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Observations of the mixing in the Baltic thermo- and halocline layers

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ABSTRACT

The vertical and horizontal mixing of a passive contaminant in the Baltic thermo- and halocline layer has been investigated in a series of experiments during spring and fall. The vertical mixing is very weak. The horizontal spreading is a factor of 2–5 lower than in the surface layer. The integrated vertical transfer across the halocline layer is of the same order of magnitude as the integrated transfer in the coastal boundary zone. It is found that the wind is an essential source of energy for the mixing. The significance of the wind-forcing is demonstrated by the correlation between the fluctuations of the wind and (1) the fluctuations of the oxygen content in the deep water and (2) the fluctuations of the stability across the halocline layer, during the present century. The results support the view that the overall decrease of the oxygen content in the deep waters is due to other factors than an alteration of water exchange characteristics.

1. Introduction

The Baltic has a positive water balance, the fresh water supply being about 2% per year of the total volume, and the connections with the open ocean are shallow and narrow. As a consequence the stratification is very stable, with a year-round halocline layer at a depth of 50 to 70 m in the open Baltic. During most of the year there is also a thermal stratification. The importance of the Baltic as an environmental resource to the surrounding countries is evident. Several on-going research programmes are oriented towards understanding the Baltic ecosystem, the coupling between physical and biological processes and the possible interference of man (e.g. ICES, 1974). In this connection the turbulent mixing and transport of substances, the vertical exchange between various layers, as well as time variation and spatial distribution are of great interest. In this note some results on the vertical and horizontal mixing of a passive contaminant, obtained in the southern part of the Baltic, will be presented.

The mixing was studied experimentally by tracer technique. In connection with the tracing, environmental observations were carried out of salinity, temperature and current profiles so as to obtain an adequate time-series of profiles with a vertical resolution of 2–5 m. Usually these measurements were carried out by means of moored current meters and shipboard profiling from an anchored ship. Salinity and temperature profiles were taken also during tracing. During the intense experiment Baltic 75 (see, e.g., Keunecke et al., 1975) in the Bornholm basin, however, more sophisticated instrumentation such as automatic current profilers were used, providing a more detailed background information. In addition a series of surface layer experiments were performed during Baltic 75 using a photographic observation technique (Schott et al., 1976).

The tracer technique has been described in detail earlier (Kullenberg, 1971, 1974a). A density adjusted solution of rhodamin B dye is injected in a pre-selected subsurface layer of thickness 2–3 m. The injection is made by means of high pressure, forcing the solution via a hose out through a vertical diffusor closed at the lower end. During the operation, which takes 3–5 min, the ship is heaving to so that the hose is kept vertical. Before the injection a parachute drogue is released at the selected depth. The injection is made fairly close to

Table 1. *Characteristics of the experiments*

Date	Position of injection	Mean depth (m) of dye	Mean depth (m) of bottom	Mean wind (m/s)	Injected amount (kg)	Duration of tracing (h)	Mean velocity (m/h)	Drift direction (towards)
Present:								
May 8	N 55° 31.2 E 15° 34.8	40–45*	85	8	8	9	260	065°
May 9–10	N 55° 32.1 E 15° 35.8	35–55†	80	3	4	42	275	070°
Previous:								
1969								
Dec. 9	N 54° 58.5 E 14° 00.5	22–27‡	45–50	3	2	17	910	229°
Dec. 10–12	N 55° 04.5 E 14° 08.0	30–35‡	45–50	6	1	10	450	083°
1970								
Aug. 29	N 55° 32.6 E 15° 56.0	26–32§	85	4	1	11	220	064°
Aug. 30	N 55° 34.3 E 16° 03.5	48–53†	85	3	1	26	195	068°
Sept. 1	N 55° 26.0 E 16° 00.0	50–54¶	85	7	1	8	310	064°
Sept. 2	N 55° 33.0 E 15° 48.0	26–29**	85	3	0.5	11	95	270°

* Above halocline.
† Top of halocline.
‡ Thermohaline layering.
§ In thermocline.
¶ In halocline.
** Below thermocline.

the drogue surface buoy which is used for positioning. In most cases the drogue and the dye will drift with only a slow relative separation. When the separation becomes too large a second drogue can be released. The dye concentration distribution is mapped by means of a winch-operated *in situ* fluorometer (Kullenberg, 1969) towed behind the ship in an up-and-down (saw tooth) fashion. The concentration and the temperature distribution are continuously recorded on an analog recorder. The ship is cruising systematically over the area at a speed of 1–3 knots. The position is plotted continuously relative to the drogue as well as in Decca coordinates.

The experiments were performed in the Arkona and Bornholm basins and characteristics are given in Table 1. Those labelled present were carried out during the Baltic 75 experiment whereas the other ones have been partially reported on previously (Kullenberg, 1974a). The mean drifts of the dye spots illustrate the variability of the Baltic currents. The close agreement between the deep dye spots as regards both drifting direction and velocity is noteworthy. The variability appears to be more pronounced in the 20–30 m layer. Generally the winds were weak during the experiments with insignificant local influence on the mixing in the halocline layer.

2. Data reduction and analysis

2.1. *Vertical mixing*

From the dye observations vertical and horizontal mixing parameters are determined. Usually the vertical dye distribution is characterized by relatively thin layers, often pulse-formed, with sharp boundaries.

Some examples are given in Fig. 1. The shallow dye boundary normally does not show any lobes whereas the deep boundary often displays disturbances. This suggests that the upper layer contains more shearing agencies, keeping the dye layer sharp, than the deep layer. It also suggests that mixing goes downwards as well as upwards, and a plausible mechanism for this in the halocline would be by small-scale breaking internal waves

