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Oceanographic Processes in the Baltic Sea

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Contents

1. General Description

The Baltic is a tideless semi-enclosed sea consisting of several basins, which are connected by sills or channels. A map of the bathymetry of the Baltic is shown in Fig. 1. The Baltic has only narrow connections to the North Sea through the Danish straits which reduce the water exchange distinctly. Water transport is also hindered by the shallow Darss Sill (maximum depth 18 m). This sill is the western boundary of the Arkona Basin (maximum depth 45 m), which is connected by the Bornholm Channel to the Bornholm Basin (depth maximum 95 m). At the eastern flank of the Bornholm Basin, the Stolpe Channel follows, which is connected to the central Baltic Sea, the Gotland Basin with a maximum depth of 250 m. The northern part of the basin extends to the East into the Gulf of Finland and towards the North, through the Aland Sea with its numerous islands, into the Gulf of Bothnia.

Being located in the latitudinal range from 54 to 66°N, the Baltic Sea is exposed to sub-arctic climatic conditions in the North and temperate conditions in the central and southern parts, and subject to highly variable winds and many storm events. The northern parts of the Baltic are regularly ice covered during winter. General overviews of the oceanographic features of the Baltic Sea are given, for example, in Omstedt (2004), Stigebrandt (2001), Kullenberg (1981) and Feistel et al. (2008).

Owing to the substantial fresh water input from several rivers, the Baltic Sea has a positive water balance. The annual total contribution of all rivers is about 450 km^3 while the excess of precipitation over evaporation amounts to 60 km^3 per year. This would raise the sea-level by a little more than 1 m per year if the Baltic Sea were not connected to the North Sea. As a result of the freshwater surplus and the restricted water exchange, the Baltic is one of the largest brackish water seas in the world with a water volume of about 21,000 km^3.

In the eastern and northern parts of the Baltic Sea the salinity is low (about 4 g/kg), but increases towards the Danish Straits. Owing to the freshwater supply and the near bottom inflow of saline water, the Baltic is characterized by a strong vertical stratification,

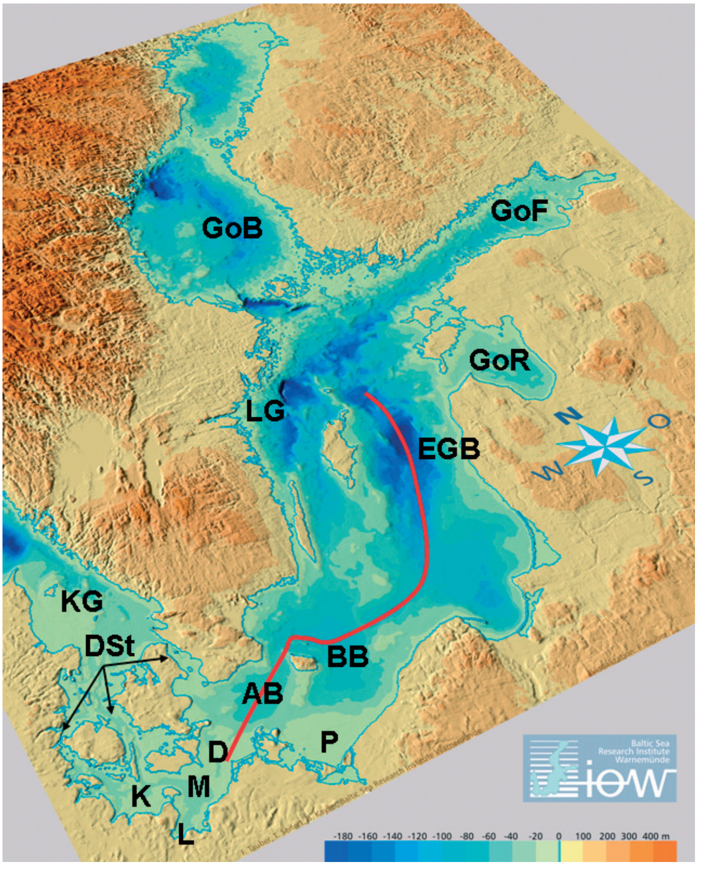

Fig. 1: Topographic map of the Baltic Sea (SEIFERT et al., 2001): Kattegat (KG), Danish Straits (DSt), Kiel Bight (K), Lübeck Bight (L), Mecklenburg Bight (M), Darss Sill (D), Pomeranian Bight (P), Arkona Basin (AB), Bornholm Basin (BB), Eastern Gotland Basin (EGB), Landsort gauge (LG), Gulf of Riga (GoR), Gulf of Finland (GoF), and Gulf of Bothia (GoB). The bold red line indicates the ship track for the records shown in Figs. 2 and 3

where a strong halocline separates the brackish surface water from the more saline bottom water.

The halocline is associated with a high vertical stability, implying a weak vertical exchange through the halocline. As a consequence, hypoxia or even anoxia occurs frequently in the deep waters below the halocline. The general structure of the salinity and oxygen distribution is shown in Fig. 2 by means of a section along the path of the maximum depths. The poor ventilation of the deep water below the halocline favours variations in oxygen and occasional switching to hydrogen sulphide. This affects the biogeochemistry in the bottom water and the upper sediments.

