Review Vertical mixing in the Baltic Sea and consequences for eutrophication – A review

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Abstract In the transition area between the North Sea and the Baltic Sea entrainment processes dominate the vertical mixing in the inflowing saline bottom water. The hot spots of these processes are located at the Darss Sill and the Bornholm Channel in the western Baltic Sea. In the central Baltic Sea the horizontal advection of saline water in deep layers below the permanent halocline dominate the temporal changes and associated transports. This is accompanied by the turbulent vertical transport through the halocline into the surface layers. The related vertical salt transport into the entire surface mixed layer estimated by various methods is slightly above 30 kg/(m² a). During stagnation periods, the residence time of the deep water in the Eastern Gotland Basin increases roughly by a factor of five. Vertical mixing through the halocline is drastically reduced when inflows are lacking, the potential processes of diapycnal mixing are discussed to the present knowledge. The turbulent motion resulting from breaking internal waves is capable of turbulent transports through the halocline corresponding to the estimates of the salt transport into the surface mixed layer. The actual knowledge about boundary mixing due to internal waves in the Baltic Sea is poor. Mesoscale eddies may contribute to the vertical mixing, but it is not known whether they really do and which of the possible direct and indirect mixing mechanisms is most effective. Near-bottom currents induced by inflow events likely enhance vertical mixing. Coastal upwelling certainly contributes to the vertical transport, but the depth of its origin and the volume transport are hard to determine for large-scale quantifications. The short spatiotemporal scale of turbulent transports through the halocline resulting in a weakening of the halocline during summer together with the mixing of the entire surface layer down to the halocline in winter form a consistent description of the vertical salt transport. It is hypothesised that the longer residence time of the deep water during stagnation periods results from the lack of energy imported by the inflows and directly or indirectly feeding the diapycnal mixing processes. The vertical transport of nutrients such as the phosphate is quantitatively not sufficiently understood and needs further interdisciplinary research activities.

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1. Introduction – General aspects of vertical mixing

Turbulent mixing plays a major role for the dynamics of marine ecosystems. For the surface mixed layer of the open ocean and shelf sea waters this has been illustrated first in the classical publication of Sverdrup (1953) where he showed that spring phytoplankton blooms are initiated when the surface mixed layer depth becomes smaller than a critical depth. This idea has been later refined by the concept of critical turbulence (Huisman et al., 1999) where the depth of the mixing layer (zone of active vertical mixing), in contrast to the mixed layer, is considered instead. Vertical mixing is also responsible for the maintenance of deep chlorophyll maxima, zones of relatively high phytoplankton abundance fuelled by upward turbulent nutrient fluxes within the thermocline below nutrient depleted surface water (Sharples et al., 2001).
In the Baltic Sea, a European semi-enclosed marginal sea (Fig. 1), turbulent mixing plays a much more complex role for the dynamics of the marine ecosystem. Due to its positive freshwater budget and the episodic events of salt water inflows through the narrow and shallow straits connecting the Baltic Sea with the saline North Sea, the central Baltic Sea is permanently stratified with a halocline located about 60 m below the surface. The Baltic Sea produces large amounts of organic matter sinking into the stratified deeper water where it mineralizes and thus lowers the oxygen concentration in the water. Major Baltic Inflows (MBIs) (Matthäus and Franck, 1992) occurring on the decadal time scale are the main process which ventilates these depleted deeper water. During the stagnation periods between two MBIs the near-bottom water of the deeper basins typically becomes anoxic with the consequence that large amounts of phosphate are released from the sediments. By means of complex and not fully understood vertical transport mechanisms this phosphate reaches the surface water of the Baltic Sea in winter, inducing there a nitrogen to phosphorus ratio significantly lower than the physiological Redfield ratio of 16. Once the spring phytoplankton blooms have depleted the near-surface nitrate, nitrogen fixing cyanobacteria take advantage of the excess phosphate, such that in warm and calm summers massive cyanobacteria blooms are characteristic for the Baltic Sea.

In many senses the processes leading to the upward transport of phosphate are non-trivial. In-situ turbulence measurements in the stratified water below the halocline of the Baltic Sea result in a vertical turbulent transport with eddy diffusivities of the order of $10^{-5} \text{m}^2\text{s}^{-1}$, whereas vertical eddy coefficients around $10^{-4} \text{m}^2\text{s}^{-1}$ are obtained by simple salt budget estimates. Thus other processes than internal mixing must take place in the Baltic Sea, the identification and quantification of which is a major open question in Baltic Sea research.

Stigebrandt (2003) has sketched a conceptual model for the vertical circulation in the central Baltic Sea. In Fig. 2 this overview is refined by listing all known relevant vertical mixing and transport processes in the Baltic Sea.

Episodic overflows over the Belt Sea sills transport saline water into the entrance areas of the Baltic Sea where they form dense bottom currents (Section 4.1). Those are subject to entrainment of less saline ambient water lowering their salinity and thus their density (Section 4.2). Once the density of the bottom currents equals the density of the ambient water (due to entrainment and deeper propagation) they interleave with the ambient water masses and ventilate them. Because of volume conservation the deep inflows generate a compensating uplift of the water column in the central Baltic Sea (Section 3.3). Vertical turbulent transport is enhanced by internal mixing due to inertial waves and internal wave breaking (Section 5.1), as well as by mesoscale Baltic Sea Ed-dies (the so-called Beddies) (Section 5.3). The latter may also contribute to boundary mixing when they propagate towards the sloping sea bed. Other potential boundary mixing effects are internal wave breaking at the sloping sea bed or shear-induced convec-tion (Section 5.2), or near-bottom currents induced by inflow events (Section 5.4). Coastal upwelling has the potential of transporting sub-halocline water towards the surface and irreversibly mixing it into less saline surface water due to differential advection (Section 5.5). During winter surface cooling will establish statically unstable stratification leading to convective entrainment of surface water into the halocline. Similarly, the surface wind stress causes shear-induced entrainment into the halocline. During summer the halocline is protected against entrainment from above by a seasonal thermocline below which internal-wave mixing erodes the halocline (Sections 2.2 and 3.2). Finally, surface waves generate near-surface turbulence and modify the surface mixed layer in a complex way. Breaking surface waves inject tur-
bubulence into the surface mixed layer and thus have a strong impact on the near-surface dynamics (field data by Terray et al. (1996) and model-data comparisons by Craig (1996) and Stips et al. (2005)). The interaction between wave-induced Stokes drift and mean shear generates pairs of counterrotating vortices aligned with the wind direction, the so-called Langmuir Circulation (LC). LC is strongly modified by turbulence and typically homogenizes the surface mixed layer (Large Eddy Simulation study by McWilliams et al. (1997) and the field data by Thorpe et al. (2003) and Gemmrich and Farmer (2004)). Since the mixing due the presence of surface waves is characteristic to all wind affected surfaces of natural waters, it will not be discussed in this review in further detail. Because of the character of the Baltic Sea as an almost enclosed marginal sea, tidal mixing does not play a significant role in the Baltic Sea.

The manuscript is organised as follows: after this brief introduction, the bathymetry and hydrography of the Baltic Sea is presented in Section 2 with special emphasis on inflow events in Section 2.1 and the surface mixed layer dynamics in Section 2.2. In Section 3 the vertical salt transport through the halocline is estimated by means of various methods, which are water and salt balances (Section 3.1), the quantification of pycnocline erosion (Section 3.2), the quantification of uplifts by inflows (Section 3.3), and the construction of an analytical diapycnal response function (Section 3.4). In Section 4 pathways of dense inflows in the western Baltic Sea are reconstructed (Section 4.1) and entrainment estimates are made for these inflows (Section 4.2). In Section 5 the potential mechanisms for vertical mixing in the Baltic Sea are discussed in detail as already listed above. The ecosystem perspective is presented in Section 6, mainly showing the responses of marine ecosystem parameters to inflow events. A final discussion of Baltic Sea vertical mixing in Section 7 concludes this review.

It should be noted that the unit g/kg for the salinity is used here. The numerical value of absolute salinity in grams of salt per kilogram of seawater is higher than Practical Salinity (PS-78) by about 0.5% (Jackett et al., 2006), which is an irrelevant deviation for most estimates made in this paper. For the calculation of the mass transport of salt, values of absolute salinity rather than Practical Salinity are required. The best estimate for the absolute salinity of Standard Seawater is the Reference-Composition Salinity (Miller et al., 2008). Salinities reported in g/kg in this paper are expressed in this new salinity scale.

# 2. Bathymetry and hydrography of the Baltic Sea

The Baltic Sea is a landlocked sea. The water exchange with the open ocean only takes place through the North Sea. The Baltic Sea is divided into different deep basins connected by narrow sills and channels (Fig. 1). The exchange transition area is the Kattegat (K). It is divided into different deep basins connected by narrow straits and the open ocean only takes place through the North Sea. The Baltic Sea is bounded by the Gulf of Bothnia which is separated from the central Baltic Sea or Baltic Proper by the Åland Sea (AS). The Gulf of Bothnia has a north–south extension with the Bothnian Sea (BS) in the south and the Bay of Bothnia (BoB) in the north.

Saline inflows from the North Sea produce a lateral surface salinity gradient throughout the whole Baltic Sea with high salinities of about 25 g/kg in the transition area of the Kattegat and low salinities around 5 g/kg in the Gulf of Bothnia. Compared to the open ocean and the North Sea the salinity in the Baltic Sea is generally low due to large amounts of fresh water provided by river discharges with an annual run-off of about 436 km³ resulting from the huge drainage area in conjunction with the humid climate. In addition, the annual water budget amounts to 224 km³ precipitation, 184 km³ evaporation, and 947 km³ surface water outflow (Brogmus, 1952; HELCOM, 1993) and is compensated by an appropriate inflow of nearly 500 km³ saline water from the North Sea. The water exchange between the Baltic Sea and the North Sea is restricted by the connecting belts and the Øresund, because they are particular narrow and shallow.

A permanent halocline separates the surface water from the deep water in the basins. The halocline in the AB is found in about 35 m to 40 m depth. In the EGB it is significantly deeper at 70 m to 90 m depth (Stigebrandt, 1987a; Elken, 1996). This prominent halocline is found throughout the southern and central Baltic Sea. During summer a seasonal thermocline develops at depths between 10 and 30 m (Matthäus, 1984). The thermocline disappears in winter due to cooling and mixing of the surface water by wind and convection. However, these processes do not affect the deep water which is isolated from the surface water by the halocline. When passing the sills, about 50% of the “old” Arkona deep water is replaced by inflowing dense water. During strong inflow events the exchanged volume of Arkona Deep Water exceeds 100 km³ (Omstedt and Axell, 1998). Because of its high density the inflowing saline water spreads along the bottom and is subject to further entrainment of brackish surface water on its way along the chain of deep basins (Köuts and Omstedt, 1993).

During stagnation periods without inflows of highly saline and oxygenated water from the North Sea, the oxygen in the Baltic deep water becomes depleted and hydrogen sulfide may occur. The thermohaline properties in the deep layers are mainly determined by advection. Vertical mixing occurs only by diffusion and turbulent exchange (Stigebrandt et al., 2002).

## 2.1. Inflow of high-salinity seawater from the North Sea into the Baltic Sea

Owing to an open connection to the North Sea and the Atlantic Ocean, the Kattegat contains seawater of high salinities with typically 17 g/kg at the surface and exceeding 30 g/kg in the deep layers. In contrast, the inner Baltic Sea has a fresh water excess causing climatological surface salinities up to 10 g/kg with strong vertical and lateral gradients (Feistel et al., 2008). The resulting climatological density differences in the Belt Sea drive a permanent near-bottom inflow of saltier and a near-surface outflow of more brackish water, severely hampered by the narrow and shallow Danish straits, their tidal oscillations and the permanently changing wind conditions in the westwind belt.
Only under specific meteorological circumstances, when either very strong or very calm wind conditions prevail for typically 20 days or longer, and the water column in the Belt Sea is either well-mixed or extremely strongly stratified, significant amounts of very salty water can penetrate into the inner Baltic Sea. Once it has reached the Arkona Basin it can propagate eastward across the sills and basins, substituting stagnating water masses there and replenishing the Baltic Sea salt reservoir. Such inflow events occur irregularly, from repeated events within a single year to stagnation periods lasting for a decade. The significant impact of such events on the physical, chemical, and biological status of the Baltic Sea has intensively been investigated in the past 50 years. A recent review on these studies was given by Matthäus (2006). The way these inflows depend on the global climatic change, and the complex non-linear cascades of processes they trigger are only partly understood yet.

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

- They are driven by barotropic pressure gradients, i.e. sea level differences.
- They appear during persistent westerly gales (mostly in autumn, winter, spring).
- They import salt (typically 2 Gt) into the Baltic Sea along with a water import of typically 200 km³.
- They import oxygen-saturated water (typically 1 Mt of O₂) in winter and spring.
- They pass through Öresund 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, 2004; Mohrholz et al., 2006):

- 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 Sea along with water volume export.
- They import oxygen-deficient water, but ventilate the deep Baltic basins by entrainment.
- They pass only through the Great Belt/Darss Sill gateway.

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

Matthäus and Franck (1992) identified 90 major inflow events during the 80-year period from 1897 to 1976, i.e. about one event per year on average. Since then the frequency of such events has dramatically decreased (Matthäus et al., 2008). The main index for estimating the strength of major inflows is the related total mass of imported salt: more than 3 Gt for very strong inflows, 2–3 Gt for strong and 1–2 Gt of salt for moderate inflows (Fischer and Matthäus, 1996; Feistel et al., 2003b).

As a prerequisite for a barotropic inflow of saline deep water into the EGB, easterly winds lasting for several weeks are needed to lower the sea level of the Baltic Proper by some 10 cm below normal (Matthäus and Franck, 1992). Afterwards strong winds from westerly directions must push saline North Sea water through the Belts and the Öresund into the south-west Baltic Sea. Usually, these conditions occur in late autumn, winter, or early spring when the saline water of the Kattegat is well ventilated due to strong mixing and cooled as well as the haline stratification is destroyed for a number of days. The westerly wind has to persist for 5 or more days to push enough water across the Dars and Drogden Sills, which then will interleave and mix with the “old” deep water in the AB. Only if this “new” deep water has a higher density than the “old” deep water in the subsequent basin, it can replace the deep water in that adjacent basin, steered by internal pressure gradients. After the inflow event, this renewal front takes at least three months to propagate into the central Baltic basins (Hagen and Feistel, 2001; Feistel et al., 2003b).

The volume capacity of the deep Baltic basins in connection with the mixing of the deep water into the intermediate water layer above controls the residence time of the deep water in the basins and thus their salt balances. For illustration, a volume of approximately 40 km³ fills the deepest layers of the central basin of the Baltic Sea, the EGB, between 190 m and 245 m depth (Hagen, 2004), typically containing 0.5 Gt of salt. During and after inflow events, large lateral pressure and salinity gradients between the central EGB with low pressure and its rim with high pressure are forcing the newly arriving saline water at the rim to flow along the bottom topography contour and circulate cyclonically within the basin (Lehmann and Hinrichsen, 2000). Peak velocities of 40 cm/s were measured, slowing down to 3 cm/s after a period of 3–6 months (Hagen and Feistel, 2004), thus providing the deep water with turbulent kinetic energy.

