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Oxygen Depletion in the Kattegat

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A mathematical model driven by meteorological forcing and sea level differences has been established in order to describe the water exchange and vertical mixing in the Kattegat and Samsø Belt.

The model has been run with a time-step of 1 day for the period April – October from 1961 to 1981. The mixing has been represented by equations from the theory of turbulence with the normally used coefficients, i.e. no calibration has been necessary. Results from earlier investigations have been used to describe the flows in the Belts and the Sound.

The vertical exchange between the upper, brackish and the lower, saline water is strongly dependent on the meteorology. A small meteorological activity results in small exchange flows which implies that the oxygen supply to the lower layer becomes smaller and unfavourable biological conditions may occur. A simple calculation has been carried out on the assumption that the chemical and biological oxygen demand is a first order process with a time scale for the oxygen decay, which is halfed for each 10° increase of temperature. Thus the model describes the effect of the *meteorological* activity on the oxygen concentration, all other parameters kept constant.

Description of Data

The Baltic Transition Area is one of the areas in the World for which the longest hydrographical time series exist. This permits the development of a model for description of changes which take place on a climatic time scale. The requirements of the model must, however, comply with the type of data available, the sampling

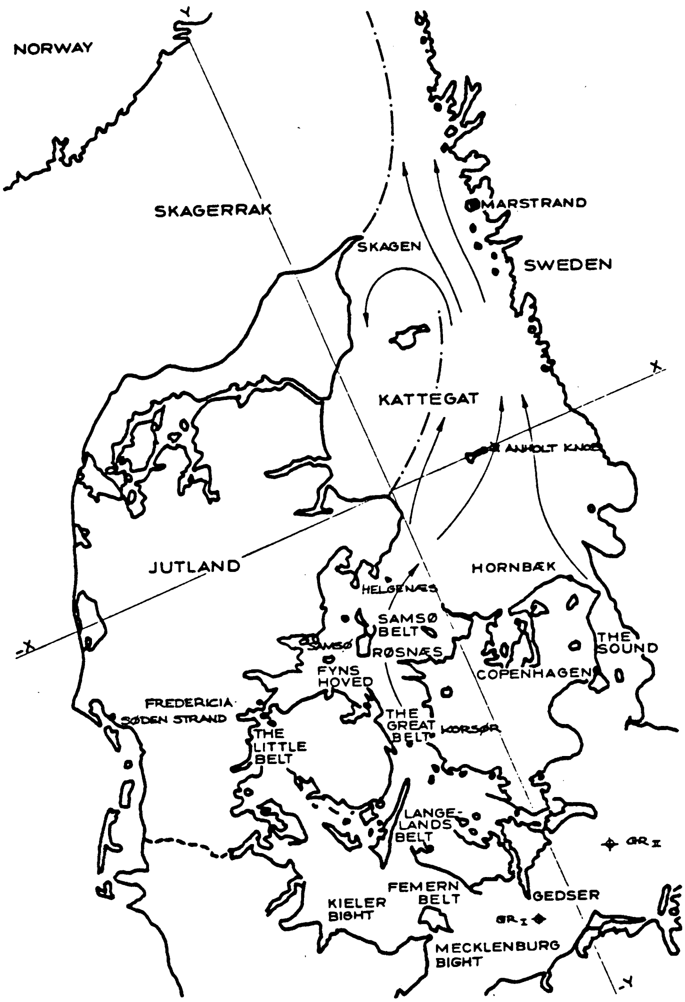


Fig. 1. Map of the inner Danish Waters.

interval, the station network and the existence of possible gaps in the series. Most of the data required for the present model are available for about 100 years.