The annual cycle of solar radiation drives a yearly cycle of surface temperatures and the formation of a seasonally varying thermocline. Hence the vertical temperature distribution shows a strong seasonal cycle with the formation of a warm upper layer of 10 to 20 m thickness and a strong thermocline below. Examples of the temperature distributions along the

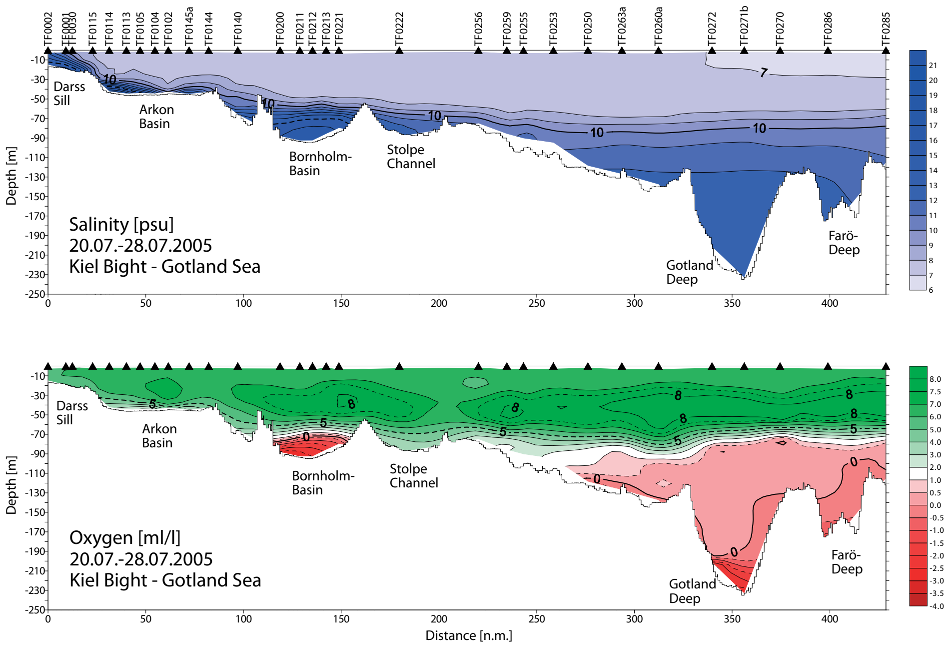

Fig. 2: Horizontal-vertical sections of salinity (top) and oxygen (bottom) as recorded during a Baltic monitoring cruise in 2005 (see Fig. 1). Negative oxygen is used to indicate hydrogen sulphide

path of the maximum depths in winter and summer are shown in Fig. 3. Vertical profiles of temperature and salinity show usually stronger gradients than suggested by the vertically and horizontally interpolated distributions plotted in Figs. 2 and 3.

The permanent halocline and the seasonal thermocline are important for the ecosystem of the Baltic Sea. The horizontal and vertical salinity gradients associated with the permanent halocline set the long term conditions for the marine ecosystem. Salinity gradients affect osmotic regulations of living cells and, therefore, the number of species in the Baltic is relatively small. The annual cycle of temperatures, i.e., the formation of the thermocline in spring and its disappearance in fall, defines a seasonal time scale. Irregular weather patterns and storms with time scales of a few days are typical for the latitudes between 50 to 60°N and generate a variety of mesoscale currents in the Baltic Sea, such as eddies, coastal jets and associated up- and downwelling patterns with time scales of a few days to weeks. Mesoscale features can influence the distribution patterns of plankton, eggs and larvae etc., because particulate biomass can either be dispersed or concentrated by the combined effect of swimming or sinking in conjunction with advection and turbulent diffusion.

2. Processes and Time Scales

The most energetic part of the currents is due to wind forcing. Frequent synoptic-scale cyclonic activity is typical in the Baltic area and implies a high variability of the prevailing westerly winds. Maximum speeds are reached during November to March. In May and June

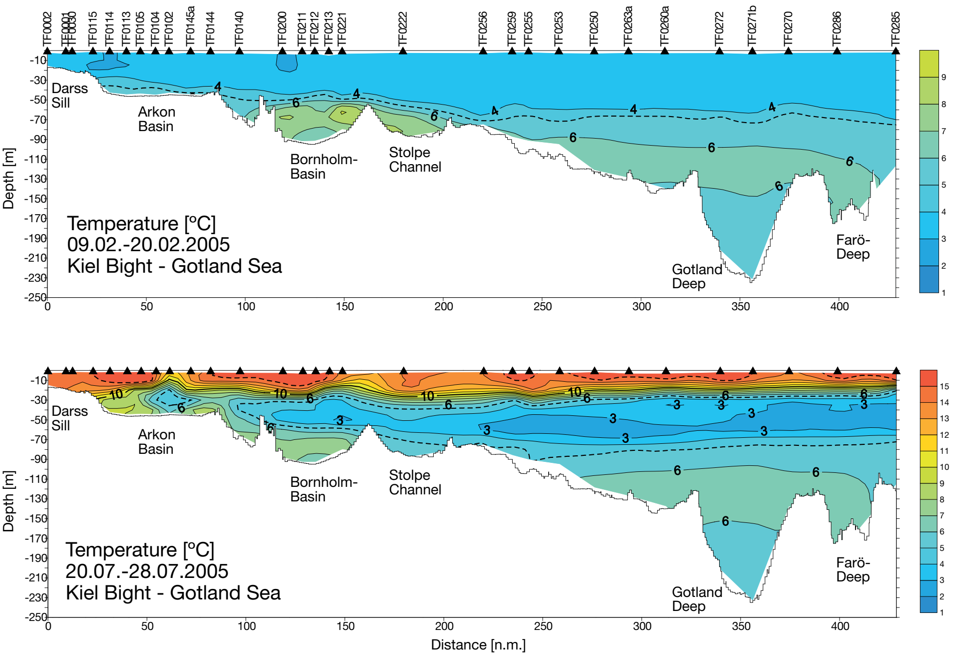

Fig. 3: Horizontal-vertical sections of temperature as in Fig. 2 as recorded during Baltic monitoring cruises (see Fig. 1) in winter (top) and in summer (bottom)

the average wind is minimum. In the high frequency range the wind generates waves which produce turbulence in the upper mixed layer and can mix the whole water column in the shallower regions.