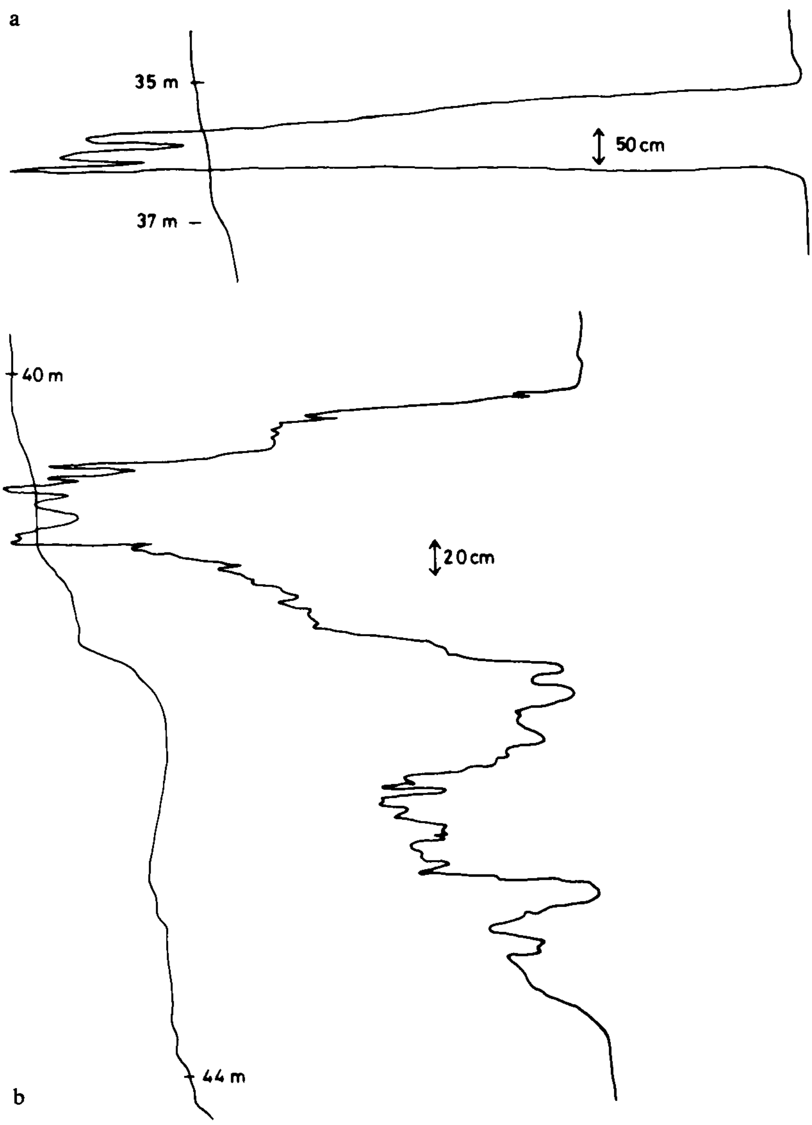

Fig. 1. Examples of profiles through dye layers obtained during the experiment 9–10 May 1975, with the instrument going downwards. Depth of injection 44–46 m. Relative temperature (thermistor) decreasing to the right.
a. Pulse-formed layer after 4.2 h of tracing
b. Interleafing layers after 14 h
c. Inversion structure after 17.5 h
d. Pulse-formed layer in temperature minimum after 30.3 h
e. Layers related to temperature decrease on top of halocline, after 34.6 h (right) and 34.8 h (left).

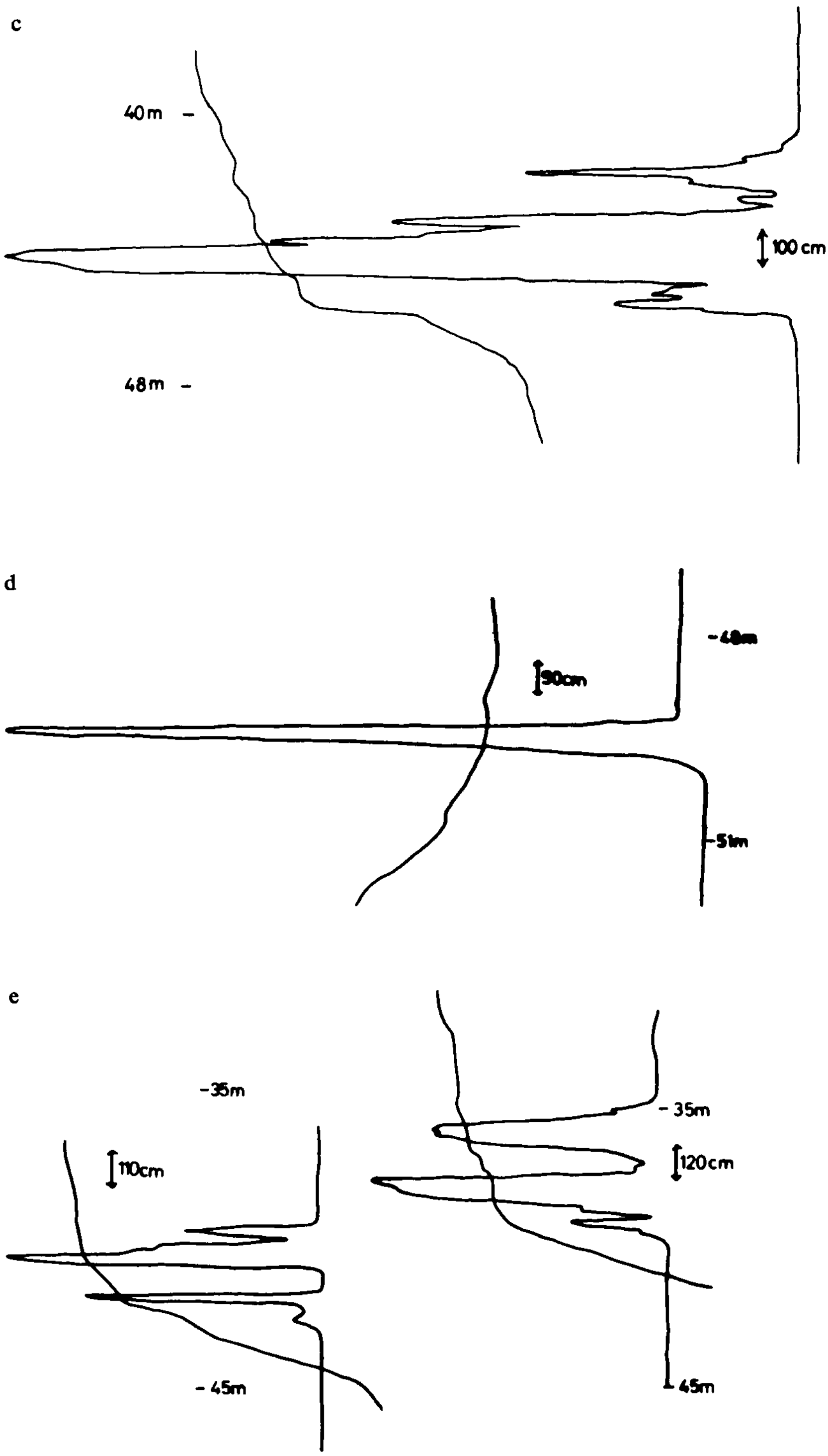

Fig. 1—contd.