The Baltic Sea counterpart to the oceanic deep convection processes is the estuarine circulation. Random inflow events transfer dense water to the bottom layers and lift up the original ambient water. Slow vertical transport dilutes the bottom water subsequently and carries the salt into the brackish surface layer above the halocline, from where it is flushed into the Kattegat in the form of the Baltic Current. Thus inflow events are likely the key to the most important vertical transport mechanisms in the central Baltic Sea since they are the driving force of this “Baltic Conveyor Belt”. The estimated residence time of the Baltic deep water of the order of 20 years suggests that the interplay between inflow events and vertical transport is controlling the hydrophysical properties of the Baltic Sea on the time scale of decades (Meier and Kauker, 2003; Feistel et al., 2006a; Meier et al., 2006; Lass and Matthäus, 2008; Feistel et al., 2008; Hagen and Feistel, 2008).

2.2. Surface mixed layer dynamics

The bulk sea surface salinity in the Baltic Proper is determined by the freshwater fluxes into the surface layer, i.e. the net freshwater flux through the surface resulting from precipitation and evaporation, melting and freezing, as well as the lateral freshwater fluxes due to river run-off, and the rate of upward turbulent salt flux through the uppermost pycnocline. The latter is defined by the seasonal thermocline in summer and the permanent halocline in winter.

The summer and the winter halocline differ essentially as also discussed in Section 3.2: During summer the thermocline protects the halocline against erosion by surface mixed layer turbulence. Therefore the halocline is relatively weak in summer. In contrast, the absence of the thermocline in winter allows the surface mixed layer to reach down to the halocline, which is abraded by the associated turbulent mixing. In this way the halocline is restored and the vertical salinity gradients in the halocline reach their maximum.

Here we refine the picture of the seasonal near-surface cycle by numerical simulations with a high-resolution one-dimensional model. Such models are applicable in situations where vertical processes dominate over horizontal processes. This is the case for the Baltic Proper on the seasonal time scale (Section 3.1). The water column model applied here is the General Ocean Turbulence Model (Umlauf et al., 2005, www.gotm.net) which has successfully repro-
duced observations of near-surface mixing processes (Burchard and Bolding, 2001; Burchard et al., 2002). The setup used here is similar to a decadal simulation recently carried out for the Baltic Proper (Burchard et al., 2006). Here we present simulation results for the years 1989 and 1990, after the model has been spun up during the whole year 1988. The geographical position for the model simulation is at 20°E, 57.3°N in the centre of the Eastern Gotland Basin at a water depth of about 250 m. Only the upper 100 m of the water column are simulated. Prognostic equations are numerically approximated for momentum, potential temperature, salinity, and two turbulent properties, namely the turbulent kinetic energy $k$ and its dissipation rate $\epsilon$ (Burchard et al., 2006 for details of the setup). The estimate of 3 m/a for the vertical uplift at the 10 g/kg isohaline resulting from inflows has been used as the vertical advection velocity at 60 m below the surface (Section 3.3). This vertical velocity was set to zero at the surface and at 100 m depth for continuity reasons. From these two depths it was linearly interpolated to the value of 3 m/a at 60 m depth. With this setup a net

![Temperature in °C](image1)

![Salinity in g/kg](image2)

![Brunt-Väisälä frequency squared in s⁻²](image3)

![Eddy diffusivity in log₁₀(Kₑ/m²s⁻¹)](image4)

**Fig. 3.** Simulation results for potential temperature (upper left), salinity (upper right), squared Brunt-Väisälä frequency (lower left), and eddy diffusivity (lower right). Shown are results for the years 1989 and 1990 for the upper 80 m. For the squared Brunt-Väisälä frequency all values larger than 0.001 s⁻² are shown in red, all negative values are shown in black.

**Table 1**

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<td>12.74</td>
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</tbody>
</table>

Time series 2005 of salinity from the Gotland Deep measured by SMHI, expressed in the Reference Salinity Scale (Millero et al., 2008). The halocline, defined here by the strongest vertical gradient, is marked in bold. The virtually homogeneous surface layer is shown in italic on grey background. The steepest halocline of 2.13 (g/kg)/10 m is observed in January, together with the deepest extension of the mixed layer, down to the halocline. Quantity: Reference Salinity in g/kg Year: 2005 Station: BY 15 Lat: 57°20’ Lon: 20°03’.
surface freshwater flux of 1.26 m/a (partially accounting for unresolved lateral freshwater fluxes) was sufficient to balance the vertical salt flux into the surface mixed layer in such a way that the annual surface mixed layer salt budget was closed. The model resolution was 0.5 m in the vertical and 10 min in time. These choices ensured that all numerical schemes were converging and the simulation results were not significantly affected by numerical artefacts.

The results of this simulation for potential temperature, salinity, eddy diffusivity, and squared Brunt-Väisälä frequency are shown in Fig. 3. It should be noted that the graphics are composed of snapshots at midnight. The temperature shows a clear annual cycle with a summer thermocline at about 20 m depth, which builds up in mid April and is eroded down to 60 m in late December. During maximum surface warming the surface mixed layer depth is less than 10 m. The annual cycle of salinity is much weaker. It has a surface amplitude of about 0.3 g/kg. This is a 50% underestimation compared to the salinity amplitude given in Table 1. This disagreement may result from the seasonal variability in the freshwater flux which has been neglected in this simulation. During winter the surface mixed layer extends downwards into the main halocline, which leads to an erosion process with upward salt flux and increasing surface salinities during the winter season. Once a thermocline is established in spring the halocline is protected from surface mixing. However, internal mixing processes (Section 5.1), which are here parameterised by means of a minimum TKE value and the length scale limited by the Ozmidov scale, weaken the gradients in the halocline during the summer season. These process descriptions are strongly supported by the simulation results for the eddy diffusivity and the squared Brunt-Väisälä frequency. Directly below the surface mixed layer the squared Brunt-Väisälä frequency shows values of more than 0.001 s\(^{-2}\), specifically in the thermocline during summer and in the halocline during winter. The weakening of the summer halocline can be seen clearly. Near-surface stratification is mostly unstable, since all results are from midnight snapshots (“night convection”). Exceptions are the winter months, during which the temperature has fallen below the maximum density temperature, in particular during the winter 1988/1989. As a result further cooling at the surface stabilises the water column. Eddy diffusivity reaches values of up to 0.1 m\(^2\)/s during winter cooling. Interestingly, values of 2–3 \(\times\) 10\(^{-6}\) m\(^2\)/s are simulated in the halocline, which is consistent with the observations discussed in Section 5.1.

Comparisons between these simulation results and observations of temperature and salinity (not shown here) show that the simulated mixed layer depths are generally too shallow, which may be a result of missing horizontal processes. However, this simulation qualitatively reproduces the relevant processes in the surface mixed layer of the Baltic Proper. Detailed observational studies of these dynamics including turbulence measurements would be required in order to better quantify near-surface mixing.

3. Vertical salt transport through the permanent halocline

This section has the focus on the salt balance in the central Baltic Sea and the associated vertical transport of salt to provide an easy measure and an estimate of the order of magnitude of any upward tracer transport through the permanent halocline which is connected to this by the corresponding volume transport. The estimates are made by several different approaches to form a consistent picture. Some of them are independent of the detailed mixing processes, but some others are already connected to certain transport mechanisms, mainly the erosion of the pycnocline by deep convection in winter.

3.1. Bulk estimates from water and salt balance

Since the only source of sea salt in the Baltic Sea is the input from the Kattegat/North Sea in the form of episodic injections into layers beneath the permanent pycnocline, and its only sink is the near-surface outflow of brackish water into the Kattegat, the salinity above the pycnocline is a robust integral measure of the net upward transport through the pycnocline. The related bulk parameters have been known for more than a century (Knudsen, 1900) and have since then been discussed by various scientists (Fonselius, 1962; Jacobsen, 1980; Matthäus and Schinke, 1999; Rodhe and Winsor, 2002; Fonselius and Valderrama, 2003; Meier and Kauker, 2003; Stigebrandt and Gustafsson, 2003; Meier et al., 2006).

The climatologic surface salinity 1900–2005 of the Baltic Sea is available from e.g. the Baltic Atlas of Long-Term Inventory and Climatology, BALITIC (Feistel, 2006; Feistel et al., 2008). During the residence time of about 30 years of the riverine water at the surface, the slow but perpetual entrainment of salt from the layers beneath is forming the well-known NE–SW surface salinity gradient (Meier and Kauker, 2003; Meier et al., 2006; Feistel et al., 2006a). The freshwater input of about 500 km\(^3\)/a causes a comparable export of brackish water (Knudsen, 1900; Matthäus, 2006) with a salinity of about 8 g/kg. Thus the Baltic Sea exports about 4 Gt of salt per year, and imports the same amount on average (Feistel and Feistel, 2006). With regard to the surface area of 130 000 km\(^2\) at 60 m depth, the mean diapycnal salt transport rate is about \(J = 30\) kg/(m\(^2\) a). This mass of salt originates from the Kattegat, imported with a mean inflow rate of another 500 km\(^3\)/a, balanced by the same outflow volume with lower salinity. The seasonal signal of this exchange flow is strongest in the Belt Sea with an amplitude of 4 g/kg in the surface salinity off Warnemünde, and is observed up to the Bornholm Basin with a difference of 0.2 g/kg between February and August (Matthäus, 1978).

The climatologic salinity distribution in the surface water is suggesting further useful information about the lowest-order dynamical processes in the Baltic Sea. If regions not too close to the river mouths or the Danish Straits are considered and evaporation and precipitation are balanced, salt is transported only upward from the deep to the surface water, and the salinity of a particular parcel in the surface layer can only increase in time. Under these assumptions, salinity is a measure of the surface water age, and the lateral salinity gradient reflects the climatologic advection trajectories. The water from the river Neva takes about 30 years to cross the Baltic Sea over a distance of 1500 km (Meier et al., 2006; Feistel et al., 2006a), corresponding to an average advection speed of 0.2 cm/s. For the typical BALITIC atlas grid cells with dimensions of about 100 km \(\times\) 50 km (Feistel et al., 2008), the related residence time of surface water is 1–2 years. This estimate justifies the dynamical consideration of a particular 1\(^1\) \(\times\) 1\(^1\) water column in terms of a one-dimensional vertical model on the monthly to seasonal time scale, thereby ignoring the influence of lateral advection. If, alternatively, surface advection velocities in the Baltic Sea were significantly higher than estimated here and, say, forming a fast gyre circulation on the basin scale, the observed lateral salinity gradient would not persist in a stable manner.

Near the bottom, in contrast, current velocities of the order of 3 cm/s are measured in the central Baltic Sea even in the absence of inflow processes (Hagen and Feistel, 2004). Thus seasonal changes in the deep water of an 1\(^1\) \(\times\) 1\(^1\) water column are likely to be influenced by advective exchange processes between adjacent cells. Velocities are significantly higher during inflow events. The leading front of a typical inflow event, like 1997 or 2003, travels the distance of 800 km from the Danish Straits to the Gotland Basin within 3–4 months, i.e. with an average speed of 10 cm/s (Hagen and Feistel, 2001; Feistel et al., 2003b; Hagen and Plüschke, 2009).
Summarising the estimates made before, temporal changes on the monthly scale encountered in a selected water column of the size $1^\circ \times 1^\circ$ are likely attributed to horizontal advection in the near-bottom water but rather to vertical exchange in the surface water. This working hypothesis is borne in mind in the following when we shall discuss certain observations and possible explanations thereof.

3.2. Erosion of the pycnocline

The time series measured at the Gotland Deep in 2005 by the Swedish SMHI may serve as an instructive example for the seasonal evolution of the halocline (Table 1). The year 2005 was a year of very weak inflow activity, with only decaying remnants of the inflow events observed in 2002 and 2003 (Nausch et al., 2006; Feistel et al., 2006a). Levels below 100 m are unaffected by the seasonal cycle. The pycnocline, defined by the strongest vertical gradient, is weak during the existence of the summer thermocline (May–September) and is strong in winter (January–February). The steepness is correlated with the homogeneity of the surface layer, i.e. the winter convection, but also with surface salinity, which rises in fall/winter and falls in spring/summer.

A likely explanation of this observation can be that in summer the pycnocline becomes weakened under the protection of the thermocline, while the winter convection erodes its uppermost part and distributes the salt enclosed over the whole surface layer, restoring the steep density leap at its lower boundary (Section 2.2). Between the minimum surface salinity of 6.85 g/kg in August 2005 and the maximum of 7.46 g/kg in February 2006, we find a difference of 0.61 g/kg. The corresponding difference between minimum and maximum of the mean seasonal cycle is 0.52 g/kg (Matthäus, 2005) and 7.46 g/kg in February. In reality, this process is likely more constrained to the pathway of the inflowing tongue, rather than being spread homogeneously over the whole Baltic Sea area.

The strongest perturbations on timescales between hours and years are Major Baltic Inflows (MBI) from the Kattegat. During an event lasting 10–20 days, typically 200 km$^3$ of water flow into the Baltic Sea, importing about 2 Gt of salt into the deep water (Fischer and Matthäus, 1996; Feistel et al., 2003c, 2006a; Matthäus, 2006; Matthäus et al., 2008). The near-bottom salinity in the Arkona Sea during an inflow event is 20 g/kg or higher, i.e. only about 100 km$^3$ of the inflow volume are injected into the deep water below the pycnocline. Independent of future mixing processes on its way to the central basins, this volume elevates the pycnocline with its surface area of 130,000 km$^2$ by 80 cm on average. This is a rather small number compared to the typical lowering of the halocline in winter by some 10 m in the Gotland Basin. If this uplifted layer, assuming its typical salinity as $S = 10$ g/kg, is eroded during the vertical convection in the following winter(s), 1 Gt of salt becomes entrained and mixed into the surface layer with its total volume of about 15,000 km$^3$, thus increasing its salinity by less than 0.1 g/kg on average. In reality, this process is likely more constrained to the pathway of the inflowing tongue, rather than being spread homogeneously over the whole Baltic Sea area.

Within the deep basins the uplift of ambient water masses by the newly arriving one is much more pronounced than it is observed at the pycnocline. During the inflow of September 1997, for example, estimated 56 km$^3$ of the inflow water with a salinity of about 12 g/kg arrived at the Eastern Gotland Basin, entirely substituting the residing water volume of 38 km$^3$ beneath 150 m depth there, and pushing it into the adjacent basins further north and west (Hagen and Feistel, 2001). The original volume increase was estimated to 140 km$^3$ from the sealevel rise during the inflow process (Hagen and Feistel, 2001), its salt import to 1.2 Gt (Matthäus, 2006).

The pycnocline uplift during the MBI 2003 as well as its seasonal erosion is well seen from the distribution of hydrogen sulfide as a tracer in the ambient deep water in Fig. 30. The oxygen-rich inflow near the bottom (green, beginning in April/May 2003) causes an elevation of the stagnant water (yellow) below the pycnocline from 100 m to 70 m depth, i.e. by almost 30 m. Subsequent erosion, in particular between February and April, gradually suppresses the excited vertical oscillation of the pycnocline.