2.1 Waves and Seiches

Compared to sea level elevations by tides and winds in the open ocean, the sea surface variations of the Baltic Sea are rather weak. Owing to its small spatial extent the tidal range is limited to some centimeters only (MÜLLER-NAVARRA, 2003), and the tidal signal propagating from the North Sea into Skagerrak and Kattegat is filtered out by the narrow and shallow Danish straits.

Therefore, the sea level of the Baltic Sea oscillates like in closed basins (so-called seiches). According to WÜBBER and KRAUSS (1979) the first mode comprising the entire Baltic has a period of 31 hours and forms a cyclonically rotating wave with the amphidromic point in the northern Baltic proper. Hence, the sea level observed at the tide gauge at Landsort is a reliable measure for the filling level of the Baltic Sea. The second mode shows two separate seiches of the Bothnian Sea and of the central Baltic including the Gulf of Finland and the Gulf of Riga with periods around 26 hours. The relative change of sea level by seiches is 5–10 cm in the open basins and up to 30–40 cm in the western Baltic, the Bothnian Bay, and especially in the eastern Gulf of Finland.

Stronger surface elevations are generated by wind set-up (Windstau). For example, long-lasting easterly or westerly winds can change the filling level of the Baltic Sea by ±50 cm on average. The wind induced piling-up of water at the coasts may rise to 1 m and more. Extreme high water levels can be generated when such a set-up flows back in resonance with the seiches, while a storm from the opposite direction forces a wind set-up at the other side of the basin. This was the case in November 1872, when a floodwater of 2–3 m height damaged the German and the Danish coasts (ROSENHAGEN and BORK, 2008).

The height of the surface waves, which are produced by the exchange of momentum between the atmosphere and the sea, increases essentially with the wind speed and with the fetch, which is the directional length over which a nearly constant wind acts upon the sea surface. The fetch is limited in the relatively narrow Baltic Sea to some 200–700 km along the main axes of the open basins. Wave models show that, despite of the fetch limitation, a saturated sea state develops for stronger winds within 12–24 hours with significant wave heights of 9 m in the central Baltic and up to 7 m in the south-western Baltic. In correspondence to the seasonal variation of the wind, the monthly mean wave heights vary between 1–3 m in winter (October–February), whereas the average is less then 1 m in summer (see for instance JOENSSON et al., 2002).

In Fig. 4 the mean significant wave height (in colours) is shown as taken from a simulation for the decade 1990–1999. The computation was done with an adaptation of the parametric wave model of SCHWAB et al. (1984), using a resolution of 3 nautical miles (approximately 5.5 km). The maximum of the simulated waves are indicated by contours (full lines for 1–4 m, and broken lines for half meter steps). To validate the model results, data from a wave rider buoy were used. The wave rider was operated by GKSS (Institut für Küstenfor-

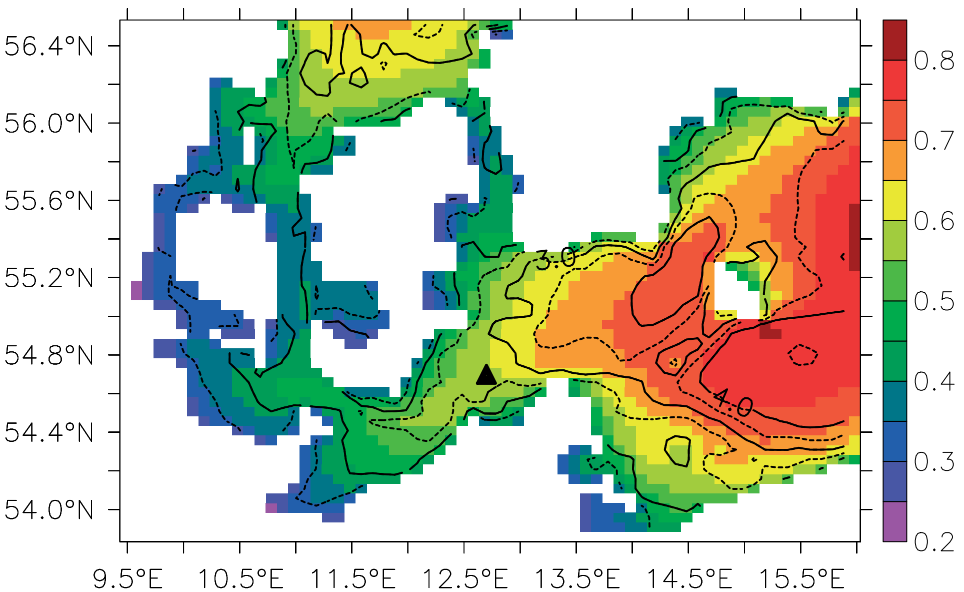

Fig. 4: Mean significant wave height (colours, in meters) and maximum waves (contours) in the period 1990–1999 simulated with an adaptation of a parametric wave model (SCHWAB et al., 1984) with ERA-40 wind forcing (UPPALA et al., 2005). The triangle shows the position of the GKSS wave rider buoy used to validate the model results