Sea level has been measured at 10 locations with recording tide gauges since about 1890, the purpose being to establish a reference level for the Geodetic Survey. These stations are Esbjerg, Hirtshals, Frederikshavn, Århus, Fredericia, Slipshavn, Korsør, Hornbæk, Copenhagen and Gedser. A few more stations have been installed later. Hornbæk and Gedser have been selected to represent the sea level change along the Belt Sea (Fig. 1) and Fredericia has replaced Hornbæk when the latter was missing. Fredericia and Gedser had a digital recorder installed in 1972, Hornbæk in 1973. Prior to that time the data were read from analog (ink)

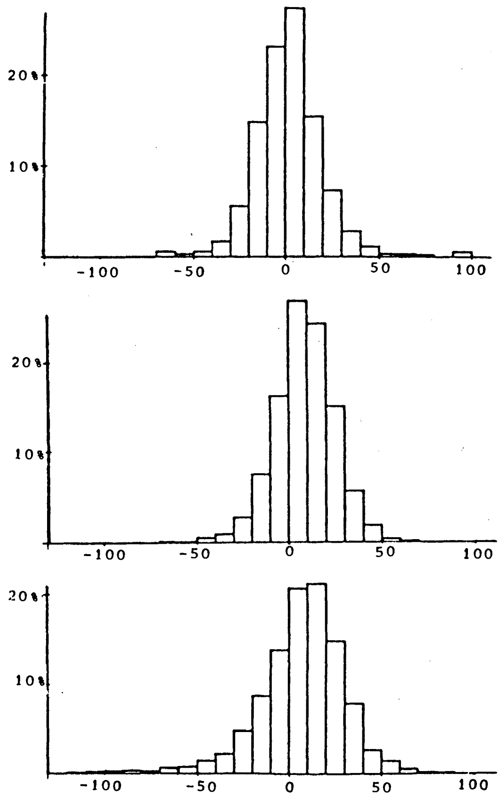


Fig. 2.
Distribution of daily averages of sea water levels, 1/4-15/10, for the years 1960-1981.

a) Hornbæk,
b) Gedser,
c) difference Hornbæk-Gedser.

recordings at hourly intervals. Short gaps in the series are caused by malfunction of the recorder, while longer gaps are due to the fact that the latest analog recordings have not yet been digitized. This is why the model was not run for 1971 and 1972.

The sea levels used for the model are daily averages related to DNN (Danish Normal Zero). DNN was an equipotential surface when established, but different rates of isostatic land-rise have since then distorted it. The present estimate is that the DNN zero at Hornbæk is positioned 1.5 cm above the zero at Gedser within a mean error of 6 cm (Bedsted Andersen and Kejlsøe, private comm. 1977). This small and uncertain zero offset was not taken into account. The distribution of daily averages for Hornbæk and Gedser and for their difference Gedser-Hornbæk are shown in Fig. 2. It is recognized that the most frequent sea level is in the range 0-10 cm.

The wind supplies energy to the upper layer and creates turbulence which causes erosion (entrainment) of the bottom layer. The wind-induced sea level set-up which governs the barotropic water exchange through the Belt Sea is another effect, but it doesn't enter the model through the wind, as the sea level difference is used directly as a forcing function. The representation of the wind presents a problem.

First, the energy input to the sea is proportional to the cube of the wind, so a small error in observing the wind will be increased relatively appr. 3-fold in the calculated energy. Second, the wind near the ground has a considerable spatial variation because of varying topography, and wind observations should be pooled from a number of coastal stations in order to create a reliable estimate of the wind at sea. A single sea station, Anholt Knob, does however, exist. Furthermore, wind observations were formerly carried out without use of instruments and are therefore likely to contain 'subjective' noise. This may bias the long-term changes to be analyzed.

Another method is to calculate the geostrophic wind and determine how it may best be reduced to represent the wind at sea level. This method offers the advantage that the air pressure has always been objectively determined and a long term bias is less likely to be present. On the other hand, instrumental zero-offsets create non-existing winds, erroneous readings may create very high wind speeds, and since the stratification in the near-surface atmosphere is not known, it will be necessary to introduce a constant reduction factor for the geostrophic wind. However, as water temperature is almost always lower than air temperature from April to October, a stable stratification is likely to persist.

The decision was in favour of the objectivity of the geostrophic method, and a large triangle, Skagen – Sæden Strand – Christiansø, was selected. A fourth air pressure station, Jønkøping, was kept as a reserve should one of the three corners be missing. The geostrophic wind was calculated from the daily averages of the synoptic pressure observations.

The size of the triangle introduces the risk of losing high speed local winds which are typically associated with pressure patterns of small dimensions, but since years and not single days are to be compared, it is assumed that this irregularity is statistically smoothed out. A comparison with the observed wind at Anholt Knob Lightship (Fig. 3) has shown that a fairly good fit to the observed wind is obtained by reducing the geostrophic wind to 63% and turning it 13 degrees towards low pressure.