(Woods, 1968). It is also clear that dye layers become trapped on relatively sharp temperature steps where they are gradually stretched by the shear.

For the pulse-formed layers Kullenberg (1971) developed a model of vertical mixing, assuming that the pulse form was maintained by the vertical current shear opposing the tendency of vertical mixing to create a smoother form. Over limited periods of time, the observed thickness is virtually constant, and in the model the thickness is treated as constant. From the model an effective vertical mixing coefficient K_z is determined

$$K_z = \frac{H^2}{\pi^2(t_2 - t_1)} \ln \frac{C_1}{C_2} \quad (1)$$

where H is the layer thickness, C_1 and C_2 the maximum concentrations at times t_1 and t_2, respectively. The expression has been used to calculate the mixing coefficients given in Tables 2 and 5a. The pulse-formed layer is identified by means of its shape, position in the temperature structure and thickness (Fig. 1). It is often possible to identify several layers of slightly different depths in different thermal structures.

2.2. *Horizontal mixing*

A smoothed horizontal concentration distribution is obtained by plotting the mean concentrations along the track of the ship and contouring the plot. For obtaining a reasonable distribution the track of the ship must cover the dye cloud in several directions. It is assumed that the distribution is horizontal, and no correction is made for the depth variations which may occur. An integration over the distribution, assuming an average thickness and vertical concentration, should yield an acceptable mass balance, regarded as such when in the range 50–100%. From the horizontal distribution various horizontal mixing parameters can be determined.

The observed distribution may be transformed into a rotationally symmetrical one by calculating the equivalent radius r for the circle corresponding to the area covered by a given isoline. The arguments behind this treatment are related to the structure of the horizontal motion in the sea, assuming that the mixing is due only to homogeneous horizontal turbulence (Okubo, 1962). This approach should be used with caution in near coastal waters.

There exist a number of rotationally symmetrical models describing the mixing by means of one single parameter (e.g. Okubo, 1962, for a review). The present author has used the Joseph–Sendner model (Joseph & Sendner, 1958), where the basic assumptions are that a mean diffusion velocity P can be defined, and that the mean square separation is proportional to the initial separation. The model distribution is

$$C(r, t) = \frac{M_0}{2\pi \cdot H_0 \cdot (Pt)^2} e^{-r/Pt} \quad (2)$$

where M_0 is the injected amount, H_0 the initial thickness, t time and r the distance from the moving point of injection. It is assumed to be given by the equivalent radius. The model requires that the contaminant is homogeneously distributed in a layer of given thickness which it does not leave. In reality the dye leaks out from the layer by vertical mixing and the initial thickness can be affected by both shear and vertical mixing. Thus

Table 2. *Vertical mixing coefficient for experiments during Baltic 75*

Date	Depth (m)	Thickness (m)	Δt (h)	C_1/C_2	$K \cdot 10^6$ (m^2/s)	Mean $K \cdot 10^6$ (m^2/s)	N^2 (s^{-2})
May 8	40–50	0.50	2.7	7.5	3.8		
	40–50	0.40	0.8	2.7	5.2	4.5	$9.0 \cdot 10^{-5}$
May 9	30–40	2.65	2.8	2.3	5.7		
May 10	30–40	1.00	6	5.4	8.0	6.8	$3.3 \cdot 10^{-5}$
	45–50	1.00	9	1.4	1.1		
	45–50	0.90	2.2	1.3	2.4	1.7	$3.1 \cdot 10^{-4}$
	50–55	0.20	2	11.1	1.4		
	50–55	0.20	1.3	1.7	0.8		
	50–55	0.20	3.3	18.8	1.0	1.0	$2.8 \cdot 10^{-3}$

Table 3a. *Horizontal diffusion velocity P from the concentration distribution, 9–10 May only*

Diffusion time (h)	Concentration $\cdot 10^{10}$ (ton/m^3)	Equivalent radius (m)	P (m/h)	Mean P (m/h)	Mass balance (%)
13.3	2	310	3.4		
	10	240	3.4		
	100	130	3.1	3.3	70
18.5	10	330	3.9		80
29.0	400	116	3.9		80
	100	280	4.0		
	10	400	2.4	3.4	
40.0	10	420	4.5		50
	5	510	4.0	4.3	

Table 3b. *Horizontal diffusion velocity P from observed values of maximum equivalent radius r_{max} and maximum diffusion time T_{max} for given concentration isolines, 9–10 May only*

Diffusion time (h)	Concentration $\cdot 10^{10}$ (ton/m^3)	Equivalent radius (m)		P (m/h)	Mean P (m/h)
		observed	predicted		
27.5	100	290	146	5.3	
47	100	0		4.3	4.8
28	200	245	103	4.4	
40	200	0		3.5	4.0
40	10	420	460	5.3	

the observed amount of dye and the observed thickness is used for calculating P.

The rotationally symmetrical variance σ_r^2, commonly defined as

$$\sigma_r^2 = \frac{\int_0^\infty r^2 C(r, t) r\,dr}{\int_0^\infty C(r, t) r\,dr} \tag{3}$$

is obtained by integrating over the observed transformed distribution. The Joseph–Sendner model predicts the relation

$$\sigma_r^2 = 6P^2 t^2 \tag{4}$$

which is also used to determine P for comparison with the results obtained from (2). By means of (2) other predicted characteristics, such as the maximum equivalent radius r_m and the maximum time T_m for a given concentration to exist, can be calculated and compared with observed values. Assuming that the diffusion velocity is constant, we find the time, T_m, when the equivalent radius for the isoline C approaches zero to be

$$T_m = \frac{1}{P}\sqrt{\frac{M_0}{2\pi H_0 C}} \tag{5}$$

The maximum expected radius r_m found by putting $dr/dt = 0$ is

$$r_m = 2Pt$$

$$r_m = \frac{1}{e}\sqrt{\frac{2M_0}{\pi H_0 C}} \tag{6a, b}$$

In many applications a so-called horizontal diffusion coefficient K_h is used corresponding to a length scale l_h. These are commonly defined as follows

$$K_h = \sigma_r^2/4t, \qquad l_h = 3\sigma_r \tag{7a, b}$$

The results of this analysis are presented in Tables 3a and b.