schung, Geesthacht) in the frame of the MORWIN project (http://morwin.baw.de) and was deployed near the permanent monitoring station on Darss Sill, the location of which is indicated by a triangle symbol in Fig. 4. The data comparison showed that the significant wave height is reproduced by the wave model with a scatter of ±0.25 m explaining 70–80 % of the data variance. On average the simulated wave heights under-estimate the real sea state by 20 % (slope of 0.8 of the regression line between model and data). That is partly caused by the wind forcing taken from the reanalysed weather model data set ERA-40 (UPPALA et al., 2005), since transient local wind maxima are smoothed out by six-hourly forecasts within a spatial resolution of 60–120 km.

Despite of these restrictions, the wave simulation shows realistic mean wave heights below 0.9 m in the south-western Baltic. In the sheltered bights of Kiel and Lübeck a low sea state is found. Maximum significant wave heights of 2–3 m are obtained in the Mecklenburg Bight, and up to 4 m in the Arkona Sea. However, the assessment of wave action in the coastal zone requires the application of more elaborate models, including swell and wave breaking, with a significant higher resolution of the model grid and the forcing data. For a recent comprehensive overview of waves, tides and seiches in the Baltic Sea see SCHMAGER et al. (2008).

2.2 Currents

Although the winds can be strong during gales, the variance of the wind is high while mean values are small. Consequently, the currents are dominated by the transient phases of the oceanic responses to episodic forcing. Virtually there are no significant permanent currents which deserve their own name. Weak general circulation patterns can be detected by the large scale distribution of surface salinity, which points to a small basin-wide cyclonic water motion. In the central Gotland Basin, a relatively persistent low frequency dense bottom flow has recently been detected by HAGEN and FEISTEL (2004).

Strong current signals are often observed in the western Baltic and in the Danish straights, where the flow is driven by local and remote forcing. In particular, large scale sea level gradients between the Kattegat and the Baltic Proper can generate strong gradient currents in the narrow channels of the transition area between the Baltic and the Kattegat even at low wind episodes.

The responses of the sea to wind forcing are in particular seen as coastal jets, usually associated with up- and downwelling patterns (see e.g. FENNEL and STURM, 1991; FENNEL and SEIFERT, 1995 and LEHMANN et al., 2002). The upwelling is well documented by satellite images of the sea-surface temperature (SST), where the cold intermediate winter waters wells up near the coast and leave a very clear signature (see the example shown in Fig. 5).

The spirals and eddy-like patterns have typical scales of 5 to 20 km and demonstrate a rich mesoscale circulation. Since the spatial scales of the meteorological forcing and irradiation are much larger, say several hundreds of kilometres, it is clear that the mesoscale structures reflect the response of the sea. The upwelling signal off the Polish coast refers to the typical response of the coastal waters to eastern winds. The cold water area in the south-western Baltic shows upwelled water that spreads over the entire channel due to the off-shore Ekman transport.

An illustration of the high variability of the current signals in the channel of Fehmarn Belt is shown in Figs. 6 and 7. In Fig. 6, the three snapshots of the currents through the Fehmarn Belt, observed in November 1994, show inflowing waters. However, the current is not homogeneous but varies over the cross section over a range of up to 100 cm/s.

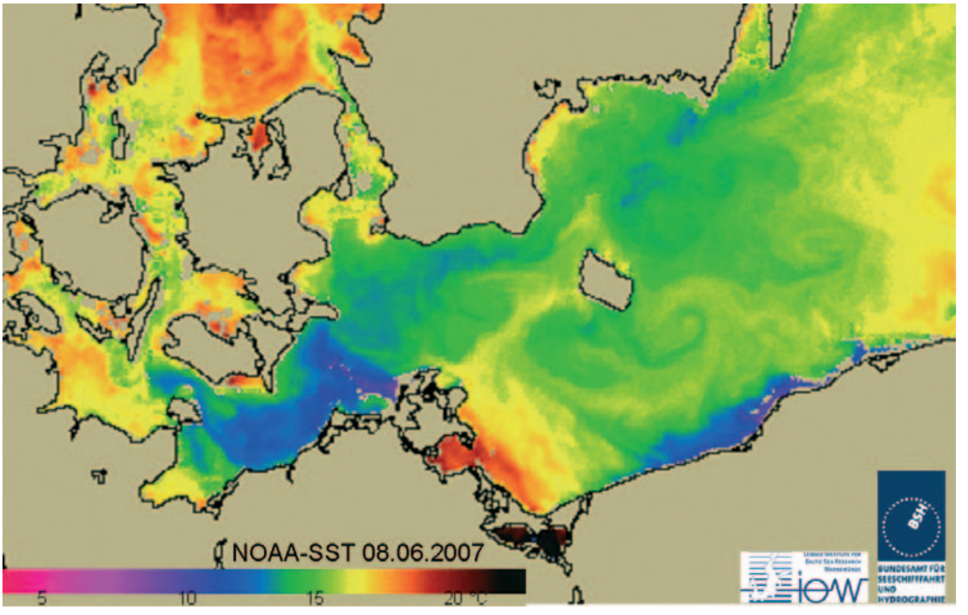

Fig. 5: Satellite image of the south-western Baltic in June 2007.
The blue areas indicate cold upwelled water

The second example, shown in Fig. 7, was observed in summer 1995. The saline stratification is stronger than in autumn, and the isohalines indicate a geostrophic balance of the currents and the cross-channel pressure field. The current has a remarkable spatial cross-channel structure and varies in the vertical and horizontal direction. The speed ranges from –80 to +80 cm/s.