Air temperature for the Kattegat is represented by Anholt Knob Lightship and for the Belt Sea by Gedser Rev.

Sea surface temperature has been represented as air temperature (above). It has been observed once a day (0800).

Air pressure, air temperature and sea surface temperature are available from the establishment of the network of lightships, i.e. about 100 years.

The incident short-wave radiation (300-2,500 nm) is complete for the entire period 1961-81. Data are available back to 1955 from Landbohøjskolen. It has been measured at Landbohøjskolen, Copenhagen, and it represents the sea after being corrected for the difference in albedo.

Final Remarks. A complete list over the availability of data in the period 1961-81 is depicted in Fig. 4.

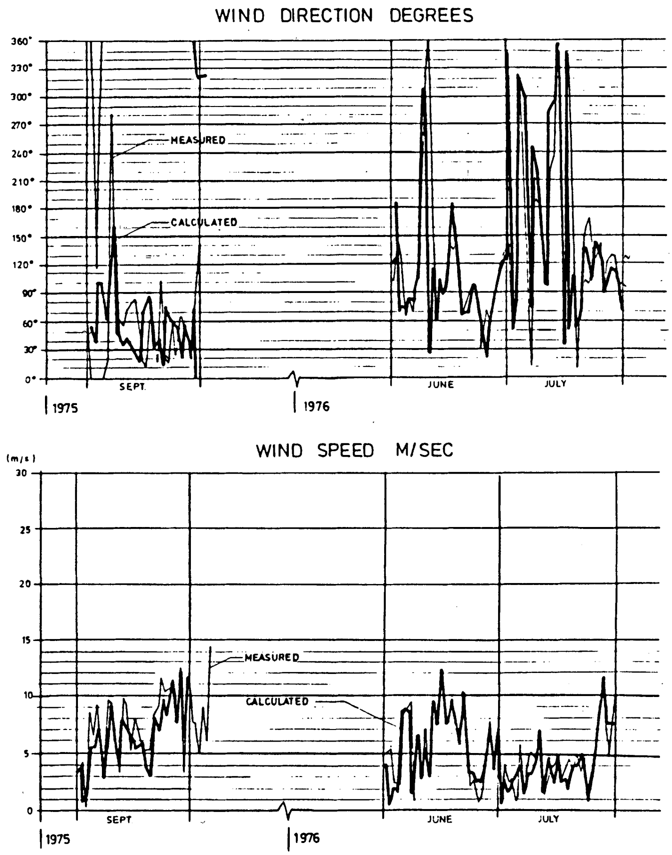


Fig. 3. Comparison of winds measured at Anholt Knob lightship with winds derived from geostrofic winds.

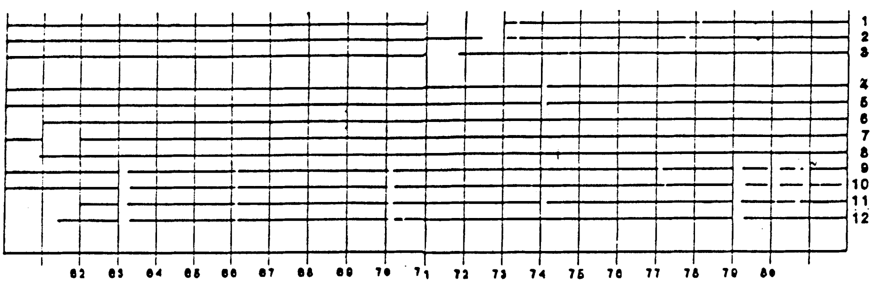


Fig. 4. Available data in the period 1961-81 for the parameters 1) Sea level Hornbæk 2) Sea level Fredericia 3) Sea level Gedser 4) Air pressure Skagen 5) Air pressure Sæden Strand 6) Air pressure Christians Ø 7) Air pressure Jønkøping 8) Short Wave Radiation 9) Sea surface temperature Anholt Knob 10) Sea surface temperature Gedser Rev 11) Air temperature Anholt Knob 12) Air temperature Gedser Rev.