In some cases the observed distributions are elongated and asymmetrical. Then it is more attractive to use the distributions directly for interpreting the mixing rather than transforming them into rotational symmetry. The longitudinal and transversal variances, σ_x^2 and σ_y^2, are calculated using the common definitions

$$\sigma_x^2 = \frac{\iint x^2 C\,dx\,dy}{\iint C\,dx\,dy}$$
$$\sigma_y^2 = \frac{\iint y^2 C\,dx\,dy}{\iint C\,dx\,dy} \quad (8a, b)$$

We still assume vertical homogeneity in the layer. A rectilinear coordinate system is used and adjusted so that the origin falls in the center of mass. The results are given in Tables 4 and 5b.

2.3. *Environmental data*

In the present work attention is only given to the stratification. As a measure of this the mean Brunt–Väisälä frequency squared is used in Table 2. This is determined by averaging all the vertical profiles of salinity and temperature obtained during the period of the experiment. The vertical resolution is 2.5–5 m.

3. Results

The experiments were conducted during relatively calm conditions. The dye was traced in layers of varying degree of stratification. The surface wave action was slight or absent, implying that the observed fine-structure of the vertical dye distributions was not an artifact of ship motion. The layered structure was clearly related to the temperature, salinity and current distributions in the water. An extensive discussion of the dye distribution in relation to the temperature dis-

Table 4. *Horizontal symmetrical variance σ_r^2 and corresponding diffusion velocity P, horizontal mixing coefficient K_h and scale l_h, longitudinal variance σ_x^2 and transversal variance σ_y^2, 9–10 May only*

Diffusion time (h)	σ_r^2 (m²)	P (m/h)	K_h (m²/h)	l_h (m)	σ_x^2 (m²)	σ_y^2 (m²)
5.8	$3.7 \cdot 10^3$	4.3	160	180	—	—
13.3	$1.6 \cdot 10^4$	3.8	290	380	$5.3 \cdot 10^4$	$3.1 \cdot 10^3$
18.5	$2.9 \cdot 10^4$	3.8	390	510	—	—
29.0	$6.1 \cdot 10^4$	3.5	530	740	$9.4 \cdot 10^4$	$1.6 \cdot 10^4$
40.0	$1.2 \cdot 10^5$	3.5	730	1000	—	—

Table 5a. *Previous results, vertical mixing*

Date	Mean depth (m)	Thickness (m)	Δt (h)	C_1/C_2	$K \cdot 10^6$ (m²/s)	Mean $K \cdot 10^6$ (m²/s)	$\overline{N^2}$ (s⁻²)	$\overline{\mathrm{Ri}}$
1969								
Dec. 9	25	0.60	0.4	1.8	16			
		0.40	0.9	6.2	9	12	$1.1 \cdot 10^{-3}$	0.4
1970								
Aug. 29	28	0.50	1.1	3.2	7.3			
		0.50	1.2	8	11.8	10	$4.8 \cdot 10^{-4}$	—
Aug. 30	50	1.00	0.9	1.1	3.9			
		0.50	3.5	2.5	1.8			
		0.75	4.0	1.6	1.8	2	$2.0 \cdot 10^{-3}$	2.0
Sept. 1	53	0.30	2.7	44	3.5	—	$2.7 \cdot 10^{-3}$	—
Sept. 2	26	0.25	4.0	23	1.5	—	$5.3 \cdot 10^{-4}$	0.9

Table 5b. *Previous results, horizontal mixing*

Date	Diffusion time (h)	σ_r^2 (m^2)	P (m/h)	K_h (m^2/h)	l_h (m)	σ_x^2 (m^2)	σ_y^2 (m^2)
Dec. 9	10.5	$4.3 \cdot 10^4$	7.2	1004	630	$3.7 \cdot 10^5$	$3.0 \cdot 10^3$
Aug. 30	7.5	$1.6 \cdot 10^4$	6.8	540	380	$2.3 \cdot 10^4$	$4.1 \cdot 10^3$
	16.7	$2.4 \cdot 10^4$	3.9	360	470	$1.4 \cdot 10^5$	$3.4 \cdot 10^3$
Sept. 2	6.7	$1.2 \cdot 10^4$	5.7	470	330	$1.5 \cdot 10^4$	$5.6 \cdot 10^3$

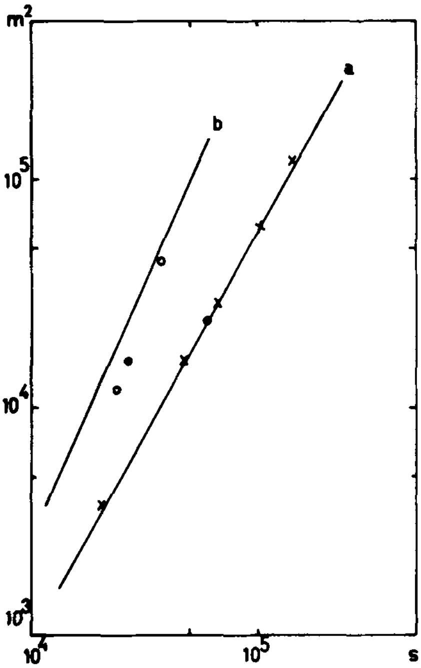

Fig. 2. Symmetrical variance σ_r^2 vs diffusion time t. × Baltic 75, ● previous deep, O previous shallow; line a: present results $\sigma_r^2 \propto t^{1.8}$; line b: $\sigma_r^2 \propto t^{2.3}$, from Okubo (1971).

tribution has been given by Kullenberg (1974a, b), and only a summary will be given here.

Several different types of dye layers are observed: pulse-formed with very sharp boundaries, layers with more ragged boundaries, layers with concentration inversions and leaf-structures with several sheets of dye separated from each other by water free from dye. This shows that there is fine-structure present in the water. In the deep experiments at 50–55 m depth the dye was found in the typical Baltic temperature minimum. Although the most common form was pulse-like, several cases of ragged boundaries and interleafing structure were observed (Fig. 1). These were found in very weak or no temperature gradient. Even after 40 h of tracing, well-defined, pulse-formed layers were detected. The thickest layers of the order of 2–3 m thick were invariably situated in the homogeneous parts of the temperature structure.

In the thermocline experiments around 25 m depth the dye was usually detected in relatively thin layers, often related to local features or disturbances in the temperature gradients. The thinnest layers, peaks, of the order 20 cm thick were found in regions of locally increased gradients. Most common were 15–30 cm thick layers apparently trapped in a local weakening of the basic temperature gradient. The difference between the deep and shallow experiments demonstrate the significance of the vertical current structure.