2.3 Adjustment Processes

It is illuminating to look at the response of the sea to a sudden onset of wind. The scenario starts with inertial oscillations and inertial waves along with Ekman transports. Coastal boundaries and wind stress curls generate convergences or divergences of the Ekman transport, which produce vertical water motions and generate pressure gradients and geostrophically adjusted flows, such as coastal jets. Eventually, the adjustment of the flow results basically in currents along isobaths.

The response scenario can be characterized by several time scales: the geostrophic adjustment is related to the time needed by inertial waves to travel along the channel-like Baltic Sea (crossing time = c_1 L, where c_1 is the first mode phase speed of inertial and Kelvin waves and L is a cross-sectional length (scale) of the Baltic channel. The phase speeds vary both spatially among the basins and seasonally due to the cycle of the thermocline (FENNEL et al., 1991). The adjustment time scales range from about one day in the Arkona and Bornholm basin to about two days in the central Baltic. The coastal flows are also shaped by Kelvin waves, or more general, topographically trapped waves, which propagate around the basins and set up alongshore pressure gradients.

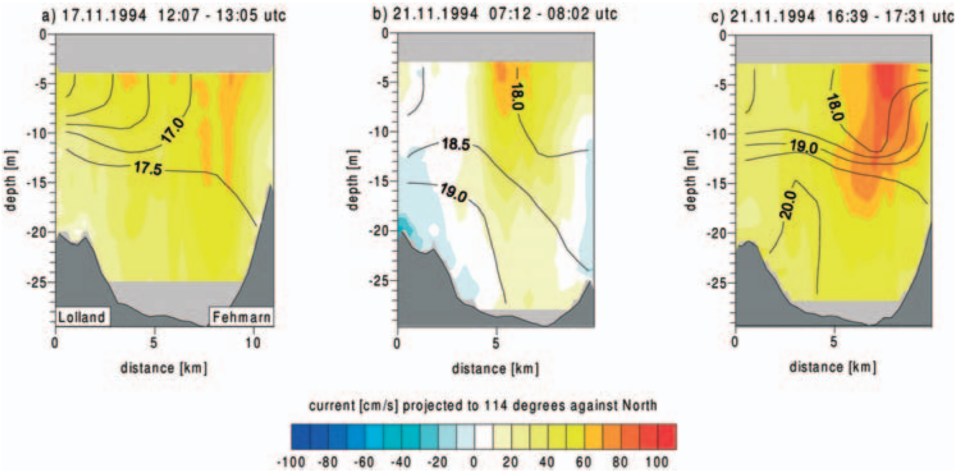

Fig. 6: Three snapshots of salinity stratification (contours) and flow (colours) through the Fehmarn Belt during autumn conditions. The flows show basically an inflow, but the flow pattern is structured and varies over a range of almost 1 m/s

The corresponding time scales are given by the travel time around the basins and are about 12 days in the Arkona Sea, 9 days in the Bornholm Sea, and 27 days in the central Baltic. These time scales often exceed the synoptic scales of weather patterns, which range from a few days to a few weeks.

Semi-enclosed systems can globally be characterized by residence times, which can be derived from the ratio of the water volume to inputs or outputs in terms of fluxes through the system boundaries. For the Baltic Sea, the residence time varies in the range between 15 to 35 years, depending on the choice of in- or output. However, this type of timescale does not represent the residence time of constituents, say dissolved nutrients, in a water parcel. The fate of a water parcel is affected by a variety of different processes acting on different temporal and spatial scales. Apart from the physical transport processes, i.e., advection and diffusion, the characteristic time scale of nutrients depends also on the way how matter is cycling through the food web. Hence attempts to characterize semi-enclosed systems by just a few numbers should also reflect the physical biological interaction.

2.4 Salt Balance and Major Inflows

The salt balance is maintained by a dynamical equilibrium between near-bottom inflow of saline water, outflow of brackish surface water and vertical salt diffusion through the halocline. From the conservation of mass (Knudsen theorem) it can be derived that the annual outflow of brackish water is approximately 1,400 km^3, while the near bottom inflow of saline water is about 830 km^3 per year.

The deeper water can only be ventilated through horizontal propagation of dense saline bottom water, which is formed in the transition areas between the North and the Baltic Sea and cascades over the sills until it arrives in the central basin. In particular, irregularly occurring major salt water inflows renew the deep water in the central Baltic (e.g. MATTHÄUS and FRANCK, 1992). The inflow of dense saline deep water near the bottom occurs in particular during major inflows when a substantial water volume passes the Darss Sill and advances through the Arkona Basin into the Bornholm Basin. In the Bornholm Basin the dense water must first fill the basin till the depth level of the Stolpe Channel (60 m) before it can propagate further into the central Baltic, where it can renew the bottom water.