The salt content of the entire Baltic Sea is about 120 Gt (Wingsor et al., 2001; Rodhe and Wingsor, 2002), which rises by about 1–5% in the course of a strong MBI event, but shrinks by 4% regularly every year due to the outflow to the Kattegat (Feistel and Feistel, 2006). In order to substitute the mean annual surface salt loss of 4 Gt/a by an equivalent supply from the deep water, the required apparent mean vertical advection speed at the $S = 10$ g/kg isohaline would be $w = 3$ m/a on average.

3.4. Diapycnal response function

The temporal development of the salinity at the surface (above 20 m) and in the deep water (at 200 m) between 1968 and 2005 at the Gotland Deep is shown in Fig. 4. In the upper series, inflow events are visible by the salinity leaps, followed by a slow decay. This decay is particularly apparent in the long stagnation period from 1978 to 1993. In the lower series, the readings scatter stronger, but reflect the deep-water record with a delay of about 10 years, but with smaller amplitude and temporally smoothed. From the simple linear 2-box vertical exchange model,

$$\frac{\partial S}{\partial t} = D \times [s_d(t) - S]$$

(1)

describing the “answer” $S(t)$ of the surface salinity to the “signal” $s_d(t)$ of the bottom salinity (Feistel et al., 2006a), a linear response function can be derived by solving the differential Eq. (1), as

$$S(t) = D \int_0^\infty \exp(-\tau T_E) s_d(t-\tau) \, d\tau$$

(2)

The memory, $T_E = 13a$, and the deep-water residence time, $1/D = 21a$, can be estimated from mean salinity differences and the signal delay between the two depth levels as shown in Fig. 4 (Feistel et al., 2006a). The related residence time of the surface water

![Fig. 4. Salinity at the surface (above 20 m) and in the deep water (at 200 m) between 1968 and 2005 at the Gotland Deep.](image-url)
appears to be 33 years. These values agree very well with estimates derived earlier by several workers from entirely different arguments, e.g. by modeling Eulerian tracers or water age diffusion (Meier and Kauker, 2003; Meier et al., 2006; Meier, 2007).

None the less, the model according to Eq. (2) is an extreme simplification of the real complexity of vertical exchange. It is derived from the temporal behaviour at merely a single location, but its inherent time constants correspond to basin-scale processes of the whole Baltic Sea, which comprise dynamically so different regions as the Belt Sea or the Bothnian Bay. Another deficiency of the model is the fact that it predicts an exponential decay of the bottom salinity during stagnation periods, i.e. without salt inflow \( \Phi_B(t) = 0 \), from the complementary balance relation to Eq. (1)

\[
\frac{d S_B}{dt} = \Phi_S(t) - D \times S_B
\]  

(3)

As Fig. 4 clearly shows, the curve at 200 m between 1983 and 1993 is concave rather than convex, i.e. the salt loss is accelerating rather than slowing down. A possible explanation is that the salt depletion in the deep water reduces the density gradient at the pycnocline, thus permitting the winter convection to gradually deepen its erosion of the ”lid” (Feistel et al., 2006b). Such a self-accelerated decay ends in a total destruction of the halocline already after a finite period of time, in contrast to an exponential law, which takes infinitely long to let the stratification disappear.

The mean vertical transport rate \( D \), corresponding to a deep-water residence time of \( 1/D = 21 \) years, includes the occasional up-lift processes of deep water by inflow events as shown in Fig. 4. In the absence of the latter ones, as observed during the long stagnation period, the typical salinity decrease at 200 m is about 1.5 g/kg 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 residence times suggest that deep water uplift during inflow events and the subsequent erosion of the pycnocline is by far the dominating vertical transport mechanism in the Baltic Sea (Feistel et al., 2006a, 2008).

The related transport rate across the pycnocline into the surface layer of about 60 m thickness can correspondingly be computed from Eq. (1) as a salt flow of

\[
J = D \times S_B \times \rho \times 60m = 34 kg/(m^2 a).
\]  

(4)

in excellent agreement with the estimates derived in Sections 3.1 and 3.2 from rather different models. During the Great Stagnation Period 1983–93 without major inflows, this value appeared reduced by a factor of 5 due to the longer residence time of about \( 1/D = 100 \) a used in Eq. (4), so that probably 80% of the long-term vertical salt transport originates from occasional inflow processes and the subsequent elevation and erosion of the halocline, and only less than 10 kg/(m² a) are due to other mechanisms.

A similar estimate was made earlier by Ozmidov (1994a), who took a deep-water volume of 200 km³ with salinity 20 g/kg, possessing a (guessed) residence time of 10 years beneath the thermohalocline with a surface area of 80,000 km², and obtained a vertical salt flow of \( J = 5 \) kg/(m² a). Alternatively, derived from a diffusion equation approach to the vertical salinity profile measured in the Eastern Gotland Basin in April 1993, the vertical salt transport estimated by Ozmidov (1994a) is about \( J = 1 \) kg/(m² a), i.e. even smaller. Evidently, the related vertical salt transport of 0.4 Gt/a estimated by Ozmidov (1994a) can explain only 10% of the observed salt export of 4 Gt/a by the brackish Baltic Current.

We finally note that the vertical salt flow of \( J = 34 \) kg/(m² a) amounts to a related potential power density of \( P = g \times J = 11 \) µW/m², corresponding to the potential energy stored in the uplifted mass of dissolved salt in excess to the ambient mass of water.

4. Mixing in straits

This section focuses on the entrance area of the Baltic Sea addressing the mixing and dilution of inflowing saline bottom water on its way through channels and over sills to the central parts of the Baltic Sea. The amount of dense water arriving in the deep basins of the Baltic Sea depends on the net amount of dense water flowing over Darss Sill (DaS) and Drogden Sill (DrS) into the Arkona Sea (Fig. 5) and on the rate of entrainment of brackish water into the inflow water along the pathway of the dense water towards the Baltic Proper.

Considering the inflow along a chain of subsequent sills and basins, starting from the Kattegat, through the Belt Sea into the Arkona Sea, through the Bornholm Channel (BCh) into the Bornholm Sea and furthermore through the Słupsk Furrow into the Eastern Gotland Basin, Köuts and Omstedt (1993) estimated the amount of volume flow increase due to entrainment for each of these regions. Analysing salinity and temperature observations during two decades, they came up with 79% for the Belt Sea, 53% for the Arkona Sea and 28% for the Słupsk Furrow. For the Bornholm Channel and the Bornholm Sea no mixing was identified.

4.1. Dense water pathways

In order to obtain a more detailed picture of the regional and temporal distribution of entrainment and a deeper understanding of the underlying processes, the pathways of dense inflowing water along the major chain of subsequent straits and basins need

Fig. 5. Detailed map of the western Baltic Sea covering the Arkona Basin as indicated on the left side (map A). The arrows in map B on the right side indicate the direction to the Darss Sill (DaS) and Drogden Sill (DrS). Indicated in red are the transects north of Kriegers Flak (RF) and across the Bornholm Channel (BCh), shown in Figs. 7 and 8, respectively. The bathymetry is drawn in contours with a spacing of 5 m, the inner contour in the Arkona Basin is at 45 m depth.
to be identified. Because of the episodic nature of these inflows and the general undersampling problem, this is actually not an easy task. The classical picture is that the dense bottom currents adjust to a major balance between Coriolis force and pressure gradient, whereas bottom and interfacial friction are of secondary importance (Liljebladh and Stigebrandt, 1996). This has the consequence that water masses spilling over the Darss and Drogden Sills are basically spiralling cyclonically along the rims of the Arkona Basin (Fig. 5), dragged by some small frictional effects into the core of the Arkona Basin (Lass and Mohrholz, 2003). These feed episodically into an up to 15 m thick bottom pool of dense water resting in the relatively plain eastern part of the Arkona Sea (Stigebrandt, 1987b), generating a Kelvin-wave type cyclonic circulation pattern (Lass and Mohrholz, 2003). This pool has a residence time of 1 to 3 months (Lass et al., 2005), and is reduced in volume by permanent leakage through the Bornholm Channel (Walin, 1981).

This view of cyclonically spiralling dense bottom currents into the Arkona Sea has recently been challenged by means of numerical modelling (Burchard et al., 2005; Lass et al., 2005; Burchard et al., 2009) and field observations (Sellschopp et al., 2006; Umlauf et al., 2007). Burchard et al. (2005) carried out a simple numerical lock exchange experiment for the Arkona Sea, using the three-dimensional hydrostatic model GETM (General Estuarine Transport Model, Burchard and Bolding, 2002). The Arkona Sea was initially filled with brackish water of homogeneous salinity of 8 g/kg. The model was then forced with elevated water levels and high salinity of 25 g/kg in the Øresund, leading to a strong overflow over Drogden Sill. On its southward propagation the plume turned eastwards and passed along the northward slope of Kriegers Flak (KF) (Fig. 5). Only a small fraction of the plume followed the expected pathway west of Kriegers Flak, thus contradicting the earlier assumptions of basically geostrophically balanced inflows. The separated dense bottom currents joined south-east of Kriegers Flak and meandered into the eastern part of the Arkona Sea. This was independently confirmed by another idealised numerical model study carried out by Lass et al. (2005), based on the Modular Ocean Model (MOM-3, Pacanowski and Griffies, 1999) applied to a numerical model of the North Sea and Baltic Sea. There an inflow event was forced by sea surface elevation gradients and westerly winds, resulting first in an inflow through the Øresund with similar dynamics as obtained by Burchard et al. (2005). About 5 days after Drogden Sill is passed by saline water a saline overflow occurs at Dars Sill, which subsequently runs down a submarine terrace, flowing cyclonically along the southern rim of the Arkona Basin, and stratifying over the denser plume from the Øresund. Parts of these rather complex dynamics have been confirmed by Sellschopp et al. (2006) who observed an inflow event through the Øresund in Jan./Feb. 2004. They found that most of the dense water flowing over Drogden Sill did afterwards pass through the channel north of Kriegers Flak. Burchard et al. (2009) set up a realistic numerical model for the western Baltic Sea covering the period of observations carried out by Sellschopp et al. (2006). They obtained good agreement between the observed and simulated salinities at the positions of the Drogden Sill, Dars Sill and Arkona Sea moored stations (Fig. 6) and between the observed and simulated salinities and currents north of Kriegers Flak. With such a validated numerical model various analyses of derived properties, which cannot be directly quantified by observations, can be carried out as discussed in the next subsection.

These observations were further confirmed by Umlauf et al. (2007) who observed a strong inflow over Drogden Sill via the channel north of Kriegers Flak in November 2005 with improved instrumentation at high spatial resolution. During several transects they found a distinct cross-sectional structure of the dense bottom current with a strong pycnocline leaning towards Kriegers Flak, with the dense core moved by the cross-channel Ekman-transport towards the northern slopes of the channel, a characteristic transverse circulation, and a complex pattern of high and low mixing (Fig. 7). This pattern is repeated across the Bornholm Channel (Fig. 8). For the Słupsk Sill this characteristic transverse structure of dense overflows into the Baltic Sea has also been observed by Paka et al. (1998, 2006). They attributed the dislocation of the dense core to the north and the vertically erected isopycnals near the bed of the southern slope to the mechanism of arrested Ekman transport.
layers (Garrett et al., 1993), balancing the southward near-bed Coriolis acceleration by the transverse density gradient. The possible impacts of these dynamic features on entrainment are discussed in the next subsection.

The outflow from the Bornholm Sea over the Słupsk Sill into the Słupsk Furrow shows a high variability as well. Meier et al. (2006) list three different overflow regimes, (i) simple overflow when isopycnals in the Bornholm Sea are higher than the sill depth, (ii) overflow with strong shift of the dense bottom current core towards the southern slopes of the sill, and (iii) eddy-like episodic exchange across the sill. Each of these overflow regimes may be subject to various mixing and entrainment processes which are not yet well investigated.

4.2. Entrainment

The data analysis of Köuts and Omstedt (1993) located the regions of increased mixing of inflows into the Baltic Sea, but did not discuss the processes causing the mixing. In a more detailed and process-oriented investigation Lass and Mohrholz (2003) identified three major mixing mechanisms, (i) wind mixing in the vicinity of sills, (ii) differential advection in the head region of the dense bottom currents, and (iii) shear-induced entrainment of ambient water into dense bottom currents. However, the wind related mixing can only be effective at relatively shallow depths, such as the Darss and Drogden sills, whereas in the deeper Bornholm Channel and Słupsk Furrow the dense bottom currents are mostly protected against wind mixing by an overlaying stratified layer. The differential advection mechanism which increases mixing by shearing denser water over less dense water in the presence of density increasing in flow direction has been suggested earlier by van Aken (1986).

The entrainment across the pycnocline on top of the dense bottom current due to interfacial shear has been investigated in much detail since several decades (Turner, 1986; Baines, 2001; Cenedese et al., 2004). This generally results in a dependence of the entrainment parameter $E = W_E/U$ (with the entrainment velocity $W_E$ representing the rising velocity of the pycnocline and the mean downslope velocity $U$) on the Froude number $Fr = U/(gH)^{1/2}$ (with the reduced gravity $g$ and plume thickness $H$), which is of the form

$$E = cFr^a$$

with the positive non-dimensional parameters $a$ and $c$. The entrainment formulation derived by Stigebrandt (1987a), based on the
Kato-Phillips entrainment formula (Kato and Phillips, 1969) and calibration of a horizontally integrated numerical model, parameterises this as

\[ E = 2C_d R_f F \]

using the notation of Arneborg et al. (2007), with the bed friction coefficient \( C_d \) and the bulk flux Richardson number \( R_f \), which is the ratio of the buoyancy flux and the shear production of turbulence. Stigebrandt (1987a) applied this with constant values of \( C_d = 0.0035 \) and \( R_f = 0.035 \). By validating the turbulence model GOTM (General Ocean Turbulence Model, Umlauf and Burchard, 2005) with time series observations of current, salinity, and turbulence dissipation rate in a dense bottom current north of Kriegers Flak, Arneborg et al. (2007) refined the entrainment formula by considering the effect of the Ekman number \( K = \frac{(C_d U)}{(f H)} \) (with the Coriolis parameter \( f \)) and the Froude number \( F_r \) on the bulk flux Richardson number \( R_f \):

\[ E = aC_d K^b F_r^c \]

with \( a = 0.084 \), \( b = 0.6 \) and \( c = 2.65 \).

Based on mass conservation (Sellschopp et al. 2006) estimated the entrainment velocity between Drogden Sill and the region north of Kriegers Flak to \( \bar{w}_E = 3 \times 10^{-6} \) m/s. Inserting the observed values of \( C_d = 0.0037 \), \( U = 0.5 \) m/s, \( f = 1.2 \times 10^{-4} \) s\(^{-1} \), \( H = 10 \) m and \( g = 0.089 \) m/s\(^2 \) into the formulation Eq. (5) by Arneborg et al. (2007), which gives a Froude number of \( F_r = 0.53 \) and an Ekman number of \( K = 1.54 \), the entrainment velocity is \( \bar{w}_E = 3.75 \times 10^{-5} \) m/s, about one order of magnitude higher than the estimate derived from a larger scale spatial average mass conservation consideration. This demonstrates how highly variable in time and space entrainment may be (Fig. 9).

One complication is the complex transverse structure of dense bottom currents as shown by Umlauf et al. (2007) (Fig. 7). The dissipation rate, representing the mixing intensity, shows four clearly distinct regions. There are strongly elevated levels of dissipation rate in the bottom boundary layer of the dense bottom current, at the left side where vertical isohalines are advected towards the center of the channel by the Ekman-transport associated with the strong bottom turbulence, and in the sharp pycnocline itself where the local shear has a maximum. In contrast to that, a region of low dissipation rate is visible in the core of the dense bottom current.