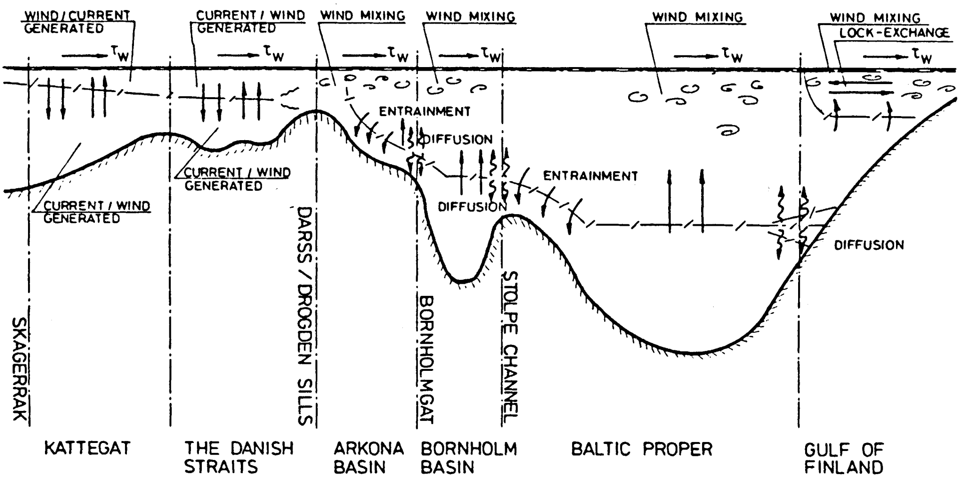


Fig. 5. The Kattegat and the Baltic, indicating mixing processes.

Mathematical Description

Introduction

This model describes the currents in the Kattegat during the summer half where the flow takes place under stratified conditions. The wind and Coriolis forces change the interface position, resulting in large shift of water between the various parts of the Kattegat. However, the wind is not sufficiently strong for this during the summer half. The summer half is to be defined as the time from the last winter storm to the first storm in the autumn. The model can, at present, not handle strong winds, especially if they are from the north. However, as the oxygen depletion takes place during late summer, this is not of importance for the present application.

If a frontal zone develops in the Kattegat the lower layer comes to the surface and oxygen can enter it without restrictions. This is one of the reasons why the oxygen conditions are favourable during winter. In the summer half on the other hand, the upper layer forms a barrier to downward oxygen transport. The only oxygen, which reaches the lower layer during this period, originates from the Skagerak (by horizontal advection) or from diffusion and downwards entrainment through the halocline (caused by currents in the lower layer). During periods of calm, hot weather a termocline develops near the surface and prevents windgenerated mixing. A longitudinal section of the Kattegat, Belt Sea and Baltic is shown in Fig. 5.

In the following, the equations for motion, mass, volume, continuity and the boundary conditions are discussed.

The Northern Boundary Condition

The depth of the upper layer in the Kattegat is dependent on the choice of boundary condition imposed for the outflow of brackish water from the Kattegat. Observations indicate, that the condition should be that of a normal critical cross section known from channel flow i.e. a section with the densimetric Froude number close to unity (Ljøen 1976).

The assumption is supported by observations which show the Coriolis force to develop an eastern current along the Swedish coast, while the currents in the western Kattegat along the eastcoast of Jutland appears to be a kind of large dead water zone but with superposed currents generated by the interface movements, Fig. 1.

The current, once having left the northernmost point of Skagen, is governed by a purely frontal condition along the western boundary and a considerable contraction begins. It converges towards a critical cross-section halfway up in the Skagerak. After having passed this section the water enters the Norwegian Coastal Current, which is governed by the wind or by critical cross-sections along the Norwegian coast (Ljøen 1976, Schulz and Thorade 1936).

The equation of motion at the critical cross-section is at right angle to the current

$$\rho_u \, y \, f_c \, U + \tau_{XW} = \Delta \rho_u \, g y \frac{dy}{dx} \tag{1}$$

where ϱ_u is the density of the upper layer, y the upper layer depth (a function of x), f_c the Coriolis parameter and U the current velocity which is assumed to be uniformly distributed over the critical section. τ_{XW} is the wind shear stress at a right angle to the current, g the acceleration of gravity, and $\triangle$ the relative density difference between the two layers. Inertial terms and pressure contributions caused by density variations across the flow have been neglected because they are small. Further the interfacial friction has been neglected.