If the density of the inflowing water is less than that of the bottom water, it will replace the water sheet of equivalent density within the halocline. This 'interleaving' pushes old water upwards while ventilating the halocline. Since major inflows are initiated by strong winds, which imply strong vertical mixing, a substantial part of the saline water will be mixed with brackish surface water in the Arkona Sea and leave the Baltic (MATTHÄUS and LASS,

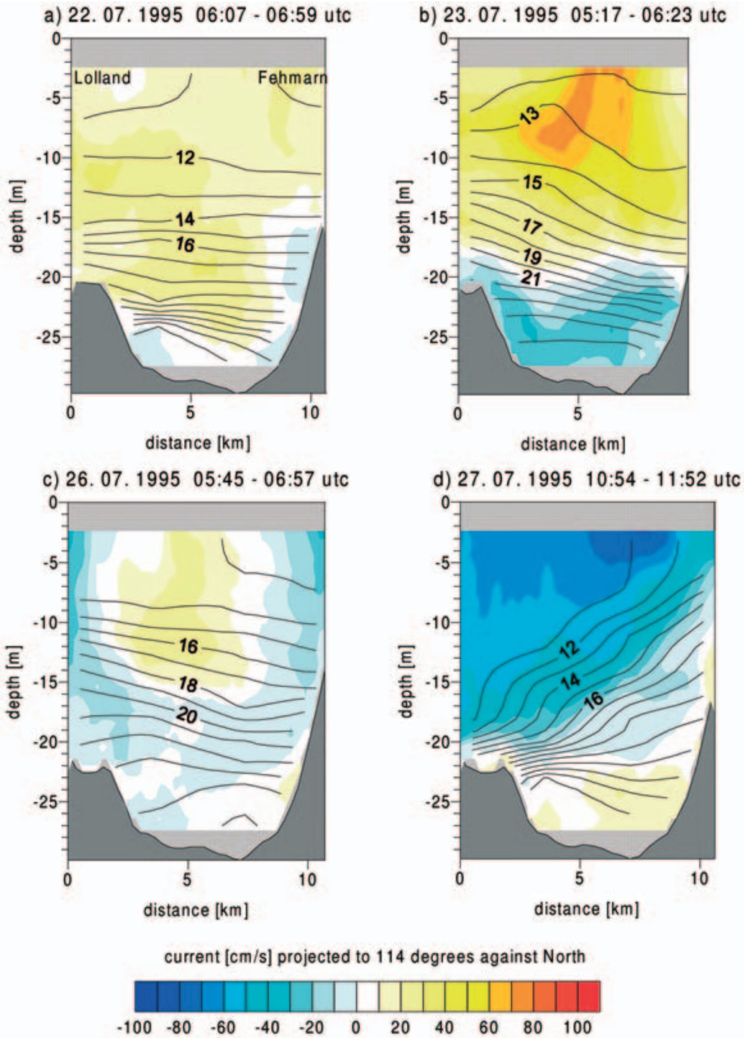

Fig. 7: Four snapshots of salinity stratification (contours) and flow (colours) through the Fehmarn Belt during summer stratification. The shape of the isohalines indicates the geostrophically adjusted status of the currents. The speed ranges from +80 to –80 cm/s, for inflow and outflow, respectively

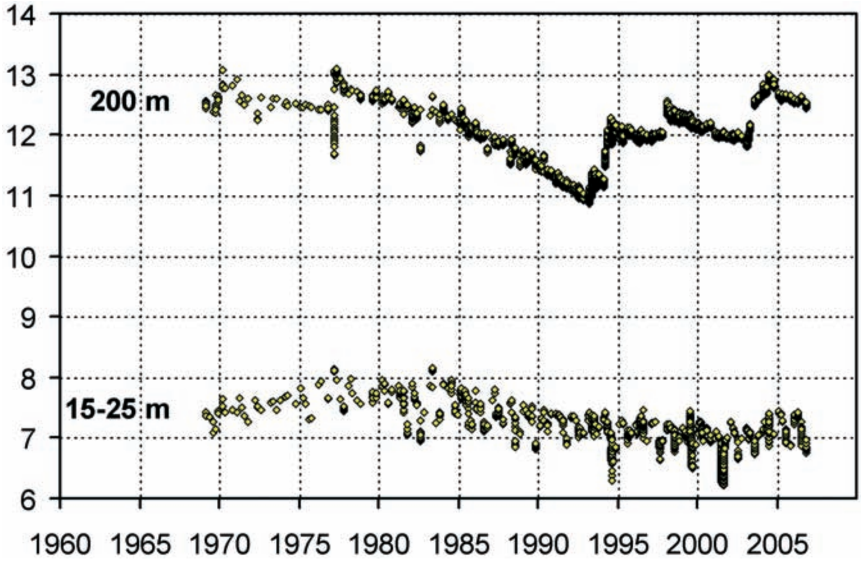

Fig. 8: Time series of salinity in the deep water and near the surface at a station in the central Baltic. The data are taken from the data base of the IOW

1995). The time scale of the propagation of inflowing dense bottom water toward the Gotland Basin ranges from a couple of weeks to several months.

Examples of the effect of salt water inflows into the deep waters of the central Baltic Sea are shown in Fig. 8 in terms of a long time series of salinity observed in 200 m depth. A strong major saltwater inflow occurred in 1976, as is indicated by the sudden increase in salinity. This event was followed by an unusual long stagnation period until in 1993 a new major inflow was observed. After this event, the sudden slight increases in salinity of the bottom near waters give a clear indication of a series of smaller inflows during 1997 to 2005.