Observations from the Bornholm Channel in November 2005 clearly show a similar pattern (Fig. 8). From this we conclude that substantial mixing and subsequent entrainment must also be present at the Bornholm Channel. For the Słupsk Sill substantial mixing can be inferred from detailed hydrographic surveys (Piechura and Beszczynska-Möller, 2004; Paka et al., 2006).

Because of the high spatial and temporal variability of entrainment processes and the small value of the entrainment velocity, mixing in straits is substantially undersampled. A successful method to still obtain quantitative estimates of mixing in straits over long periods, is the application of high-resolution numerical models for these regions. Arneborg et al. (2007) have already demonstrated this for the position north of Kriegers Flak with a one-dimensional example. It is essential that these models are able to reproduce the basic processes associated with mixing and the few observations which are available. Using the same turbulence module as Arneborg et al. (2007), Burchard et al. (2009) have set up the GETM model for the western Baltic Sea and Kattegat region for the period September 2003–May 2004. They could reproduce time series observations at the Darss and Drogden sill stations.

Fig. 9. Decadal logarithm of simulated vertically integrated, time averaged turbulent vertical salt flux in \([m^2 g/kg]/s\) in the Arkona Sea area for the period September 2003–May 2004 taken from Burchard et al. (2009). Isobaths are indicated as contours.
and in the Arkona Sea as already shown in Fig. 6. The vertically integrated, time averaged turbulent vertical salt flux has been calculated utilising this validated model (Fig. 9). Areas of strong mixing can be clearly identified in the pathways of dense bottom currents through the Øresund towards north of Kriegers Flak and through Darss Sill as well as over the submarine terrace east of it. Moreover, the strongest salt fluxes are observed across the Bornholm Channel. Effective mixing rates calculated by the model may however be artificially increased due to numerical mixing (Burchard and Rennau, 2008) such that some attention is necessary when analysing mixing rates from numerical models.

The strong mixing observed in the Bornholm Channel is in disagreement with the findings of Kõuts and Omstedt (1993), who concluded that mixing in the Bornholm Channel is irrelevant. How can this be explained? A closer look at the dissipation rates in Fig. 7 shows that increased dissipation is present in the region of the halocline, located above the quiet core of the plume with minimum dissipation rates. Thus the entrainment cannot be mediated through bottom friction, but can be fully explained by interfacial instabilities. Although the mixing effect of this entrainment is substantial, the dynamic effect on the dense bottom current due to the entrainment stress may be small. Thus the fact that the dense bottom current is basically in geostrophic balance (Kõuts and Omstedt, 1993) does not contradict to the fact that entrainment is large. Actually, Sellschopp et al. (2006) could show that the shape of the density profiles north of Kriegers Flak can be largely explained by the thermal wind equation, even though the same data imply strong entrainment at this position at the same time (Arneborg et al., 2007).

Mixing in the Słupsk Furrow has been estimated to be substantial by Kõuts and Omstedt (1993). Although direct observations of entrainment activities in this region have not been carried out, this high entrainment is indicated by the fact that the Słupsk Furrow near-bottom areas are almost always well ventilated with oxygen while the water masses are mostly poor in oxygen at the same depths in the Bornholm Sea (Feistel et al., 2006a). This can be explained by oxygen-rich surface water being entrained into the oxygen-depleted dense bottom currents during the passage over the Słupsk Sill.

The natural mixing processes in the western Baltic Sea may actually be modified by massive off-shore constructions such as bridges and wind farms. During the planning phase of the Great Belt Link and the Øresund Bridge a zero-blocking solution has been postulated, meaning that the bridge constructions must have no influence on the volume flux through the Øresund and Great Belt (Hansen and Møller, 1990). Stigebrandt (1992) estimated that a flow reduction by only 0.6% would result for the Øresund Bridge with almost no measurable effect for the Baltic Sea hydrography. Møller et al. (1997) carried out laboratory experiments with obstacles dragged through stratified water and found no significant mixing effect. However, recent observations of currents, stratification, and turbulence upstream and downstream from the western part of the Great Belt Bridge carried out by Lass et al. (2008) revealed significant effects of the bridge piles on the vertical flow structure. They observed von Karman straits in the wake of the bridge piles with the potential to generate internal waves. These may propagate away from the constructions and release their energy into diapycnal mixing remotely (Section 5.1). In recent years extensive off-shore wind farms have been planned in the western Baltic Sea. Many of them are located near the coast at shallow depths, where dense bottom currents do not occur, but it is the intention to locate some of these projected wind farms in overflow areas such as around Kriegers Flak. Their foundations may also act as obstacles for dense bottom currents with the effect of increased entrainment. However, bulk estimates of the mixing effects of these existing or projected constructions and their potential impacts on the ecosystem of the Baltic Sea have not been carried out so far.

5. Other potential mechanisms of vertical mixing

After the erosion of the pycnocline by deep convection in winter seems to be identified as the main vertical transport mechanism from the permanent halocline into the entire mixed layer in the central Baltic Sea in Section 3, this section is devoted to some other vertical mixing mechanisms which are assumed to be relevant in the Baltic Sea. According to Section 3, they may gain in importance temporarily during stagnation periods or on timescales shorter than one year. Furthermore, they may have regionally or at depth levels other than the halocline significant impact. The knowledge about the mechanisms discussed in this section is quiet different. It ranges from quiet well known to completely unexplored for the Baltic Sea. In addition, it has to be mentioned that they are not necessarily independent from each other or may respond to the same forcing conditions.

5.1. Mixing due to inertial waves and internal wave breaking

5.1.1. Dissipation of wave energy to turbulent motion and mixing

Lass et al. (2003) have shown that there exists a well defined internal turbulence regime in the Baltic Sea which is embedded between the surface layer and the bottom layer turbulence regime. Turbulence below the surface mixed layer is independent of the actual wind forcing at the sea surface and occurs in intermittent patches due to pelagic and benthic disintegration of internal waves that are sub-grid phenomena in most models. The general view has been that non-linear interaction of internal waves cascades energy to small scales where it subsequently supports turbulence in the ocean interior. This can be quantified in different ways depending on the assumptions made about the wave field and about the nature of the interactions. The Garrett–Munk model of the internal wave field (Garrett and Munk, 1975) has been used for most discussions of open ocean conditions. Predictions of the energy contributed by the internal wave field to turbulence and ultimately lost to dissipation in the course of mixing the fluid were reviewed by Gregg (1989).

There are areas in the ocean and in marginal seas where the internal wave field deviates from the Garrett–Munk model. The internal wave field in the Baltic Sea deviates in several aspects from that of the ocean. Firstly, internal tides are lacking in the Baltic Proper due to the virtual absence of barotropic tides. Secondly, there are no permanent geostrophic currents in the Baltic Sea whose temporal variations are associated with the radiation of inertial waves due to the adjustment to the varying geostrophic balance (Gill, 1982). The dominant process of the generation of internal waves in an enclosed sea is expected to be the radiation of barotropic inertial waves generated at the coasts in order to maintain the zero flux boundary condition in the presence of fluctuating Ekman transports in the surface layer of the open sea. Barotropic inertial waves passing stratified water on a sloping bottom generate baroclinic inertial waves. This may explain why baroclinic inertial motions are so energetic in the Baltic Sea (Gustafsson and Kullenberg, 1936; Kielmann et al., 1973). Although the shape of the internal wave spectrum in the Eastern Gotland Basin is similar to that of the Garrett–Munk spectrum in the frequency space, Lass et al. (2003) found that the internal waves below the halocline were characterised by a dominance of waves with upward phase propagation. This implies that the spectrum of internal waves in the Baltic Sea can not be modelled by the Garrett–Munk spectrum. Polzin et al. (1995) were able to find a scaling of the dissipation rate of turbulent kinetic energy in terms of the frequency distribu-
tion of energy within the deep-ocean internal wave field for wave fields that differed from the Garrett–Munk model.

Wave–wave interaction parameterisation assumes that the energy flux is directed from the large energy containing waves via weak non-linear interaction toward the small scale waves, which become unstable and finally break into turbulent motion. The rate $\varepsilon$ of dissipation of turbulent kinetic energy depends on the energy flux through the spectrum of the internal waves and depends on its parameters. The parameterisation of Henyey et al. (1986), Gregg (1989), Polzin et al. (1995), which is based on the wave–wave interaction mechanism, predicts $\varepsilon \propto N^2$ with the Brunt-Väisälä frequency $N$ and has been successfully compared to dissipation measurements in the open ocean (Polzin et al., 1995).

Using WKB scaling of the internal wave velocities, Gargett and Holloway (1984) have shown that the dependence of the dissipation rate depends on properties of the internal wave spectrum. In case of narrow band internal waves the averaged kinetic energy production term and hence the dissipation rate in the stratified ocean is $\varepsilon \propto N$. In the broad band case (GM-spectrum) it is $\varepsilon \propto N^3$. In the ocean thermocline averaged dissipation rates are reported to vary systematically with $N$ between $\varepsilon \propto N$ and $\varepsilon \propto N^2$ (Gregg, 1987; Polzin et al., 1995). MacKinnon and Gregg (2003, 2005) found an $\varepsilon \propto N$ scaling for data collected over the continental slope. The parameterisation $\varepsilon \propto N$ was also successfully used in a vertical-circulation model for the deep water of the Baltic Sea (Stigebrandt, 1987a). Lass et al. (2003) verified this parameterisation by direct dissipation measurements in the Eastern Gotland Basin and found evidence that the proportionality factor is the sum of the kinetic and potential internal wave energy, i.e.

$$\varepsilon = \alpha (E_{\text{kin}} + E_{\text{pot}}) N$$

with $\alpha = 0.001$.

This result suggests that the dissipation of kinetic energy in the stratified layers may vary in space and time as the total energy of the internal wave field varies in the Baltic Sea.

5.1.2. Vertical mixing in the interior

Estimates of the vertical turbulent diffusivity in the Baltic Sea have been performed mainly with three different methods.

Beneath the halocline the Baltic Sea acts like a filling-box system maintained by horizontal advection from inflowing sea water, upwelling along the rims of the basins, and vertical diffusion through the halocline. Using information about the temporal changes of salinity and the advection of the bottom water into the deep basins, estimates of the long-term averaged vertical diffusivities have been made for the deep water of the Baltic Sea by several authors (Shaffer, 1979; Stigebrandt, 1987a; Matthäus, 1990; Axell, 1998). However, vertical diffusivities estimated by these methods represent the effects of both turbulent vertical mixing and upwelling.

For numerical models the vertical turbulent diffusivity $k_v$ is parameterised in terms of the Brunt-Väisälä frequency $N$ by

$$k_v = \frac{a}{N}.$$  

It was tuned by varying the constant $a$ until the observed and modelled variations of the stratification were in maximal agreement. Stigebrandt (1987a) obtained in his horizontally integrated model $a = 2.5 \times 10^{-2} \text{m}^2/\text{s}$. In this type of model the upwelling at the rims of the basins is not resolved.

A refined one-dimensional numerical ocean model of the southern Baltic Sea was used by Axell (2002) to investigate suitable parameterisations of unresolved turbulence in comparison with available observations. The turbulence model is a $k-\varepsilon$ model that includes extra source terms of turbulent kinetic energy production by unresolved breaking internal waves and Langmuir circulations.

The energy for deep water mixing in the Baltic Sea was provided by the wind. A range of values for the power of $N^{-n}$ in the proportionality relation to $k_v$ was tested in hundreds of 10-year simulations of the southern Baltic Sea. It was concluded that $n = 1.0 \pm 0.3$ and that a mean energy flux density to the internal wave field of about $(0.9 \pm 0.3) \times 10^{-3} \text{W/m}^2$ is needed to explain the observed salinity field. Finally, it was also shown that Langmuir circulations are important to be included when modelling the oceanic boundary layer. Using a fully 3-dim circulation model of the Baltic Sea, which resolves coastal upwelling, Meier (2001) obtained a somewhat smaller constant of $a = 1 \times 10^{-2} \text{m}^2/\text{s}^2$.

Direct estimates of diapycnal exchange coefficients have been made by Kullenberg (1977) from dispersion measurements of injected dye tracer in the thermo- and halocline of the Arkansas Basin and the Bornholm Basin in the Baltic Sea.

The turbulent diapycnal exchange coefficient in stratified water can be estimated according to Osborn (1980) assuming a balance between the production of turbulent kinetic energy, the buoyancy flux, and the dissipation of turbulent kinetic energy

$$k_v = \Gamma \frac{E}{N^2},$$

where $\Gamma = 0.2$.

Dissipation measurements in the Eastern Gotland Basin were performed by Lass et al. (2003) during winter stratification in April 1999 and during summer stratification in September 2000. Dissipation profiles were measured about every 10 min over a time interval of about 9 days. This provided a data set allowing for the estimation of quite reliable averaged dissipation profiles in spite of the huge intermittency of dissipation in stratified water (Fig. 10). The dissipation decreases from the surface to a depth of about 50 m. Maximum dissipation is observed in the halocline while it decreases below the halocline to an absolute minimum in the deep water until it increases again in a bottom boundary layer with a thickness of about 10 m.

There are significant differences between the dissipation profiles measured in winter and summer stratification below the surface mixed layer.

During summer stratification the thermocline is located at about 20 m depth and the dissipation decreases in the core of the intermediate winter water in the depth range between 40 m and 60 m to a minimum level, which is usually observed in the stratified layers well below the halocline. In winter the dissipation is much stronger at the bottom of the brackish winter water since it belongs to the surface mixed layer in winter. The dissipation in the halocline is larger by one order of magnitude during summer stratification compared to winter stratification. Since the Brunt-Väisälä frequency in the halocline does not change by more than 20% during the seasons, the total energy of the internal waves must be higher by about a factor of 10 according to Eq. (6). In summer the dissipation in the bottom water below the halocline is by 50% lower than the dissipation during the winter stratification except for the bottom boundary layer, where the dissipation increases again by a factor of 10 compared to the minimum dissipation in the centre of the bottom water body. The annual variations of the Brunt-Väisälä frequency below the halocline are quite low in the Eastern Gotland Basin. This suggests that the total energy of the internal waves below the halocline is significantly lower during the summer than during the winter stratification.

Diapycnal mixing in the turbulent regime below the surface layer was estimated by a relation according to Osborn (1980) which is based on the balance between the shear turbulent production, the work on buoyancy forces, and the dissipation rate assuming a constant Richardson flux number (Fig. 11).