From these observations possible mixing mechanisms may be inferred. Using observations obtained with the present technique from several different areas, Kullenberg (1974a) concluded that the shear instability mechanism appears to be important, possibly in combination with breaking internal waves. In the Baltic it is likely that intrusions also play a significant role. The persistency of well-defined layers suggest that the mixing is characterized by intermittency. It is possible, and in agreement with the shear instability mechanism, that the mixing is completely determined by mixing events occurring on a relatively small scale. The form of the ragged profiles suggests that the mixing goes both upward and downward across the halocline. It is not an entrainment type of mixing.

The effective vertical mixing coefficient has been found earlier (Kullenberg, 1974a) to depend upon the wind, the stratification and the shear. In the

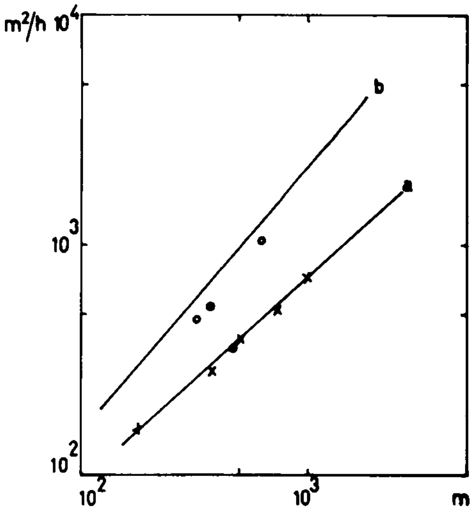

Fig. 3. Symmetrical horizontal diffusion coefficient K_h vs length scale l_h. Symbols as in Fig. 2; line a: present results $K_h \propto l_h^{0.9}$; line b: $K_h \propto l_h^{1.1}$, from Okubo (1971).

present experiments the wind velocities were weak and the mixing cannot be related to the local wind. It is clear that the stratification, as given by N^2, influences the mixing considerably. The data suggest that K is approximately inversely proportional to N, but much more data are needed to substantiate any such relationship. It is quite clear that the vertical mixing is very weak. The results from the different experiments conform, and the values are also in agreement with what is generally found in strongly stratified areas. The mean Richardson numbers given in Table 5a show that the layers were dynamically as well as statically stable. The mixing across the halocline can be expected to be stronger during the winter than during the summer. However, there is no indication of strong variations between different years. This of course requires that the forcing and transfer mechanisms are the same.

As regards the horizontal mixing, the mean diffusion velocities fall within a fairly narrow range, 3.3–7.2 m/h. These values appear very realistic in comparison with other reported results (Joseph et al., 1964; ICES, 1973).

For the surface layer the diffusion velocity is larger. The surface experiments during Baltic 75 give values in the range 9.7–21 m/h (Schott et al., 1976), i.e. a factor of about 5 larger than in the halocline layer.

The horizontal mixing is scale dependent. Therefore one cannot expect P to be constant over all scales. The relation between horizontal mixing and time is demonstrated in Fig. 2, giving σ_r^2 vs time. For comparison the regression line given by Okubo (1971) for the ocean surface layers is included. For the halocline layer in the Baltic the slope is less than in the ocean and the magnitude of the oceanic mixing becomes successively larger for increasing times. This demonstrates the scale effect. Talbot (1975) found that the slope varied considerably depending upon the area. For coastal waters the slope generally is less than 2, in agreement with the present results.

It is noted that the Joseph-Sendner model predicts $\sigma_r^2 \propto t^2$, which is quite close to what is observed in shelf seas. This suggests that the model gives a relatively good fit to the data. This was demonstrated more extensively by Schott et al. (1976) analyzing the surface experiments during Baltic 75. The present P-values obtained independently by means of (2, 4, 5, and 6a) also agree quite well. However, the prediction of the maximum radius found from (6b) does not agree with the observed values. The number of data is too low for drawing any firm conclusions, but it indicates the shortcomings of rotationally symmetrical models. Due to the combined action of vertical mixing and vertical shear, the so-called shear effect, the vertical mixing out of the layer has a considerable influence on the apparent horizontal mixing. Kullenberg (1972, 1974a) developed a time-dependent shear diffusion model and found that such a model was able to predict the apparent horizontal mixing fairly well, for scales up to about 1 week. It is quite clear that this mixing mechanism is very important in coastal waters and in those layers of the ocean as well, where there exists a vertical shear of some magnitude.

The horizontal diffusion coefficient K_h depends upon the length scale l_h (Fig. 3). For comparison the Okubo diagram (Okubo, 1971) has also been included here. In the Baltic the slope is less than in the oceanic surface layers, as is to be expected. The scale covered by the present experiments is limited, and one should be careful not to extrapolate the results too far. Boyce & Hamblin (1975) studied the spreading of a contaminant in the Great Lakes. They found that the effective horizontal coefficient giving a distribution in reasonable conformity with observations, was in good agreement with the value obtained by extrapolating the results given by Kullenberg et al. (1973) for Lake Ontario. These were obtained by the present technique, and the findings suggest that an extrapolation towards basin wide scales may yield useful estimates of the effective mixing.

4. Discussion

We will use the results to discuss some features of the overall mixing conditions in the Baltic.

First, consider the relative importance of the vertical mixing in the interior and in the coastal boundary zone. The significance of the latter has been demonstrated by Walin (1972a, b). The relative significance of the vertical transfer in these areas may be studied by expressing the vertical flux in the form

$$Q = K_z \cdot A \cdot t \frac{d\bar{C}}{dz}$$

where Q is the flux of the substance C and A the area. Schaffer (1975) is carrying out a long-term study of the coastal boundary zone. Taking averages over a 4-month observational period Schaffer (pers. comm.) found an average flux of salt of $2.3 \cdot 10^{-6}$ kg/m², s. For the open Baltic proper hydrographic observations suggest an average gradient of 2–3‰/10 m. The present data give an average of $\bar{K}_z = 2.2 \cdot 10^{-6}$ m²/s for the 50–60 m layer, giving a flux of $0.55 \cdot 10^{-6}$ kg/m², s, or a factor of 4.2 less than the average flux in the coastal boundary zone. We have assumed that the values of K_z found in the Bornholm basin are representative for the Baltic proper and also that they are representative for the whole year. The static stability across the halocline in the Gotland basin is often less than in the Bornholm basin which might imply slightly larger vertical mixing there. It should be noted that the interior mixing rates were obtained in calm conditions during spring and fall, using data from short period experiments. It is likely that the mixing is larger in strong wind conditions and during winter time. Hence the long-term average of the vertical mixing in the interior is probably slightly larger than the value used here.