A comparison of the long time series of salinity near the surface and the deep water, as shown in Fig. 8, implies an upward transport of salt. The decrease of salinity of the deep water after the inflow of 1976 till the next inflow in 1993 is also reflected in the surface salinity, but starts with a delay of about 10 years. This is a clear indication of a slow effective upward salt transport (see FEISTEL et al., 2006). The near-surface salinity shows also strong seasonal variations that reflect freshwater pulses due to seasonally varying river discharges and melting of ice.

The effective vertical diffusion is to some extent caused by vertical turbulent diffusion through breaking internal waves, which may be generated by the interaction of inertial oscillations and topography (e.g. FENNEL and SCHMIDT, 1992). Further mechanisms that drive upward salt transports are displacement of water layers in the halocline through interleaving of inflowing waters and upwelling of the halocline at steep slopes due to Ekman recirculation.

The freshwater reaches the Baltic mainly through river runoff. The river plumes propagate like Kelvin waves alongshore with the coast to the right. Without mixing and water transformation the buoyant plumes would stay near the coast, but due to wave induced mixing, wind driven alongshore flows and cross shore Ekman transports the river water is mixed and entrained into the brackish surface waters (see e.g. FENNEL and MUTZKE, 1997).

2.5 Sea Surface Temperature and Sea Ice

Synoptic basin-wide measurements of the sea surface temperature (SST) of the Baltic Sea have become available with the deployment of thermal radiation sensors on polar orbiting satellites since the 1980ies. SIEGEL et al. (2006; 2008) have compiled and analysed a continuous series of monthly mean SST fields of the Baltic Sea for the period 1990–2005 and found a distinct seasonal cycle with minimum temperatures of 0–2 °C in ice free areas during February and March. The warmest months are July and August when the temperature reaches 16–18 °C on average with an inter-annual variation between 15–20 °C. During the 16-year period of investigation, the annual mean SST reveals a positive trend of approximately 1 °C, which is above the average warming of the northern hemisphere. The increase of the SST varies regionally and seasonally. The strongest warming in the central Baltic and the Gulf of Bothnia occurs during summer and autumn implying a later onset of the ice formation. Similar SST trends have also been derived by MACKENZIE and SCHIEDEK (2007) from long-term observations. However, the Baltic SST shows significant correlation with climate indicators like the North Atlantic Oscillation index (NAO) only during the winter, especially March, when a slight decrease of the monthly mean surface temperature is observed.

Sea ice is formed if the SST drops below the freezing point. Owing to the low surface salinity freezing starts in the Baltic just below zero Celsius, and sea ice covers a considerable part of the sea during each winter. However, the ice thickness and the ice extent show strong regional and inter-annual variations because of the relatively large latitudinal extent of the Baltic Sea. Heavy ice conditions occur every year with a probability of 70–100 % in the Bothnian Sea, in the Gulf of Finland, and in the Gulf of Riga, whereas in the deep basins of the central Baltic Sea ice covers are observed only during very severe winters (see e.g. SEINÄ and PALOSUO, 1996). In the shallow south-western Baltic ice formation occurs if high pressure over Scandinavia or Russia leads to prolonged cold northerly and easterly winds, which force an outflow of low saline Baltic surface water. But the more frequent weather pattern are troughs associated with strong westerly winds that bring warm air from the Atlantic and push saline Kattegat water into the Baltic. Consequently, the open Arkona Basin and the Mecklenburg Bight remain usually ice free in wintertime. Only in the inner waters along the German coast fast ice of a thickness of 10–50 cm occurs regularly (SCHMELZER, 1996 and the overview in SCHMELZER et al., 2008).

The ice cover of the Baltic Sea can well be reproduced by models (HAAPALA and LEPPÄRANTA, 1996; MEIER et al., 2002a, b; LEHMANN and HINRICHSEN, 2000 and SCHRUM et al., 2003). As an example, the maximum yearly ice extent during the period from 1961–2004, calculated by the IOW Baltic Sea model, is shown in Fig. 9. The model is based on the Modular Ocean Model (MOM–3.1, PACANOWSKI and GRIFFIES, 2000) and the thermodynamic three-layer ice model of WINTON (2000). In comparison to the data compiled by SEINÄ and PALOSUO (1996) the simulation follows the essential inter-annual variations, see Fig. 9 (left panel). The yearly difference in ice extent is also realistically simulated for the south-western Baltic. During the two severe winters of 1962/63 and 1995/96 70–85 % of the area were covered by sea ice, whereas in a mean season like 2002/03 only 10–15 % are covered, as shown in Fig. 9 (right panel). Moreover, both the duration of the ice season (from mid-December till March) and the occurrence of relative maxima in ice extent (varying inter-annually