The halocline turned out to be an isolating layer with respect to diapycnal mixing. The minimum diapycnal mixing coefficient of
$2 \times 10^{-6}$ m$^2$/s was observed at this depth. Below the halocline the diapycnal mixing increases gradually to $6 \times 10^{-6}$ m$^2$/s and exhibits a local maximum of about $8 \times 10^{-6}$ m$^2$/s. Diapycnal diffusivity at the depth of the halocline was estimated by means of the salt budget method for the Eastern Gotland Basin by Matthäus (1990) and by Axell (1998). Matthäus (1990) obtained a value of $5 \times 10^{-6}$ m$^2$/s at 75 m depth while Axell (1998) reported $k_v = 11 \times 10^{-6}$ m$^2$/s at 115 m depth during spring time. The values are both larger by a factor 1.7–1.8 than the estimates of Lass et al. (2003) at the corresponding depths. Comparing the “directly measured” dissipation-based diapycnal exchange coefficients with those estimated by the bulk method, one has to take into account that the bulk method supplies long term averages, which include the vertical transport by diapycnal mixing as well as by upwelling in the Baltic Sea. This is due to the short time scale of upwelling events and the fact that the upwelled water is partly mixed with the surface water due to

![Fig. 10. Averaged dissipation of turbulent kinetic energy measured in the Eastern Gotland Basin in April 1999 and September 2000 taken from Lass et al. (2003).](image)

![Fig. 11. Averaged diapycnal mixing coefficients according to Eqs. (7) and (8) and vertical exchange coefficients estimated by means of the bulk method for the Baltic Sea by Matthäus (1990) at 75 m depth and by Axell (1998) at 115 m depth as well as by means of dye tracer dispersion at about 300 m depth in the ocean thermocline by Ledwell et al. (1998).](image)
the Ekman off-shore transport. Therefore the bulk method should provide larger values. The diffusivity estimates based on dissipation measurements agree well with the average value given by Kulcken (1977) obtained from dispersion of dye tracers released in the thermocline and halocline of the Arkona and Bornholm Basin in the Baltic Sea. This suggests that the diapycnal exchange coefficients estimated for the halocline of the Eastern Gotland Basin hold for large areas of the Baltic Sea.

Another comparison of directly and indirectly estimated turbulent diffusivities is possible by the assumption that the mean vertical salt flow through the halocline must balance the export of salt from the Baltic Sea. In Section 3.1 the required vertical salt transport through the halocline is estimated to \( J_s = 30 \; \text{kg/(m}^2 \text{a}) = 9.5 \times 10^{-7} \; \text{kg/(m}^2 \text{s}) \). Assuming that this salt transport is maintained by a vertical turbulent exchange coefficient \( k_v \) with a characteristic vertical salinity gradient in the halocline of the Eastern Gotland Basin of \( \Delta S/ \Delta z = 5 \; (\text{g/kg})/10 \; \text{m} \) we obtain \( k_v = 2 \times 10^{-6} \; \text{m}^2/\text{s} \). This suggests that a large fraction of the vertical salt transport through the halocline may be maintained by turbulent diapycnal mixing resulting from breaking internal waves. However, some additional vertical transport by other processes such as entrainment resulting from vertical convection or upwelling at the rim of the basin is expected.

5.2. Boundary mixing due to internal waves

5.2.1. Boundary mixing in lakes and the ocean
A frequently observed feature of stratified water bodies is a well-mixed bottom boundary layer (hereafter abbreviated as BBL) existing right above the sediment surface. The height of this well-mixed layer varies between a few meters in lakes and reservoirs (Gloor et al., 2000; Hondzo and Haider, 2004; Lemckert et al., 2004) up to several tens of meters in the ocean (Caldwell, 1978; Trowbridge and Lentz, 1991; Moum et al., 2004). The turbulent kinetic energy required to generate and maintain such mixed layers is usually assumed to be produced by bottom friction of basin- or large-scale currents (Fricker and Nepf, 2000; Wüest et al., 2000), shoaling and critical reflection of high-frequency internal waves (Thorpe, 1997; Imberger, 1998), or the interaction of large-scale currents with rough topography (Rudnick et al., 2003). Fully non-linear and non-hydrostatic numerical models have been able to reproduce the mechanisms of energy transfer from the primary internal wave to higher modes (Vlasenko and Hutter, 2002b; Vlasenko and Alpers, 2005), as well as the processes of shoaling and breaking of internal waves on a scale, and the subsequent generation of turbulent BBLs (Vlasenko and Hutter, 2002a).

Direct observations of enhanced mixing in the presence of boundaries were reported in numerous studies in lakes and the ocean (MacIntyre et al., 1999; Ledwell et al., 2000; Garrett, 2003; Wüest and Lorke, 2003b). Turbulence in BBLs is an interesting scientific topic in itself, but it has received particular attention due to the fact that the presence of turbulent BBLs has important implications for the global flux paths of salt, heat, and matter in closed and stratified basins, the most obvious being the potential to greatly increase basin-scale vertical mixing (Imberger, 1998; Müller and Garrett, 2004). Indeed, enhanced basin-scale vertical transport due to turbulent BBLs has been confirmed by tracer experiments in lakes (Goudsmit et al., 1997), fjords (Stigebrandt, 1979), and ocean basins (Ledwell and Bratkovich, 1995; Ledwell and Hickey, 1995). Apart from being an important contribution to the effective basin-scale vertical transport, the presence of a well-mixed BBL is also known to give rise to secondary circulation patterns that may considerably affect the whole system, e.g., by effectively reducing the bottom friction (Weatherly and Martin, 1978; Trowbridge and Lentz, 1991; MacCready and Rhines, 1993).

A different mechanism capable of generating buoyancy-driven (convective) mixing in BBL on slopes was proposed by Lorke et al. (2002), Wüest and Lorke (2003a). This mechanism is associated with up-slope currents on slopes in stratified basins, produced by long inertial-internal waves, internal seiches, and internal tides. The decrease of the cross-slope current velocity towards the sediment surface (‘law of the wall’, Wüest and Lorke, 2003b) leads to a differential transport of water masses in the cross-shelf direction. This mechanism can result in a net transport of dense water on top of light water, and hence to shear-induced convective mixing. Evidence for the occurrence of convectively-driven mixing in BBLs in the ocean was presented by numerical simulations of Slinn and Levine (2003), as well as in recent observations on the continental shelf by Moum et al. (2004).

5.2.2. Boundary mixing studies in the Baltic Sea
Compared to the ocean and many fjords, the Baltic Sea is a particular example of a stratified basin with very weak tidal forcing. In the virtual absence of barotropic tidal pressure gradients, internal tides can therefore be excluded as relevant energy sources for diapycnal mixing in the BBL. This fact makes boundary mixing in the Baltic Sea energetically more similar to lakes (Lorke et al., 2002; Umlauf and Lemmin, 2005) and fjords with weak tidal forcing (Arneborg and Liljebladh, 2001a,b), where long internal waves and internal seiches adopt the role of the internal tides in the open ocean. The spectrum of internal waves available for energizing the BBL in the Baltic Sea include short internal waves, inertial-internal waves, and coastally trapped long internal waves.

Investigations of internal wave effects on BBL turbulence are challenging because a large number of parameters has to be controlled. First, turbulence and mixing have to be measured either directly, e.g., with the help of microstructure measurements, or by considering the integrated effects of mixing using a budget method for temperature, salinity, or a tracer. Second, the internal wave field has to be appropriately resolved, e.g., using acoustic velocity profilers or high-resolution moored CTD-chains. To our knowledge, no study has been published satisfying all these criteria. Nevertheless, a number of less complete investigations of BBLs has been reported, and will be briefly summarized in the following.

The field measurements conducted during the DIAMIX project (Stigebrandt et al., 2002) in the Eastern Gotland Basin come closest to the requirements mentioned above. A detailed study of internal wave motion and their relation to mixing in the water column from the first DIAMIX cruise in winter 1999 has been published by Lass et al. (2003) (Section 5.1). Even though during these measurements all parameters were available for a detailed study of turbulence and internal wave motions, the microstructure measurements stopped well above the BBL such that no conclusions can be drawn about its turbulent structure. During the second set of experiments, however, undertaken during the DIAMIX cruise in summer 2000, microstructure measurements resolving the BBL on the slope of the basin (Fig. 12) were obtained (Stigebrandt et al., 2002). The measurements showed clear evidence for the existence of enhanced turbulence in the BBL, but so far these data have not been analyzed in depth, and hence no definite conclusions can be drawn about the physical mechanisms generating turbulence in the BBL.

Another short paper directly focusing on the effect of breaking internal waves on the stratification at the slopes of the Eastern Gotland Basin has been published by Ozmidov (1994b). This author presented a number of CTD transects across the eastern slope of the basin, demonstrating quite clearly the presence of reduced density gradients in the lowest 10–20 m of the halocline adjacent to the sediment. A model expression for the vertical diffusivity derived by Ozmidov (1983) was shown to be consistent with the observed reduction of the density gradient. However, even though
Ozmidov (1994b) briefly mentioned the deployment of a microstructure profiler during the campaign, no turbulence measurements and no direct evidence for the action of internal waves has been reported, and the paper remains inconclusive in that respect.

Breaking internal waves are not the only source of turbulence in the BBL. Coastally trapped long internal waves and internal seiches are an important contribution to the spectrum of motions in the Baltic Sea, and their occurrence has been reported in numerous contexts (Walin, 1972a,b; Pizarro and Shaffer, 1998). Their velocity fields periodically move stratified water up and down the slopes, thereby causing friction and strain in the BBL. Both effects may cause turbulence in the BBL as observed in the connection with internal seiching in lakes (discussing above, Lorke et al., 2002).

Finally, Stips et al. (1998) reported a detailed study of stratification and mixing in the BBL of the Arkona Basin in the western Baltic Sea. They estimated the dissipation rate from repeated shear microstructure profiles along with high-precision CTD data. The near-bottom current structure was simultaneously measured with conventional current meters at fixed depths. A well-mixed, turbulent BBL was observed, a few meters thick with strong fluctuations in thickness due to the effect of horizontal advection. The paper was inconclusive regarding the physical mechanisms controlling turbulence in the BBL, and the effect of internal waves has not been a focus of the study.

These evident knowledge gaps about boundary mixing processes in the Baltic Sea have recently motivated a number of focused projects. Explicitly devoted to boundary mixing processes are the ongoing projects “Quantification of shear-induced convection and bottom-boundary mixing in natural waters” (ShIC) and the “International Leibniz Graduate School on Internal Waves in the Atmosphere and Ocean” (ILWAO). The former aims at the identification and quantification of boundary-layer mixing processes, and latter focuses on the impact of internal-wave mixing, and in particular bottom boundary-layer mixing, on the basin-scale diapycnal exchange. During the “Baltic Sea Tracer Release Experiment” (BaTRE), an inert tracer gas was released in September 2007 into the deep waters of the Eastern Gotland Basin. The vertical and lateral spreading of the tracer is currently being monitored, and the basin-scale mixing rates inferred from the spreading of the tracer will be compared to mixing estimates obtained from direct microstructure turbulence measurements and arrays of moored instrumentation. The outcome of these projects will likely increase our knowledge about mixing hotspots and boundary-layer mixing, and yield an improved understanding of the transport pathways in the Baltic Sea.

5.3. Vertical mixing by mesoscale eddies

5.3.1. Mechanisms of vertical mixing by mesoscale eddies

Vortices with a horizontal scale of the order of the baroclinic Rossby-radius are commonly referred to as mesoscale eddies. Mesoscale eddies are found in nearly every region of the world ocean. They carry energy and momentum and can contain significantly different water properties compared to their surrounding. Therefore they have the ability to transport and to influence mixing (Robinson, 1983).

Mesoscale eddies are also present in the Baltic Sea. The Baltic Sea Eddies are called Beddies here. They contribute to the vertical mixing, in particular to the diapycnal mixing in the permanent halocline, mainly by two mechanisms. One of them is the vertical displacement of water and isopycnals within the Beddies caused by their rotation and geostrophic adjustment. In fact, this is rather a displacement than an ongoing transport. It shows up as a lifting or lowering of the isopycnals inside the Beddies. The other one is connected to the decay of the Beddies. During the decay process their kinetic and potential energy becomes available for mixing processes no matter if they slowly fade away by dissipation or if they are destroyed more in a collapse caused by a collision with the basin rim, for example. Unfortunately, neither the lifetime of the Beddies due to pure dissipation nor their real lifetime is known. However, assuming a lifetime of the Beddies longer than 4 month resulting from pure dissipation in relation to a horizontal basin scale of several 100 km and a drift velocity of some 1 cm/s, such a collapse due to a collision seems to be more likely as the favored decay mechanism of the Beddies than their dissipation. The energy release in such a collapse would show up at least as the mixing of the water contained by the Beddy with the surrounding water masses. Moreover, some of the energy is assumed to be radiated in the form of internal waves. Therefore it may contribute to the vertical mixing by means of the internal wave field (Section 5.1).

In addition to these more or less direct mechanisms to impact the diapycnal mixing, there is some indication that the internal
waves interact with the Beddies (Talipova et al., 1998). This is consequently leading to the widely open question about their indirect impact on the vertical mixing by means of such interactions, e.g. their effect as scattering body for internal waves or the generation of critical layers for the internal wave field resulting in enhanced internal wave breaking.

5.3.2. Observations of mesoscale eddies in the Baltic Sea

Beddies were observed in most regions of the Baltic Sea during several field measurements (Aitsam and Elken, 1982; Aitsam et al., 1984; Elken et al., 1988; Sturm et al., 1988; Elken, 1996; Zhurbas and Paka, 1997, 1999; Stigebrandt et al., 2002; Lass and Mohrholz, 2003). All these investigations report on single outstanding features in hydrographic data fields which are considered as mesoscale eddies. They were found at various depths, some cover the whole water column, but most of them were observed in the region of the permanent halocline showing no surface signature. For their diameters values between 10 km and 20 km are given. Their thickness ranges between a few meters and the entire water depth. For the Beddies located inside the halocline vertical extensions of the order of magnitude of the halocline thickness itself are found, which is around 20 m depending on the region considered. The Beddies drift with velocities of a few 1 cm/s. They spin with maximum rotational speeds between 20 cm/s and 30 cm/s and are in geostrophic balance.

In another approach, using an objective pattern recognition algorithm to automatically detect Beddies as anomalies in three-dimensional hydrographic data fields, all Beddies contained in a certain ocean volume at one time are described by their integral properties (Reißmann, 2002; Reissmann, 2006). The 12 investigated data fields cover the Arkona Basin (AB), the Bornholm Basin (BB), the Słupsk Furrow (SF), and the Eastern Gotland Basin (EGB) during both summer and winter situations, i.e., with and without a thermocline (Fig. 13). Overall, the resulting integral properties are in good agreement with those of single Beddies. In particular, their spatial scales are reflected well by the corresponding mean values. Drift velocities and rotational speeds of the Beddies were not provided by this approach because of the limited capability of the used data fields. Nevertheless, some valuable information about other properties of the Beddy fields is gained for the estimation of their possible impact on the vertical mixing. One of the most important is the number of Beddies coexisting in each region. It varies from about 15 in the AB, BB, and SF up to 30 in the EGB.

Most of the Beddies are found in or above the regional main pycnocline. This is a consequence of the smaller volume available for the Beddies at the bottom, due to the basin shape in each region, and does not mean that there are no Beddies below the main pycnocline. The horizontal distribution of the Beddies in each of the four basins is consistent with a uniform distribution.
In each region the Beddies occupy a constant fraction of about 12% of the investigated basin volume (Fig. 14). This corresponds to mean volumes of the Beddies ranging from about 1.5 km\(^3\) in the AB, around 2 km\(^3\) in the BB and SF, and more than 3 km\(^3\) in the EGB. This increase of the mean volume is accompanied by an increase of the mean thickness of the Beddies from around 13 m in the AB to 23 m in the EGB, while their mean horizontal cross-sectional area varies only little within the range corresponding to radii between 4 km and 5 km. Therefore the increasing mean volume of the Beddies can be linked to the increase of their mean thickness. This in turn may be linked to the generally increasing water depth from the AB to the EGB. Under the assumption that a large number of Beddies is located within the halocline it also can be linked to the increase of the halocline thickness.