The area of the Baltic proper including the Bornholm basin is 10^5 km² at the 60 m level (Ehlin et al., 1974), assuming this to be the average depth of the halocline layer. The area of the coastal boundary zone is estimated by assuming a width of 20 km with a length of approximately 1200 km around the Baltic proper at the 50–60 m level. Thus the ratio of the total fluxes is of the order of 1.0. The result suggests that, although the coastal boundary is more effective in generating vertical transfer, there is an approximate balance between the total fluxes in the interior and in the boundary zone. The interior flux cannot be disregarded relative to the boundary flux.

The next problem concerns the question: From where does the energy supply for the vertical mixing come? The tides in the Baltic are insignificant, and the weak estuarine driven flow does not generate any significant mixing. There remains the possibility of thermohaline convection and the energy input from the wind. The mixing generated by thermohaline convection is important for the water column above the halocline during part of the year. However, it does not penetrate beneath the halocline. Assuming that the heating/cooling essentially balance out during a year, let us consider the wind as the energy source. This is the energy source for the coastal boundary layer, and we now consider it as the driving force for the mixing in the interior as well.

Several recent studies have suggested that only a small fraction of the available wind energy is consumed for changing the potential energy of the water column (Denman, 1973; Denman & Miyake, 1973, Kullenberg, 1976). This is also in agreement with laboratory experiments (More & Long, 1971; Wu, 1973).

The amount of energy consumed per unit time and area from the surface to the depth $-H$ may be written as

$$P_E = \int_{-H}^{0} \rho_0 K N^2 dz \tag{10}$$

The total amount of turbulent energy available in the layer is then

$$T = P_E / Rf \tag{11}$$

where Rf is the flux Richardson number. The energy input per unit time and area from the wind can be given in the form

$$E = V_0 \tau_0 = k W_{10} \cdot \tau_0 = k c_d \rho_a W_{10}^3 \tag{12}$$

where k is the wind factor, V_0 the surface current, c_d the drag coefficient, ρ_a the air density and τ_0 the surface stress. Using the following values: $k = 2 \cdot 10^{-2}$, $\rho_a = 1.23 \cdot 10^{-3}$ ton/m³, $c_d = 1.3 \cdot 10^{-3}$, we find

$$E = 3.2 \cdot 10^{-8} W_{10}^3 \text{ ton/s}^3$$

The energy input to the waves is more difficult to express. It may be argued that it is at least equal to the input to the current (Phillips, 1966, p. 230). Thus the total energy input from the wind may be larger by a factor of 2–3 than that estimated by (12). The response of the Baltic to various winds is not considered here (see Walin, 1972; Kielmann et al., 1973). It is not unreasonable to expect that fluctuating winds can generate mixing.

An estimate of the range of the integral (10) over the depth interval 0–60 m may be obtained from the present data (Tables 2 and 5a). One finds the range $(6.35–2.44–0.41) \cdot 10^{-7}$ ton/s^3, using the maximum, intermediate and minimum mixing values. Several studies (e.g. Kullenberg, 1971, 1976; Stigebrant, 1976) suggest that a representative value of flux Richardson number over the present range of stabilities is $\overline{Rf} = 0.05$. Using this we find a mean value for the required total turbulent energy per unit time and area of $\bar{T} = 6 \cdot 10^{-6}$ ton/s^3. For a wind velocity of 5 m/s the input from the wind to the current is $E = 4 \cdot 10^{-6}$ ton/s^3. The yearly mean wind velocity over the SW Baltic is in the range 4–7 m/s. The total energy input from the wind is probably larger by at least a factor of 2, due to the wave energy. Thus we find that the required wind energy is available even though we have only considered the mixing to 60 m. The agreement between the estimates obtained by independent methods is in fact quite satisfactory. The result suggests that the wind is supplying at least part of the energy driving the weak but significant vertical mixing across the halocline.

It is of interest to proceed one step further and estimate the time Δt required to lower the salinity in the Baltic deep water after an inflow, by vertical mixing in the interior, to facilitate a successive renewal of the deep water. Long time observations (see e.g. Fonselius, 1969) suggest that the salinity of the volume beneath 120 m needs to be lowered approximately 2‰ between major inflows. The salt must be transferred across the halocline, a vertical distance of about 100 m on an average. The energy required per unit time and area is

$$E_1 = \frac{\Delta H \cdot g \cdot V_{120} \cdot \Delta S \cdot 10^{-3}}{2 \cdot \Delta t \cdot A_{120}} \text{ ton/s}^3$$

where ΔH *is the distance,* V_{120} and A_{120} the volume beneath and the area of the 120 m level, respectively, Δt is the time period and ΔS the salinity shift. The area and volume are found from Ehlin et al. (1974), and

$$E_1 = \frac{100 \cdot 10 \cdot 2 \cdot 10^3 \cdot 10^9 \cdot 2 \cdot 10^{-3}}{2 \cdot \Delta t \cdot 5 \cdot 10^4 \cdot 10^6} = \frac{40}{\Delta t} \text{ ton/s}^3$$

According to our observations the energy consumed for mixing is of the order of $3 \cdot 10^{-7}$ ton/s^3, which implies a time period between successive inflows of 4–5 years. This is in good agreement with observations (Fonselius, 1969; Dickson, 1971, 1973). Although this may be coincidental, it is quite clear that the results are of the right order of magnitude. Introducing the energy input from the wind as given by (12) and putting $E_1 = Rf \cdot E$, we find $\Delta t = 400\ \Delta S/W_{10}^3$ in years, with ΔS in ‰ and W_{10} in m/s; for $W_{10} = 6$ m/s and $\Delta S = 2$‰, $\Delta t = 3.7$ years.

The results suggest that the rate of renewal of the deep waters in the Baltic is sensitive to changes of the wind speed over the Baltic. Here we have considered the average wind speed, but the most effective mixing is undoubtedly generated during the passage of storms. Another measure of possible variations of the rate of renewal would therefore be the fluctuations in frequency and strength of storms over the Baltic. It should also be noted that the renewal of the Baltic deep waters depends critically upon conditions outside the Baltic (e.g. Wyrtki, 1954; Dickson, 1973). The frequency, strength and duration of westerly winds over northern Europe play a significant role (Soskin, 1963). However, the conditions in the Baltic are also important: the topography, the relative amount of river run-off, the stratification and stability of the halocline layer, and the winds.