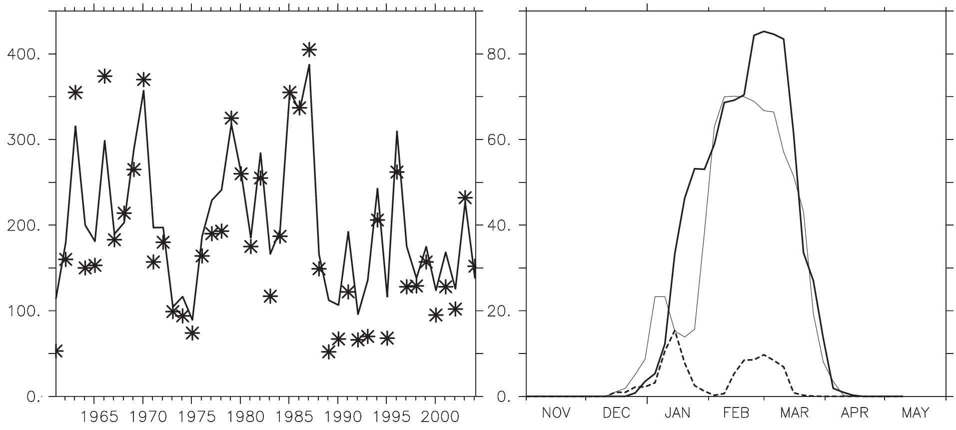

Fig. 9: Yearly maximum ice extent (10^3 km^2) of the Baltic Sea (left panel); the bold line indicates the simulated ice cover and star symbols show the estimates by SEINÄ and PALOSUO (1996). Relative model ice extent (%) in the south-western Baltic (right panel); bold, thin, and dotted lines correspond to the winters 1962/63, 1995/96, and 2002/03

from early January to end of February), are in agreement with the observations (see for example SCHMELZER et al., 2008).

2.6 Transport of Suspended Material

For the transport of matter from the river plumes towards the central basins, the main physical mechanisms are currents across the isobaths, in particular the non-geostrophic Ekman-transport.

For the transport of suspended sedimentary material from river mouths to the accumulation areas in the deep basins, fluffy layers are important. Fluffy material is assumed to consist of flocks which accumulate at the sea bottom and are easily eroded if the bottom shear stress exceeds 0.02 N/m^2 (CHRISTIANSEN et al., 2002). Strong wind events generate wave mixing in shallower areas from the surface to the bottom and re-suspend the fluffy layer material, which is then transported into the basins in a few days or weeks time. Calm situations and strong summer stratification imply a slow exchange between the coastal zone and the open sea, as was shown in KUHRTS et al. (2004) and SEIFERT et al. (2008) by means of experiments with a numerical model.

The relative frequency of fluff erosion events in the south-western Baltic is shown in Fig. 10, as derived from a model simulation with a high spatial resolution of approximately 2 km. The reference period December 1992 till October 1993 was chosen for a series of strong wind events, especially during the winter season. The bottom shear stress is evaluated from an adaptation of the wave boundary layer after GRANT and MADSEN (1979) and the skin friction layer after SMITH and MCLEAN (1977). A detailed analysis of the model fields shows that the oscillatory motion induced by surface waves is the predominant contribution to the bottom shear stress. Therefore, the probability of erosion events decays exponentially with water depth. Fig. 10 shows that no erosion of fluff is calculated for depths below 40 m, whereas in the shallow areas above 20 m erosion events are rather frequent. As a rule-of-thumb, one

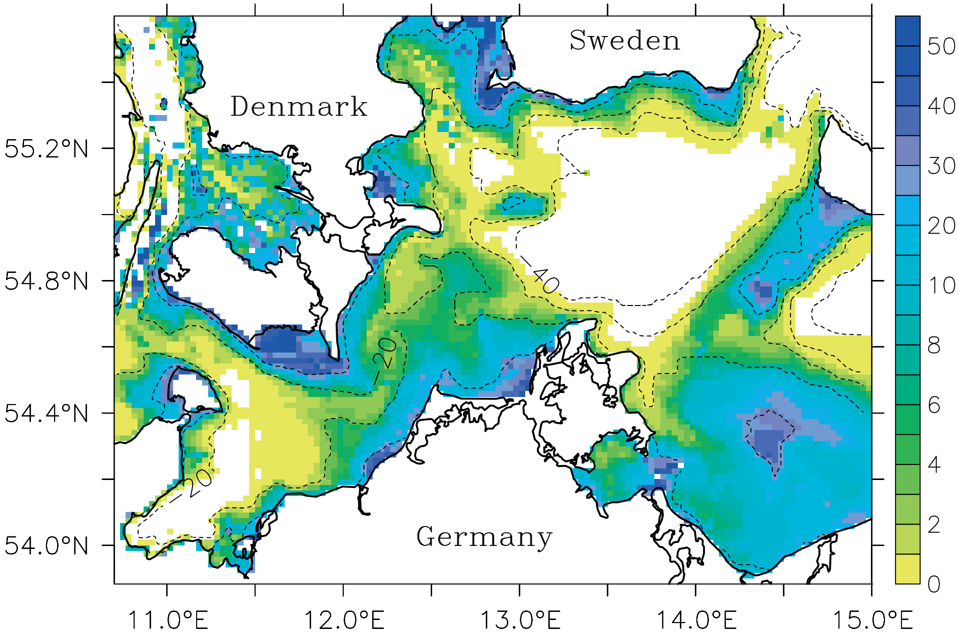

Fig. 10: Modelled frequency of occurrence (%) of fluff erosion in the period Dec. 1992 till Oct. 1993. Broken lines indicate the water depth in 10 m steps

may assume that fluffy material will be washed off from regions above 20–40 m if the wind speed exceeds 10 m/s for at least 2 days. Because of its sheltered location, the inner Lübeck Bight is an accumulation area already below 20 m water depth. In contrast to that, the coastal zone off Fischland, Darss and Hiddensee appears to be a rather active area of sediment transports. Strong winds above 15 m/s are necessary to move fine particulate matter and grainy sediments such as sand. Since the threshold for incipient motion is above 0.2 N/m² the resuspension and bed-load transport of such material occurs only in regions above 20 m depth with a probability of less than 5 % of the simulation period.

3. Acknowledgement

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