In all data fields the mean density of available potential energy of the Beddies is lower than the energy density needed to destroy the stratification in the corresponding halocline. The kinetic energy of the Beddies is not considered in this estimation. However, in most of the data fields there are at least a few Beddies found with available potential energy densities larger than that mixing energy density. The sum of their available potential energy is sufficient to destroy the stratification in a relevant fraction of the corresponding halocline of more than 20% in about half of the data fields. These data fields originate from the AB and the BB during winter situations and the SF irrespective of the stratification situation. In the EGB this fraction is negligible for all data fields. In this way some evidence is given about the regions and the season in which an enhanced vertical mixing by Beddies can be expected.

The mixing capability of the Beddies presumably is underestimated here by taking into account only their available potential energy and, consequently, neglecting their kinetic energy, even if not all of their energy is consumed by mixing processes. Finally, the total available potential energy of all Beddies in each data field is higher during winter than during summer situations. A correlation of the total available potential energy of the Beddies to the external forcing in terms of the wind stress curl could not be proved (Reißmann, 2002).

5.3.3. Estimates of the vertical salt transport by Beddies

The simplest approach to evaluate the possible impact of Beddies on the vertical mixing is to estimate the vertical transport of salt which is feasible by the Beddies. Considering only those Beddies that are located in the permanent halocline, this provides a characteristic measure of the diapycnal mixing which can be easily compared to the corresponding bulk estimates of the vertical transport.

For this purpose, typical salinities \(S_h\) of 13 g/kg, 12 g/kg, 11 g/kg, and 9 g/kg are taken within the halocline in the AB, BB, SF, and EGB, respectively (Reißmann, 2002, 2006). The respective thicknesses \(d_h\) of the halocline in the four regions is taken as 9 m, 10 m, 9 m, and 11 m using a second derivation criterion according to Reißmann (2006). The constant overall volume fraction of 12% occupied by Beddies (Reißmann, 2005) is used here for the halocline depth range in spite of a somewhat uneven vertical distribution of the Beddies. Finally, the lifetime \(\tau_p\) of the Beddies is assumed to be 4 month, because a destruction of the Beddies due to a collision with the basin rim is most likely after this time span at the latest as estimated above.

Assuming that the entire salt carried by the Beddies is transferred upward through the halocline, the diapycnal transport rates can be estimated as \(12\% \times d_h \times S_h \times p_h \times \tau_p / \tau_d\) with the density \(p_h\) within the halocline. The given values result in diapycnal transport rates of 42 kg/(m\(^2\)a), 43 kg/(m\(^2\)a), 36 kg/(m\(^2\)a), and 36 kg/(m\(^2\)a) in the AB, BB, SF, and EGB, respectively. So, they are slightly higher in the basins of the western Baltic Sea, but, overall, they are of the same order of magnitude as the bulk estimates in Section 3. This is at least indicating that the Beddies have the potential capability to contribute significantly to the vertical transport. This conclusion still holds if it is assumed that only half of the salt carried by each Beddy is transported through the halocline for geometrical reasons and/or only half of the Beddies contribute to the transport because half of them may originate from the surface layer, for instance. All these reductions of the diapycnal transport rates may be compensated, for example, by a conceivable shorter lifetime of the Beddies and/or by taking into account the energy release of the Beddies in their decay which nearly would double the mixed volume fraction of the halocline in most cases.

5.3.4. Variation of vertical mixing by Beddies in space and time

The vertical mixing due to the collapse of Beddies as a result of their collision with the basin rim, obviously, is focused to the region of the basin rim in each basin. In particular, for the vertical transport through the halocline this refers to the area around the 30 m, 55 m, 45 m, 70 m isobaths for the AB, the BB, the SF, and the EGB, respectively, according to Reißmann (2006). In contrast, any other mixing process induced by Beddies is effective uniformly in space on scales larger than the Beddies themselves in each basin. The potential energy of the Beddies which is available for the destruction of the stratification in the halocline gives some evidence that the Beddy induced vertical mixing through the halocline is strongest in SF and weakest in EGB (Reißmann, 2002, 2005).

According to the present knowledge the Beddies and, consequently, the mixing events induced by them occur uniformly in time. There is neither an indication for a seasonality nor a dependence on the meteorological forcing in terms of the wind stress curl over the Baltic Sea found for the occurrence of Beddies (Reißmann, 2002). However, it is not known whether their occurrence depends on inflow events to the Baltic Sea and its basins. Moreover, no information exists about the temporal variations of the mixing intensity induced by the Beddies. In particular, a seasonality of the mixing intensity due to the different properties of the halocline in summer and in winter is likely but not scientifically proven. With respect to the potential energy of the Beddies which is available for the destruction of the stratification in the halocline, there is some evidence that the Beddy induced vertical mixing through the halocline is stronger in winter than in summer (Reißmann, 2002, 2005).

5.4. Boundary mixing due to near-bottom currents induced by inflow events

The Eastern Gotland Basin (EGB) represents the largest volume of deep water in the Baltic Proper and the observed thermohaline mixing is typical for all deep basins of the Baltic Sea. The ultimate question is: Can deep boundary mixing be sufficiently effective to explain basin-averaged vertical diffusion and which role does the vertical shear of deep rim currents play beneath the perennial pycnocline?

For the open deep ocean Munk (1966) made an estimate of the diffusion coefficient \(k_v \sim 10^{-4} m^2/s\) to describe turbulent mixing across isopycnal surfaces. Twenty years later measurements of Moum and Osborn (1986), Gregg (1987) suggested that \(k_v\) is one magnitude smaller. Beside this open question, Phillips et al. (1986) argued that internal mixing over a sloping boundary should spread isopycnals up and down the slope and that resulting buoyancy forces drive a flow, which converges at the pycnocline causing an outflow toward the basin’s interior. After further ten years Toole et al. (1994) reported enhanced mixing near steeply-sloping boundaries. Estimated diffusion rates significantly exceeded values proposed earlier. Thus a wide range of \(k_v\) values can be found in the literature. Concerning shallow western areas of the Baltic Sea, Stips
et al. (1998) concluded from measurements of turbulent dissipation rates that no simple parameterization seems applicable, because an overall logarithmic near-bottom layer could not be verified. Model approaches relate $k_*$ to the squared stability frequency $N^2$ and the dissipation rate $e$ according to Eq. (8) with the efficiency factor $0.1 < \Gamma < 0.3$ (Lilly et al., 1974 for the atmosphere, Osborn, 1980 for the ocean). From measurements of $e$, which were carried out in the EGB during summer environmental conditions by Stigebrandt et al. (2002), it became evident that the perennial halocline/pycnocline provides an internal source layer for enhanced dissipations in between 60 and 100 m depth. There also was some observational evidence for increasing $e$ towards the basin’s steep slopes, roughly by one order of magnitude. The first case may be explained by breaking internal waves traveling within the wave guide of the pycnocline. In the sense of Munk and Anderson (1948) the second case may be attributed to regionally reduced Richardson gradient numbers $R_e = N^2/|\Delta v/\Delta z|^2$ via an enhanced vertical shear $\Delta v/\Delta z$ of rim currents $v$ following the deep isobaths.

A strong inflow of relatively warm but dense water into the EGB was reported for the winter 1997–1998 by Hagen and Feistel (2001). During this period of time two subsurface moorings recorded currents and temperature at 170 m depth, while thermometers were deployed at the 140 and 155 m horizons (Fig. 15). The water depth was 220 m at both positions. Results suggested a persistent cyclonic circulation of about 0.03 m/s, which occupied the entire deep EGB. Six subsequent inflow pulses with an overall lag of 22 days accelerated this background circulation by the factor of about two. Each of them produced lateral and vertical meanders of the deep rim current. Associated fluctuations modulated the deep thermal field on the daily scale (Fig. 16). Following Gregg (1989) such fluctuations influence the intensity of mixing on much shorter temporal scales and must be mirrored in corresponding fluctuations of the 'eddy kinetic energy' per unit mass $EKE = (\sigma_x^2 + \sigma_y^2)/2$. Here estimated $EKE$ values result from daily variances $(\sigma_x^2 + \sigma_y^2)/2$ of the zonal ($u > 0$, eastward) and meridional current component ($v > 0$, northward). On the other hand, the daily current variability is described by changes of the 'mean kinetic energy' per unit mass $MKE = (u^2 + v^2)/2$.

From the statistical point of view the MKE is based on the first (daily averages) and the EKE on the second moments (daily variances). Both estimates should be uncorrelated by definition. This means, however, that any statistical relationship between the MKE and the EKE points to a persistent energetic flux, either from the low frequent towards the high frequent spectral range of current fluctuations or vice versa. One may expect a somewhat stronger interaction between both quantities at the NE position because it directly locates in the entrance area of deep water intrusions (Fig. 15). However, one may also expect that the intensity of associated mixing processes significantly decreases along the pathway of the deep cyclonic background circulation. This is confirmed through daily temperature variances recorded at both the NE and the SW position (Fig. 16). Associated fluctuations in the north-east significantly exceed those in the south-west. This is confirmed by the comparison of logarithmically scattered values of the MKE and the EKE (Fig. 17). Obviously, the interaction between the MKE and the EKE weakens drastically on the way path of deep currents from the NE towards the SW position. This indirectly suggests that intense mixing of thermohaline properties prevails over the eastern topographic flank of the basin and, consequently, deep isopycnals climb to shallower pressure levels (Fig. 18).

Pressure levels of selected deep isopycnals are drawn for the pre-inflow (M−7) and the post-inflow situation (M−8) along the zonal section at 57.29°N. The time assignment of both field campaigns is given in Fig. 16. Thus it becomes clear that the intrusion of dense deep water caused a significant upward displacement of all density surfaces with doming above the eastern rim of the basin’s centre. Associated overall vertical velocities <$\omega>$ reached the magnitude of $10^{-6}$ m/s. They decreased from near-bottom towards intermediate layers by the factor of about four, mainly due to corresponding changes in the deep water volume (Fig. 19).

Six years of temperature and current records (31 August 1999–30 October) collected 20 m above the sea bed at the NE position, which are described in more detail by Hagen and Feistel (2004), Feistel et al. (2006a), Hagen and Feistel (2007), confirm that such upward displacements of isopycnal surfaces accompany all inflow events. For instance, the short warm water inflow of the early 2003 was immediately replaced by a cold one in April–May. About three months later, the several intrusions of warm water completely changed the hydrographic regime of the deep EGB (Fig. 20). The embedded cold inflow event peaked with a daily temperature minimum of 4.41 °C at 219 m on 15 May, but with that of 5.1 °C at 174 m depth on 20 May. Such delay in the peak values suggests that the temperature of the enclosed 45 m layer was mixed intensively during 5 days. These values point to the overall upward velocity of about <$\omega>$ = 9 m/d or $1.04 \times 10^{-4}$ m/s and demonstrate that the <$\omega>$ varies by two orders of magnitude between the 5-day and the 130-day scale (Fig. 19).

Concerning trigger mechanisms of involved diapycnal mixing, the long-term records of the NE mooring underline the importance of so-called integral effects which result from the time history of short-term fluctuations in the vertical current shear. A substantial contribution originates from local inertial oscillations with a period of about 14.4 h. The power spectra of both current components clearly show that the energetic level of this period decreases with increasing water depth. However, it also became clear that the intensity of inertial oscillations drastically increases during and
after deep inflow events. This follows from series of the tempera-
ture and the current components, which have been smoothed by
14 hourly running means. Associated standard deviations mirror
the energetic level of inertial oscillations (Fig. 21). Peak values in
the $EKE = (\sigma_u^2 + \sigma_m^2)/2$ clearly coincide with those of the tempera-
ture STD(T) after the three identified deep water intrusions
(Fig. 20). Released fluxes of kinetic energy between the
$MKE$ and the $EKE$ influenced, for instance, the 30 m layer between 174 and
204 m depth during the whole year 2003 (Fig. 22). This follows
from the overall positive regression between the $MKE$ and the
$EKE$. However, a comparable regression between the vertical cur-
rent shear $S$, which was normalized by the local Coriolis frequency
$f = 1.25 \times 10^{-4} \text{s}^{-1}$, and the $EKE$ could be only confirmed for actual
inflow situations.

This emphasizes the particular role of dense water intrusions and associated intensifications of the vertical shear in deep rim currents for regionally enhanced mixing conditions. The time his-
tory $(t)$ of daily anomalies in the vertical shear of along-slope cur-
rents $v$, i.e. $\sum (dv/dz)^2(t)$, points to the major source for daily changes in the vertical temperature gradient $dT/dz(t)$
(Fig. 23). After the deep inflow of warm water on day 73 (14 March
2003), the tendency of increasing vertical shear in the deep rim

Fig. 16. Daily temperature variances in $K^2$ (thin line) recorded with a sampling interval of 0.5 h at 140 and 155 m and with the sampling interval of 1 h at 170 m depth at both
the NE position (57°25.38’N, 20°20.83’E) and the SW position (57°04.53’N, 19°45.12’E) according to Hagen and Feistel (2001). The bold line represents five daily running
means for smoothing and vertical lines mark the inflow period lasting from 28 November 1997 until 6 May 1998 at the NE position and from 28 December 1997 until 26 May
1998 at the SW position. Downward arrows indicate the two hydrographic field campaigns denoted M-7 and M-8 along the station grid shown in Fig. 15.

Fig. 17. Logarithmically scattered daily ‘mean kinetic energy’ $MKE = (u^2 + v^2)/2$ versus corresponding ‘eddy kinetic energy’ $EKE = (\sigma_u^2 + \sigma_m^2)/2$ in 170 m depth of the
moored current meter strings NE and SW (Fig. 15). The recording length is N days and the pre-inflow is labeled by A (circles), the inflow by B (dots), and the post-inflow
situation by C (triangles and squares). Note the strict relationship between the $MKE$ and the $EKE$ at the NE position.
current was accompanied by increasing daily standard deviations of the vertical temperature gradient. However, the temperature signal was lagging the current shear tendency by four days. Thus the time scale of 4–5 days seems to be characteristic for the homogenization of thermal properties in the wake of deep inflow events.

Finally, turning back to the question asked at the beginning of this subsection, we may conclude that there is some observational evidence that strong deep water intrusions enhance the vertical shear of deep rim currents accelerating diapycnal mixing on the scale of few days.

5.5. Boundary mixing due to coastal upwelling

5.5.1. Process of vertical mixing by upwelling

For the Baltic Sea it is hypothesized that upwelling induced vertical transports of salt and nutrients are mainly contributed by coastal upwelling. The Baltic Sea is a nearly enclosed sea with a highly structured coast line. Thus at any wind direction there exists a coast which favors upwelling. The coastal upwelling is confined to an off-shore distance of the internal Rossby-radius which is of about 4–8 km in the Baltic Sea (Fennel et al., 1990). The uplift of the pycnoclines itself will not result in a vertical mixing. A differential advection is required that establishes an instable stratification and convective mixing. Two different mechanisms are known, which can produce an instable stratification due to coastal upwelling.