The rate of renewal of the deep and bottom waters in the Baltic is very important for the oxygen conditions there. The rate of oxygen consumption is dependent upon several factors: the supply of organic material, the temperature of the water, the possible presence of H_2S, directly consuming part of the oxygen in the incoming water (Fonselius, 1969; Kullenberg, 1970). The coupling of these various factors is intricate. In addition the balance between aerobic and anoxic conditions appears to be very sensitive: small variations in consumption rates, renewal rates or supply rates may change the conditions. In this connection it is of interest to investigate the correlation between the variations of oxygen content of the Baltic deep water and the fluctuations of the winds. The yearly average of the

Fig. 4. Three-year gliding mean of third power of wind speed W_{10}^3 (x) from Gedser Rev observations and oxygen content O_2% in the 150–170 m layer at station F74 (O), from Fonselius (1969) during the present century.

wind strength observed at Gedser Rev light vessel is used as a measure of the prevailing winds over the southwest Baltic proper. Oxygen data are used as given by Fonselius (1969) at station F74 at 150–170 m depth. Yearly means are used. A fairly good correlation between variations of the oxygen content and variations of the wind can be seen in Fig. 4. The cube of the wind is given. The series have not been pursued beyond 1960 because about this time the measuring technique and the unit for the wind observations were changed.

The oxygen content is not a conservative property of the water, and it is more appropriate to use the salinity for investigating the mixing, as was done in the calculations above. It should be expected that the deep water salinity variations are influenced quite strongly by the wind fluctuations. This is demonstrated in Fig. 5a, using 3-year gliding means of the salinity at 100 m depth at station F74 and corresponding means of W_{10}^3 at Gedser Rev. The correlation is quite evident. In Fig. 5b the oxygen and salinity observations at 100 m depth from station F81 (Gotland Basin) are given vs W_{10}^3. Three-year gliding means are calculated from yearly averages of O_2 and $S‰$ obtained from the ICES data bank. The wind measuring technique was changed about 1960, and it has been necessary to apply a correction factor in order to adapt the later series of data to the earlier one. The correction factor was found from overlapping values. This procedure does not, however, influence the interpretation of the results. The correlation between the variables is quite evident. The salinity being a conservative property demonstrates the influence of the wind on the water exchange and the mixing. The oxygen content shows the same general behaviour, but it is noticeable that the increased salinity after about 1950 is accompanied by a decreased oxygen content.

Although the absolute oxygen content is determined by other factors (the oxygen content is not a conservative property), the fluctuations do seem to correlate fairly well with the fluctuations of the wind strength. This is in agreement with the present results and also with those of others stressing the role of the wind for the renewal of the deep waters (e.g. Wyrtki, 1954). It should again be pointed out that the required combination of pressure and winds over the North Sea is necessary for generating an inflow. However, the mixing in the

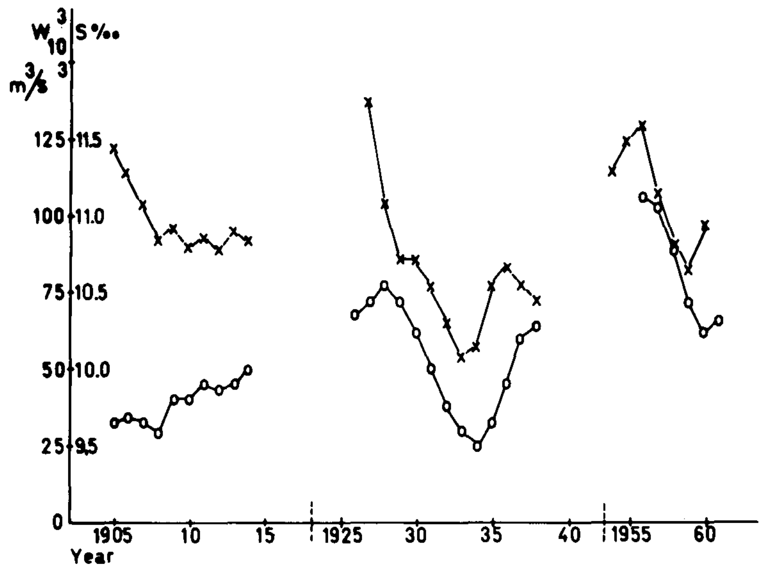

Fig. 5a. Three-year gliding mean of third power of wind speed W_{10}^3 from Gedser Rev observations (×), and salinity (O) at 100 m level at F74 (Fishery Board of Sweden Data) during the present century.

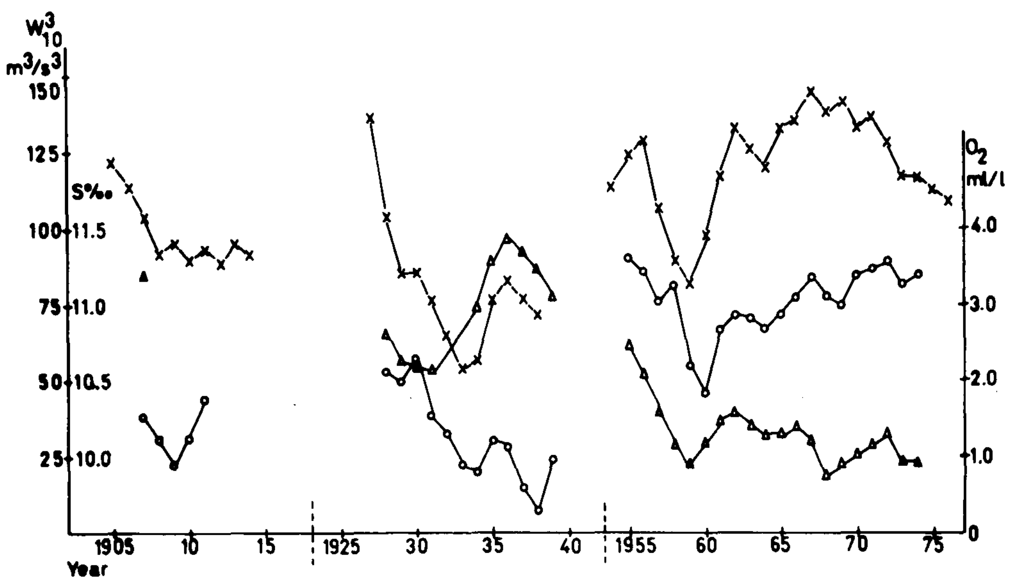

Fig. 5b. Three-year gliding mean of third power of wind speed W_{10}^3 from Gedser Rev observations (×), salinity S‰ (O) and oxygen O_2 ml/l (△), from 100 m depth at station F81 during the present century.