The linearized solution for the equation is

$$y = \left| \frac{f_c U}{g\Delta} + \frac{\tau_{XW}}{\rho \Delta g \bar{y}_c} \right| B_s \, \frac{x}{B_s} \tag{2}$$

B_s being the width and $\bar{y}_c$ the mean depth of the critical section. The mean depth $\bar{y}_c$ is calculated from Eq. (2).

$$\bar{y}_c = \left| \frac{1}{2} \frac{f_c U}{g\Delta} + \frac{\tau X_W}{\rho \Delta g \bar{y}_c} \right| B_s \tag{3}$$

Critical flow at the cross section is expressed by the densimetric Froude number, $\mathbf{F}_\triangle$, being

$$\mathbb{F}_{\Delta}^{2} = \frac{U^2}{g\bar{y}_c\Delta} = 1 \qquad (4)$$

By substituting Eq. (4) into Eq. (3), the width of the critical cross-section is determined at

$$B_s = \frac{2U^2}{\left| f_c U + \frac{\tau_{XW}}{\rho_u U^2} \Delta g \right|} \qquad (5)$$

Eq. (5) and the expression for the water transport, $Q = UB_s\bar{y}_c$, give the critical condition

$$\bar{y}_c = \sqrt{\frac{1}{2}\left(\frac{f_c Q}{g\Delta} + \frac{\tau_{XW}}{\rho\Delta g}\sqrt{\frac{Q^2}{g\Delta\bar{y}_c^3}}\right)} \qquad (6)$$

Next applying the energy equation between the critical section and the section Skagen-Marstrand, the depth, y_m, in the latter section can be calculated to

$$y_m \equiv \frac{3}{2}\bar{y}_c \frac{1}{1+\frac{1}{2}\mathbb{F}_{\Delta}{}^{2}{}_{sm}} \qquad (7)$$

where $\mathbf{F}_{\Delta sm}$ is a mean densimetric Froude number for the Skagen-Marstrand section. It is emphasized here that the velocity in this section is not uniformly distributed, Fig. 1.

Since the densimetric Froude number is usually small, south of the section Skagen-Marstrand, Eq. (7) is approximated by

$$y_m = \frac{3}{2}\bar{y}_c \qquad (8)$$

and this gives the critical condition

$$y_m = \frac{3}{2}\sqrt{\frac{1}{2}\frac{f_c Q}{g\Delta} + \frac{\tau_{XW}}{\rho\Delta g}\sqrt{\frac{Q^2}{g\Delta(\frac{2}{3}y_m)^3}}} \qquad (9)$$

This is only valid as long as $B_s<B_o$, B_o being the distance Skagen-Marstrand. If B_s becomes greater than B_o, we have a normal critical condition across B_o

$$y_m = \frac{3}{2}\sqrt[3]{\frac{Q^2}{g\Delta B_0^2}}, \quad B_0 < B_s \qquad (10)$$

Eqs. (9) or (10) makes up the northern boundary condition. They determine the relationship between the outflow and the upper layer depth.

The boundary condition is assumed valid for the entire summer half, i.e. it is assumed that outflow takes place all the time.

The boundary condition will generally not be valid in the winter when the strong westerly, northerly or southerly winds will create frontal zones in the Kattegat.

The Vertical Exchange

Mechanisms for Mixing

The exchange of water between the upper and lower layer is caused by three different mechanisms:

- Wind generated upwards entrainment
- Flow generated upwards entrainment
- Flow generated downwards entrainment

The windgenerated upwards entrainment starts with an erosion of a thermocline if present and when it becomes eroded down to the halocline, the upwards entrainment of saline water from the lower to the upper brackish layer will take place.

The flow generated entrainments are caused by the currents generated by the in- or outflow to the Baltic, and by the tides. The entrainment is calculated according to the theory presented by Pedersen (1980), with only minor changes of coefficients in the formulae for the present application, see Eq. (24). The entrainment is thus only a function of the external energy supply, i.e. the wind or the current.