Fig. 19. Vertical profile of the overall vertical velocity \(<w>\) taken from Hagen and Feistel (2001). The vertical velocity \(<w>\) was derived from spatially averaged upward displacements \(<dh>\) of six selected potential density surfaces (PD) during 130 days between pre- and post-inflow conditions labeled by M-7 and M-8 in Fig. 16. The error bars point to the 95% confidence level of the t-distribution on the base of given station numbers.
Convection will not significantly contribute to diapycnal mixing through the halocline. However, it is of great importance for the transport of nutrients from intermediate winter water into the euphotic layer (Section 6). During winter conditions without a shallow seasonal thermocline the compensation flow can cover the winter water and the upper halocline, depending on the depth of the mixed layer. Then the resulting shear-induced mixing in the bottom layer near the coast contributes to the vertical transport through the halocline. The upwelling induced mixing in the surface layer enhances the deep winter convection due to cooling.

With respect to the global warming, in the future the upwelling process may become more important for the mixing of the upper layer. The expected change in atmospheric forcing will enhance the upwelling intensity during winter. If during warm winters the surface temperature of the Baltic Sea remains above the temperature of maximum density, the convective mixing due to seasonal cooling will be reduced. In that case upwelling and enhanced wind mixing can maintain the vertical transport of nutrients into the surface layer.

5.5.2. Observations and model results from the Baltic Sea

The first observations of local coastal upwelling in the Baltic Sea were published by Walin (1972a), who described an upwelling event in the Hanö Bay. Gidhagen (1984) gave an overview on the upwelling in the Baltic Sea based on remote sensing of SST. He found coastal upwelling cells along the entire Swedish coast localised at several spots. In summer the temperature in active upwelling cells drops typically by 4 to 5 °C, although decreases of up to 10 °C were also observed. The upwelling commonly persists for a week, but events with a month duration were observed, too. Gidhagen (1984) also found a strong interannual variation in the rate of upwelling events. He tried to compile statistics of the upwelling activity from remotely sensed SST, which however had to remain incomplete due to the fact that winter upwelling events have small SST signature only. Horstmann (1983) detected upwelling at the southern coasts of the Baltic Sea caused by easterly winds.

A systematization of the upwelling zones in the entire Baltic Sea was made by Bychkova and Victorov (1987), which was also based on satellite data of SST. Their main results are summarized by Victorov (1996). 22 coastal upwelling zones were identified and de-

![Fig. 20. Daily averaged temperature in °C recorded at the horizons of 174 (blue) and 219 m (red) with a sampling interval of 1 h at the position NE (224 m water depth, Fig. 15) modified from Feistel et al. (2006a). The recording gap during the late 1990s was filled by CTD snapshot data of the Baltic Monitoring Programme (HELCOM).](image)

![Fig. 21. Fourteen hourly standard deviations in temperature records, STD(T), and the corresponding ‘eddy kinetic energy’ per unit mass EKE = (σ_u^2 + σ_m^2)/2 at three horizons of the NE position (224 m water depth) during 2003. The intrusions of warm (W = strong, w = weak) and cold deep water (C), which are identified in Fig. 20, are marked by arrows at their starting day of the year (DOY) in the upper panel.](image)
scribed by means of their size, upwelling favorable wind directions, and observed upwelling frequency and intensity.

During the last two decades a number of publications described several local upwelling cells. A prominent one is the Hiddensee upwelling cell (Fennel and Sturm, 1992; Lass et al., 1994, 1996). Fig. 24 illustrates the onset and decay of an upwelling event west of Hiddensee in July 1995. The event lasted for a week. The observed temperature differences between the core of the upwelling filament and the surrounding surface water exceeded 10 °C.

However, besides some theoretical works (Fennel, 1992) about the upwelling dynamics most of the publications remained on a descriptive level and are focussed on summer events since they can be easily detected by SST observations. The role of upwelling for the vertical salt transport has hardly been discussed. Mainly upwelling induced vertical nutrient transport from the intermediate winter water into the surface layer were discussed, in terms of cyanobacterial blooms (Janssen et al., 2004) and also its effect on the zooplankton (Kostrichkina and Yurkovskis, 1986). Until today robust estimates of the contribution of coastal upwelling to the overall vertical transport do not exist (Lehmann and Myrberg, 2008).

Since direct measurements of upwelling events by remote sensing and/or in-situ observations are difficult and cannot overcome the undersampling problem, numerical models are increasingly used to analyse upwelling events in the Baltic Sea (Jankowski, 2002; Lehmann et al., 2002; Myrberg and Andrejev, 2003). Kowalewski and Ostrowski (2005) provide one of the most recent works about upwelling using a three-dimensional hydrodynamic model. On a basis of a seven year numerical simulation they could identify 12 upwelling zones in the southern Baltic Sea, where the vertical velocities for at least two adjacent wind sectors of 45° exceed 2 × 10⁻⁴ m/s. For each of these zones a statistic of up- and downwelling was calculated. The percentage of upwelling frequency with vertical velocities above 2 × 10⁻⁴ m/s range from 4.6 to 27.1% in the specified regions.

The total percentage of periods with vertical upward velocity was in nearly all regions around 50% (41.6–54.5%), except for the east coast of Bornholm where upwelling is observed 72.8% of the time. This is the region with the most intense upwelling in the model, showing the highest annual mean of vertical velocity of about 0.55 × 10⁻⁴ m/s. A second zone of intense upwelling is the Hanö Bay with 0.24 × 10⁻⁴ m/s annual mean vertical velocity. In contrast, Lehmann et al. (2002) found extreme values of the averaged vertical velocity around 5 × 10⁻⁶ m/s, resulting in a pycnocline uplift of 0.5 m/d, which is one order of magnitude less than reported by Kowalewski and Ostrowski (2005). Strongly enhanced vertical velocities were found for NAO+ phases as a result of the prevailing atmospheric conditions. Up to date estimates of the total vertical transport due to upwelling have not been made from numerical model simulations.

5.5.3. Quantification

Although the dynamics of upwelling are well investigated, a quantitative estimate of the contribution of coastal upwelling to the basin-scale vertical mixing in the Baltic Sea is still missing and cannot be derived from in-situ measurements. This was also concluded by Lehmann and Myrberg (2008) in their actual review on upwelling in the Baltic Sea. Only hydrodynamic models will be able to supply the necessary data. However, to resolve the different-
tial advection and convective mixing in the boundary layers high vertical and horizontal resolutions would be required which have not yet been reached in basin-scale modelling of the Baltic Sea.

To estimate an upper limit of the upwelling driven vertical transport, the Baltic Sea is extremely simplified assuming a circular basin with a total surface area of 400,000 km² (radius ~ 350 km). In this case upwelling is independent from the wind direction. Wind data, taken at the Darss Sill (1995 to 2004), were used to calculate a climatological wind field. The derived monthly means of windstress were applied to estimate

![Fig. 24. Example for upwelling in the Baltic Sea. The series of SST images, compiled according to Lass et al. (1996), shows the development and decay of an upwelling filament, stretching from the west coast of island Hiddensee in north-westerly direction towards island Møn.](image-url)
the Ekman-transport \( E \) (Fig. 25) according to Eq. (9) (Csanady, 1982).

\[
E = \tau f \quad \text{with} \quad \tau = \frac{\rho_{\text{air}}}{\rho_{\text{water}}} W_{10} \cdot |W_{10}| \cdot c_d
\]

The drag coefficient \( c_d \) was assumed to be a constant of \( 1.6 \times 10^{-3} \). \( f \) is the inertial frequency at \( 57^\circ \text{N} \), and \( W_{10} \) is the windspeed 10 m above the sea surface. In case of an ocean with an infinite coast the Ekman-transport is equal to the vertical transport in the active upwelling zone at the coast. Therefore the derived Ekman-transport can be assumed as an upper limit of the vertical transport due to upwelling.

The corresponding depth of Ekman layer and horizontal velocities were approximately 20 m and 4 cm/s for typical summer conditions and 30 m and 5.5 cm/s for winter respectively. Using a mean baroclinic Rossby-radius of 6 km (Fennel et al., 1990) as measure for the extension of the upwelling zone at the coast, the vertical velocity ranges for this idealized experiment between \( 1.3 \times 10^{-4} \text{ m/s} \) in summer and \( 2.6 \times 10^{-4} \text{ m/s} \) in winter. The annual upwelled water volume amounts to 27,100 km\(^3\), which compares approximately to the total volume of the Baltic Sea. Seasonally it splits into 16,600 km\(^3\) in winter (Oct–Mar) and 10,500 km\(^3\) in summer (Apr–Sep). Related to the volume of the Baltic Sea surface layer (15,000 km\(^3\)) upwelling has the potential to turn over the surface layer twice a year.

However, in a real ocean with structured coastlines and spatial and temporal fluctuating wind fields, the situation is different from the simplified case. Eq. (9) describes the conditions after the adjustment of the current field to a constant wind forcing, which typically takes one inertial period (14.4 h in the Baltic Sea). Compared to the mean duration of wind events of the order of 2–4 days, the adjustment of current field will be a continuous ongoing process. Additionally, each inhomogeneity in the wind field and the coastline generates Kelvin waves, traveling along the coast through the upwelling areas. After passing of the Kelvin wave the dynamic balance at the coast changed and the upwelling will be stopped (Fennel and Lass, 1989). Also the upwelling can be exported by Kelvin waves into regions without upwelling favorite wind forcing. Further, a significant number of forcing events are too short to generate a vertical mixing due to differential advection, since the tilted isopycnals are relaxing before an irreversible mixing has occurred. With an Ekman layer depth of 20 m and a vertical velocity of \( 1.3 \times 10^{-4} \text{ m/s} \) the isopycnals need 1.5 days to reach the surface. It has to be concluded, that it is more or less impossible to give an robust estimate of the total vertical transport in the Baltic Sea caused by upwelling. Comparing the maximum annual mean vertical velocity from the model of Kowalewski and Ostrowski (2005) with the estimates from the circular basin, the true value of the total vertical transport can be an order of magnitude less than the upper limit given above.

6. Ecosystem perspective

The availability of nutrients in the euphotic zone determines to a large extent the intensity of primary production. Beside silicate, which is essential for diatom growth, phosphate and inorganic nitrogen compounds (ammonium, nitrite and nitrate) have to be considered. In the central Baltic Sea the spring phytoplankton bloom is nitrogen limited due to the low \( N/P \) ratio of inorganic nutrients in the surface water in winter (Matthäus et al., 2001b). After the spring bloom phosphate remains at concentrations of about 0.10 \( \mu \text{mol/l} \) in May. These conditions are suitable for the development of diazotrophic cyanobacteria later the year. Intensive blooms are frequently observed in the Baltic Sea in summer. Processes and factors that control cyanobacteria growth and their mass development are manifold and by far not completely understood. Because of their ability to overcome the shortage of nitrogen by fixing atmospheric nitrogen, phosphorus plays a key role in the regulation process (Nausch et al., 2004).

The balance of phosphate in the mixed surface layer is determined through uptake processes by autotrophic organisms, mineralization of organic matter, and vertical transport in both directions. The downward vertical transport takes place mainly in particulate form as sedimentation, whereas the upward transport is mainly passive and follows the hydrodynamic salt transport. Also upwelling processes can play a prominent role. In the last decades, especially during the 1960s and 1970s, the concentrations were strongly influenced by horizontal transport and advection due to the massive input of nutrients from the catchment area as a result of eutrophication (HELCOM, 2003). Since the end of the 1970s winter concentrations fluctuate on a high level. This suggests that the downward and upward transport are of the same order of magnitude.

In this section the present knowledge is summarized about

(a) the chemical reactions and variations of phosphate in the deep water
(b) the upward transport of phosphate which is basically physically driven.

The nutrient conditions in the deep basins of the Baltic Sea react strongly on the alternation between inflow and stagnation periods. In the presence of oxygen, phosphate is partly bound in the sediment and onto sedimenting particles in the form of an iron-III-hydroxophosphate complex. When the system turns to anoxic conditions, this complex is reduced by hydrogen sulphide and phosphate and iron(II) ions are liberated. This interplay is perfectly mirrored in Fig. 26. At the end of a long stagnation at the beginning of the 1990s phosphate concentrations of around 7 \( \mu \text{mol/l} \) were measured. The MBI in January 1993 supported by two smaller inflow events in December 1992 and April 1994 increased the

![Fig. 25. Monthly climatological wind speed at the Darss Sill (1995–2004) and its first standard deviation (left panel) and the Ekman-transport (right panel) for a circular basin using the climatological wind as forcing.](image-url)
oxygen concentrations considerably (Fig. 27) and dropped the phosphate down to values below 2 \( \mu \text{mol/l} \) (Fig. 26). With the restoration of anoxic conditions in the following years phosphate concentrations increased again. This increase was interrupted only shortly due to the effects of an exceptionally warm summer inflow in 1997 (Matthäus et al., 1998) and an inflow in autumn 2001 (Feistel et al., 2003a). A drastic decrease of the phosphate concentrations in the deep water took place after the MBI of January 2003 had reached the Gotland Basin in May 2003 (Nausch et al., 2003).

MBIs are supposed to mainly replace the deep water masses in central basins of the Baltic Sea and thus influence the nutrient concentrations there. The question arises to which extent these changes can affect the surface layer. For comparison and nutrient trend analysis usually the surface layer in winter is used, because the biological activity is low and nutrient concentrations are high in winter (Nehring and Matthäus, 1991). Furthermore, in such analysis the assumption is made that a steady state between microbial mineralization, low biological productivity, and high vertical exchange and mixing has developed at this time. This steady state lasts 3–4 months and is most discernible in the Eastern and Western Gotland Basin, where values characteristic for the winter situation are sometimes measured as late as in early April.

Fig. 28 shows the averaged annual winter phosphate concentrations in the surface layer (0–10 m) pooling 6 stations in the Eastern Gotland Basin. The steep increase in the 1960s and 1970s can be seen resulting from the intensive eutrophication during that period. Phosphate concentrations of around 0.20 \( \mu \text{mol/l} \), as found in the late 1950s and early 1960s, are the natural background concentration for the open Baltic Sea area. After the remarkable increase concentrations remain at a high level with strong fluctuations as a result of mainly internal processes. Among these the effects of MBIs have to be discussed. In Fig. 28 these events are marked. It is evident that after the inflows 1975/1976, 1983, and 1993 lower phosphate concentrations were measured whereas a comparable decrease after the MBI of 2003 was not observed.

To understand this different behaviour, the history of the inflow events has to be considered. During an inflow event salty, oxygen-rich water masses penetrate into the near-bottom layer, partly precipitate the dissolved phosphate, and lift up older, oxygen-poor water masses rich in phosphate. In case that the water layers below the halocline are relatively well supplied with oxygen, phosphate is precipitated again as iron-III-hydroxophosphate complex and phosphate concentrations remain low. This was the case for the earlier inflow events. Consequently, relatively small amounts of phosphate were transported into the surface during deep vertical mixing in winter.

