonetheless, this fact reflects the well-known complexity and non-linearity of the Baltic Sea's internal response to external forcing.

5. Acknowledgements

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