Baltic and the Transition area has a considerable influence on the depth to which the inflow can penetrate. The mixing in the Transition area is partly governed by the local winds, and the mixing in the Baltic seems to depend on the winds over the Baltic.

The oxygen content in the deep waters of the Baltic has decreased markedly since early 1900, whereas the winds do not show such a general trend. The frequency of inflows does not appear to change very much although the time span between inflows can vary (Dickson, 1973). The conclusion

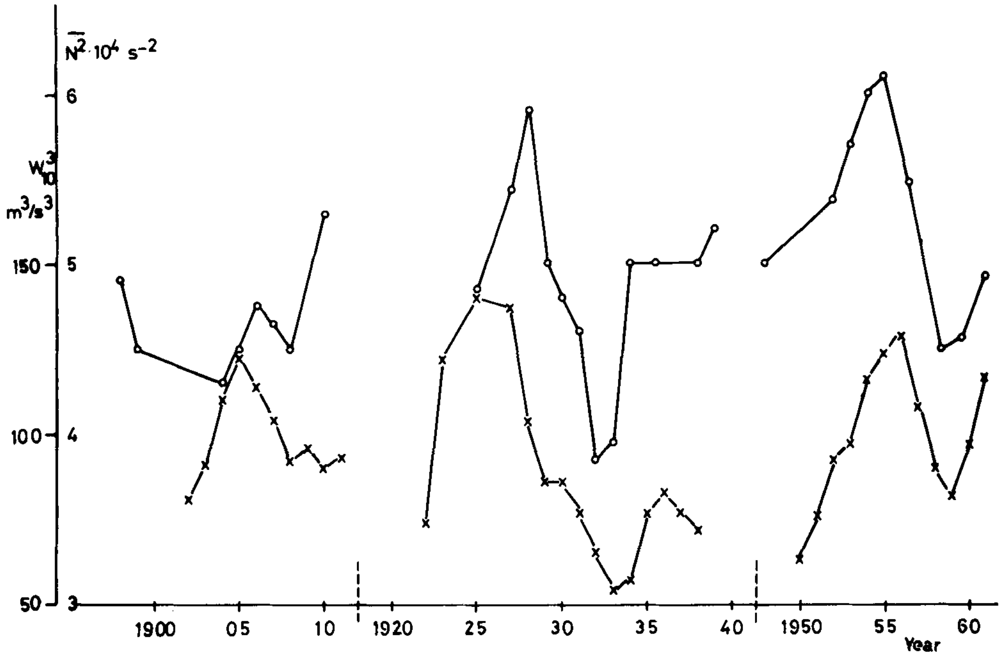

Fig. 6. Three-year gliding mean of third power of wind speed W_{10}^3 from Gedser Rev observations (×), vertical stability over the 50–100 m layer in the Gotland Basin as given by $N^2 = \frac{g}{\rho_o}\frac{d\rho}{dz}$ (O), from Fonselius (1969) during the present century.

is therefore that the decrease of the oxygen content in the deep waters is due to a decrease of oxygen in the incoming water and an increase of the oxygen consumption rate in the Baltic.

The stratification has also an influence on the vertical mixing. Fonselius (1969) found a slight increase of the average stability over the 50–100 m layer in the Landsort Deep during the period 1890–1960. In the Gotland Deep the fluctuations of the stability in the 50–100 m layer were found to be large but there was no general trend towards a significant increase of the stability. The conditions in the Bornholm Deep are similar. It is therefore not likely that the reduced oxygen content in the deep waters can be ascribed to a decreased vertical mixing across the halocline due to an increase of the stability.

The importance of the wind for the lateral transport of deep water can be demonstrated by correlating the fluctuations of the stability in the Gotland Deep, obtained from Fonselius (1969), with the wind fluctuations (Fig. 6). It is noted that low values of the stability correlate with weak winds whereas high stabilities correlate with strong winds. The wind is essential in driving the lateral inflows of water from outside the Baltic sills. The wind is also important for supplying energy for driving the vertical mixing in the Baltic.

The wind direction has not been considered here. Inflows depend critically upon the direction as well as the strength of the wind (e.g. Wyrtki, 1954). The frequency and duration of strong westerly winds have fluctuated during the present century, as pointed out by Soskin (1963). Although this can influence the frequency and strength of inflows, the analysis of Dickson (1971, 1973) suggests that the effect has not been very marked since high saline water penetrates to the Baltic entrance with a surprising regularity. The present results indicate that the water exchange between the Baltic and the outside ocean has not changed character during the present century.

5. Conclusions

The mixing in Baltic thermo- and halocline layers is considerably weaker than in the open ocean surface layer. Although the vertical mixing in the open Baltic across the halocline layer is very weak it plays a significant role for the overall conditions. The energy transferred from the wind to the water is essential for the mixing conditions. The correlation between the wind energy, the salinity and the oxygen content in the deep waters shows the importance of the wind forcing both for the mixing and the renewal by lateral inflows. The results suggest that the oxygen decrease in the deep waters during this century cannot be ascribed to variations of the water exchange pattern.

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НАБЛЮДЕНИЯ ПЕРЕМЕШИВАНИЯ В БАЛТИЙСКИХ ТЕРМО- И ГАЛОКЛИННЫХ СЛОЯХ

В течение весны и осени в серии экспериментов исследовалось горизонтальное и вертикальное перемешивание пассивной примеси в термо- и галоклинном слое Балтийского моря. Вертикальное перемешивание очень слабое. Горизонтальное растекание в 2–5 раз меньше, чем в поверхностном слое. Интегральный вертикальный перенос через слой галоклина того же порядка величины, как интегральный перенос в прибрежных пограничных зонах. Найдено, что существенным источником энергии для перемешивания является ветер. Значение ветра демонстрируется корреляцией между флуктуациями ветра и (1) флуктуациями содержания кислорода в глубинных водах, а также (2) флуктуациями устойчивости в слое галоклина, измеренными в настоящем столетии. Эти результаты свидетельствуют в пользу точки зрения, что общее уменьшение содержания кислорлда в глубинных водах вызываются какими-то другими факторами, а не изменением характеристик водообмена.