Nausch et al. (2003) have tried to budget the amount of phosphate stored in a box in the Eastern Gotland Basin. The box area below 70 m is 12,300 km\(^2\) with a volume of 343 km\(^3\). They described the phosphate content for three different depth regions: (a) between the halocline and the bottom (b) between the redoxcline and the bottom and (c) between the halocline and the redoxcline. Here the most interesting is the latter one (Fig. 29). At the end of
the stagnation period in 1992 the water body between 80 m and 125 m water depth was relatively well supplied with oxygen (Nehring et al., 1993). Therefore phosphate concentrations were comparatively low. Water renewal initially caused oxygen-poor water layers with high phosphate concentrations to be raised. Thus no further decrease between 1992 and 1995 was observed. During the stagnation period from 1995 onwards the water layers below the halocline remained extremely poor in oxygen with the result that enormous amounts of phosphate were accumulated in this layer leading to an increase of winter phosphate concentrations at the surface later on (Fig. 28).

Consequently, the lifting up of the "old" water masses after the MBI in January 2003 into the oxygen-poor water layer below the halocline resulted in an only moderate decrease of stored amount of phosphate. Thus 53 200 t P were stored in 2005 compared to 32 500 t P in 1992. Under these circumstances vertical mixing is able to transport much higher amounts of phosphate into the surface layer maintaining winter concentrations at a high level (Fig. 28). This behaviour can be described in more detail looking at the period January 2003 to December 2005 (Fig. 30). The oxygen-rich inflow near the bottom (green, beginning in April/May 2003) causes an elevation of the stagnant water (yellow) below the pycnocline from 100 m to 70 m depth, i.e. by almost 30 m. By comparing the development in the oxygen distribution with the measured phosphate concentrations substantial similarities are evident (Fig. 31). During the whole observation period high phosphate concentrations, partly higher than 4 μmol/l, were found directly below the halocline allowing for considerable ver-

![Fig. 29. Phosphate pool between the halocline (70 m) and the redoxcline (137 m) in the Eastern Gotland Basin - supplemented and updated according to Nausch et al. (2003).](image)

![Fig. 30. Oxygen/hydrogen sulphide concentration in ml/l at the Gotland Deep between 2003 and 2005 taken from Feistel et al. (2006b). Hydrogen sulphide concentration is shown as negative oxygen equivalent.](image)
tical nutrient transport, if the appropriate physical forcing was given.

Finally looking at the mean annual phosphate concentrations in 80 m water depth at the central station in the Eastern Gotland Basin between 1992 and 2005 (Table 2), the described differences become much more visible. Despite a long lasting stagnation period, which ended in 1993, the horizon below the permanent halocline contains only low phosphate concentrations due to the relative good ventilation of this water layer by intermediate inflow processes. A quite different situation was observed at the end of the following stagnation period in 2002. Nearly 3 $\text{µmol/l}$ phosphate were measured as a result of a quite poor supply with oxygen. And the situation did not significantly improve after the MBI of 2003 (Fig. 31).

In Section 3.1 a transport rate of around 30 kg/(m² a) for the vertical salt transport across the pycnocline into the surface layer of about 60 m thickness was estimated. This rate results from the erosion of a layer with a salinity of around $S = 10 \text{ g/kg}$. In combination with the measured phosphate concentrations $P_{80}$ as given in Table 2 in the 80 m depth level, a phosphate increase in the mixed layer of about $\frac{d}{dt} P_{\text{mix}} = 0.040 \text{ µmol/l/a}$ (1992) and $\frac{d}{dt} P_{\text{mix}} = 0.15 \text{ µmol/l/a}$ (1999) would hypothetically result from the balance formula

$$\frac{d}{dt} P_{\text{mix}} = \frac{30 \text{ kg}}{\text{m² a}} \times \frac{1}{60 \text{ m}} \times \frac{P_{80}}{10 \text{ g/kg}} \times \frac{1}{1000 \text{ kg/m³}} = \frac{P_{80}}{20a}$$

assuming a fixed $S:P$ ratio during this process.

This result does not necessarily mean that the transport is evenly distributed over the year. Most of the transport is realized during the deep vertical mixing in winter. Looking at their interplay, disagreements are evident between the estimated nutrient transport rates and the measured winter concentrations in Fig. 28. However, one has to keep in mind, that the water layer between the thermocline and the halocline is not completely exhausted from phosphate during the whole vegetation period. Also in the summer phosphate concentrations huge interannual differences can be observed depending on the preceding winter concentrations. For example, in summer 1997, which was a year with extremely low winter concentrations (Fig. 28), a phosphate content of around 0.20 $\text{µmol/l}$ was measured at 30 m depth. In 2005 at the same water depth around 0.55 $\text{µmol/l}$ phosphate could be detected.

In conclusion, looking at an ecosystem perspective, the vertical nutrient transport through the permanent halocline as well as the possible phosphate transport through the temporary thermocline is quantitatively not sufficiently well understood yet.

### Table 2

Mean annual phosphate concentrations at 80 m water depth at the central station in the Eastern Gotland Basin.

<table>
<thead>
<tr>
<th>Year</th>
<th>Phosphate (µmol/l)</th>
<th>Year</th>
<th>Phosphate (µmol/l)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1992</td>
<td>0.79 ± 0.14</td>
<td>1999</td>
<td>2.99 ± 0.07</td>
</tr>
<tr>
<td>1993</td>
<td>1.04 ± 0.26</td>
<td>2000</td>
<td>2.55 ± 0.50</td>
</tr>
<tr>
<td>1994</td>
<td>1.33 ± 0.43</td>
<td>2001</td>
<td>2.82 ± 0.31</td>
</tr>
<tr>
<td>1995</td>
<td>1.55 ± 0.23</td>
<td>2002</td>
<td>2.96 ± 0.22</td>
</tr>
<tr>
<td>1996</td>
<td>1.67 ± 0.30</td>
<td>2003</td>
<td>2.33 ± 0.36</td>
</tr>
<tr>
<td>1997</td>
<td>1.99 ± 0.39</td>
<td>2004</td>
<td>2.87 ± 0.29</td>
</tr>
<tr>
<td>1998</td>
<td>2.72 ± 0.23</td>
<td>2005</td>
<td>2.64 ± 0.16</td>
</tr>
</tbody>
</table>

7. Discussion

The Baltic Sea has to be roughly divided into at least two parts with respect to the dominant vertical mixing processes – the entrance area and the deeper basins east of Bornholm. In the shallow Belt Sea, gale-force winds are capable of mixing the entire water column down to the sea floor. In the deeper western parts the direct interaction between the inflowing saline water from the North Sea and the outflowing brackish water governs the vertical transport. The associated entrainment processes have strong regional hot spots as discussed in Section 4. However, the impact of the resulting vertical transport on the
water and salt budget of the Baltic Sea is hard to quantify, because the upward mixing of saline bottom water into the outflowing surface water and the downward mixing of brackish surface water joining the inflowing bottom water allow for complicated recirculation. Moreover, in certain regions this two-layer system can turn into a one-layer system of either brackish surface water or saline bottom water (Section 2). This even complicates the challenging task to determine the impact of the vertical mixing in this part of the Baltic Sea to its water and salt budget. Nevertheless, the entrainment and recirculation processes determine the properties and amount of the inflowing dense water entering the other part of the Baltic Sea, where temporal changes and the associated transports are dominated by horizontal advection of saline water in the bottom layers below the permanent halocline and turbulent vertical transport through the halocline in the surface layers above the halocline (Section 3.1). This part basically consists of the central Baltic Sea including primarily the Eastern Gotland Basin. The dominating transport processes in this part of the Baltic Sea allow for robust bulk estimations of the vertical salt transport which are the focus of Section 3. The estimated vertical salt transports can be assumed to be almost uniform in space for those regions in the central Baltic Sea which are covered by the permanent halocline. Depending on the method, the estimated vertical net transport of salt varies around values slightly above 30 kg/m² a.

7.1. General picture of vertical mixing in the central Baltic Sea

The vertical transport through the halocline into the entire surface mixed layer in the central Baltic Sea can be described consistently. The general dynamics of the surface mixed layer are presented in Section 2.2. The vertical transport into the entire surface mixed layer is basically maintained by the weakening of the halocline in summer and its erosion in winter when the surface mixed layer reaches down to the halocline because of the absence of the thermocline.

The weakening of the halocline in summer is accomplished by turbulent mixing which can be largely associated with breaking internal waves. In Section 5.1 the corresponding observed diffusivity in the range of the halocline is estimated to be sufficient to maintain a turbulent vertical salt transport which is of the order of magnitude required from the bulk estimates of the net upward transport of salt given in Section 3. However, this turbulent vertical transport is of short range of the order of magnitude of about 1 m at maximum (Section 2.2). The turbulent salt transport \( J_s \) is calculated according to

\[
J_s = k_v \frac{dS}{dz} \tag{10}
\]

with the turbulent diffusivity \( k_v \), the salinity \( S \), and the vertical coordinate \( z \). Consequently, the turbulent transport vanishes right above the halocline where the vertical salinity gradient disappears in the mixed layer above the halocline, even though somewhat larger diffusivities are observed outside the halocline (Fig. 11). This can be explicitly seen from the parametrisations of \( k_v \) given in Eqs. (7) and (8). Assuming that the variations of the potential density \( \rho_{pot} \) in the definition of the Brunt-Väisälä frequency \( N \), namely

\[
N^2 = -\frac{g}{\rho_{pot}} \frac{d\rho_{pot}}{dz} \tag{11}
\]

with the acceleration \( g \) due to gravity and the vertical upward coordinate \( z \), are mainly due to variations of the salinity \( S \), i.e.

\[
\frac{d\rho_{pot}}{dz} \approx \frac{\partial \rho_{pot}}{\partial S} \frac{dS}{dz} \tag{12}
\]

Eq. (10) results in turbulent vertical salt transports \( J_s \propto \sqrt{N} \) and \( J_s \propto \varepsilon \) for Eqs. (7) and (8), respectively. Note that the rate \( \varepsilon \) of dissipation of turbulent kinetic energy itself is proportional to \( N \) according to Eq. (6) or to powers of \( N \) up to 2 depending on its parametrisation (Section 5.1), and \( N \propto \sqrt{\varepsilon} \) with the approximation made in Eq. (12). Hence, the vertical turbulent salt transport \( J_s \) vanishes as expected for all parametrisations discussed here as the vertical salinity gradient disappears, provided that the approximation made in Eq. (12) is valid. Therefore the halocline is slowly extended upward and weakened by the turbulent transport as a result of its short range. During summer the spreading of the halocline can be clearly seen in observations (Table 1) and can be satisfactorily reproduced in simulations (Fig. 3). In winter this effect is not visible because there is no thermocline protecting the halocline from surface mixing. Accordingly, the surface mixed layer reaches down to the halocline, and erosion due to wind mixing and convection acts effectively in sharpening the salinity gradients in the halocline region. Consequently, the salt transported through the halocline by turbulent diffusion is instantaneously mixed to the entire surface mixed layer in winter. In summer surface mixing by wind and by nightly convection only reaches down to the thermocline which protects the halocline from erosion. As a consequence, the weakening of the halocline due to turbulent diffusion is undisturbed and the corresponding salt transport can be observed in terms of the slow upward expansion of the halocline during summer.

The amount of salt, which is eroded from the halocline by surface mixing in winter, is determined by the maximum depth to which the surface mixing reaches. This depth in turn depends on the meteorological conditions in winter and the vertical density gradient at the top of the halocline resulting from the weakening and expansion of the halocline due to the turbulent upward salt transport during summer. The eroded salt is mixed homogeneously into the entire surface mixed layer. In Section 3.2 the corresponding increase of the bulk sea surface salinity was used to estimate the amount of eroded salt to be slightly above 30 kg/(m² a) in the long-term mean. On the one hand, this is in good agreement with the other bulk estimates of the vertical salt transport through the halocline for the central Baltic Sea made in Section 3. On the other hand, it completes the consistent description of this transport with respect to the vertical transport within the halocline due to turbulent diffusion as discussed above and estimated in Section 5.1.

The outlined consistent description of the vertical salt transport in the central Baltic Sea is based on estimates of the turbulent salt transport within the halocline and the transport of salt into the entire surface mixed layer by erosion of the halocline. In spite of the excellent agreement of these estimates, their uncertainty, which can be a factor 2 for such estimates, would still allow for additional relevant vertical transport mechanisms as suggested in Section 5.

7.2. Specific characteristics during stagnation periods

On long time scales the salt loss below the halocline resulting from vertical upward transport is compensated by inflows of saline water from the North Sea at the bottom (Section 2.1). These inflows do not only sustain the salinity below the halocline at its long-term mean, but also reset the halocline depth to its climatological level by a corresponding vertical uplift (Section 3.3).

During stagnation periods without such inflows, the halocline is slowly lowered due to the alternating weakening and erosion of the halocline as discussed above. In addition, the salinity continuously decreases both below and above the halocline (Fig. 4).

Below the halocline this decrease obviously results from the continuing upward salt transport. Above the halocline the decrease in salinity is about a factor 2 slower than below it. This decrease of the bulk sea surface salinity may be attributed to the increase of the surface mixed layer volume resulting from
the lowering of the halocline, because the salt transported through the halocline is consequently mixed into increasing volumes of brackish water. Presuming that the environmental parameters such as freshwater surplus are almost the same during stagnation periods, this produces decreasing bulk sea surface salinities. However, the volume changes due to the lowering of the halocline can be assumed to be small relative to the total volume of the surface mixed layer. Consequently, this volume effect is expected to be of minor importance. Alternatively, the decrease of the bulk sea surface salinity may be a result of a reduced vertical salt transport through the halocline during stagnation periods. Such a reduced vertical salt transport is indicated by the about five times longer deep water residence time during stagnation periods derived in Section 3.4. Therefore it is the more probable cause of the bulk sea surface salinity decrease compared to the volume effect. Additionally, decadal fluctuations in river discharge and in the precipitation/evaporation balance influence the surface salinity.

The shorter deep water residence time during periods with small or major inflows of saline water at the bottom is associated with the corresponding uplift of the deep water and the halocline. However, the nature of the cause for the shorter deep water residence time during periods with such uplifts cannot be inferred solely from their occurrence at the same time. The shorter residence time and the corresponding intensified vertical mixing may be a direct effect of the vertical uplift as suggested in Section 3.4, but an indirect correlation of these phenomena seems to be somewhat more likely, because the inflows also import mechanical energy on various scales. This energy import and some of its consequences are illustrated in Section 5.4. The imported energy has the potential to drive or intensify nearly all mechanisms of vertical mixing discussed in Section 5. Only the vertical mixing by upwelling discussed in Section 5.5, the surface wave mixing, and the winter convection are likely to be almost independent from the energy fluctuations imported by the inflows. The vertical mixing due to the internal wave field (Sections 5.1 and 5.2) and the eddy activity (Section 5.3) is intensified during inflow events as well as the near-bottom currents (Section 5.4). Therefore the overall diffusivity directly or indirectly resulting from these phenomena most likely is increased by inflows and the corresponding energy fluctuations such as pulsating currents or the irregular uplift of the deep water. In contrast, the diffusivities would drop to a lower level during stagnation periods. Consequently, the resulting deep water residence time would be longer compared to periods with inflows.

Unfortunately, the actual quantification of the involved effects does not allow for a decision, whether these effects are sufficient to account for the difference of the deep water residence times derived from observations in Section 3.4. Note that if they were, the vertical mixing induced by upwelling would have to adjust the two different deep water residence times to their absolute values, because this is the only transport mechanism through the halocline, which can be presumed to be almost independent from inflows.

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