Wind Generated upwards Entrainment

The wind generated upwards entrainment velocity, v_E, is calculated from the following expression

$$\frac{v_E}{U_F} \equiv \frac{2.3}{6+\mathbb{R}_{iF}} \tag{11}$$

where

$$\mathbb{R}_{iF} = \frac{\Delta g y}{U_f^2} \tag{12}$$

$$U_F^2 = \frac{f_w}{2}\ \frac{\rho_a}{\rho}\, W^2 \tag{13}$$

in which w is the wind velocity (m/s) and ϱ_a/ϱ the ratio between the densities of air and sea water ($=1.3\times10^{-3}$). f_w is the friction factor which effectively is equal to 3.0×10^{-3} for Kattegat, and y is the upper layer depth (m). Normally the friction factor will be equal to 3.8×10^{-3}, but since the model calculates in time steps of 1 day the high-frequency occurrences are filtered out. This results in an effective reduction in the wind frictional coefficient.

Eqs. (11)-(13) can be reduced to

$$v_E = \left(\frac{f_W}{2}\right)^{\frac{3}{2}} \frac{1.08\times10^{-4}\,W^3}{\Delta g y + 7.8\times10^{-3}\,\frac{f_W}{2}\,W^2} \quad \text{(m/s)} \tag{14}$$

The Heat exchange to the Surface

The wind generated upwards entrainment of saline water from the lower layer is prevented if a thermocline is present. The thermocline is formed as a result of a net energy supply to the surface (heat). The surface exchange of energy consists of four parts:

1) R_{nK} – short wave radiation
2) R_{nL} – net long wave radiation
3) R_h – convective transfer of energy
4) R_e – energy exchange by latent heat (evaporation, condensation)

R_{nK} is a supply of energy *to* the surface, R_{nL} is a loss *from* the surface. R_e and R_h usually are of the same order and of opposite signs.

1) R_{nK} is calculated from the semi-empirical expression

$$R_{nK} = I_0(1-r)a_{nK} \tag{15}$$

I_0 is the insolation (W/m^2), r the albedo and a_{nK} the transmission factor. a_{nK} is determined empirically as a function of the number of hours with sun, n, and the hours with cloudiness

$$a_{nK} = (0.18 + 0.55\frac{n}{N}) \tag{16}$$

N is the number of hours with daylight. r is at sea a function of the solar elevation, and at land dependent on the type of surface, but independent of the solar elevation.

In the mathematical model the R_{nK} measured at Copenhagen have been used. These measurements may with reasonable accuracy represent the Kattegat and the Belt Sea after being corrected for the difference in albedo. The cloud cover may be denser at land than at sea as an average, but this error is so small that it is accepted.

2) R_{nL} has contributions from the outgoing radiation from the sea surface and from the reflected energy from the atmosphere. The latter depends on the degree of cloudiness and the humidity. A bulk expression for these effects are (Gottlieb 1978)

$$R_{nL} = -0.9\sigma T_s^4 + (0.75\frac{n}{N} + 1 - \frac{n}{N})\sigma T_a^4 = -0.9\sigma T_s^4 + \sigma T_a^4(1-0.25\frac{n}{N}) \tag{17}$$

the water surface temperature T_s, and the air temperature T_a, are known, but n/N (the fraction of sun hours relative to the maximum possible hours) is usually not known. We obtain then

$$\frac{n}{N} = 1.8\frac{R_{nK},\ \text{measured}}{I_0(1-r)} - 0.18 \quad \text{for } 0 \le \frac{n}{N} \le 1 \tag{18}$$

from which R_{nL} is calculated.

3) R_h is calculated from the relation

$$R_h = \rho_a C_a (T_a - T_s) C_h W \qquad (19)$$

where

ϱ_a – the density of the air (kg/m^3)
c_a – the specific heat capacity of air (J/kg/°C)
T_{luft} – the air temperature (°K)
T_s – the sea surface temperature (°K)
C_h – an empirical constant (= 0.001886)
w – the wind velocity (m/s)

4) R_e is determined from

$$R_e = L\rho_a \frac{0{,}633}{p}(e_a - e_{s,m}) C_e W \qquad (20)$$

where

L – the latent heat of evaporation (J/kg)
ϱ_a – the density of the air (kg/m^3)
P – the air pressure
e_a – the water vapour pressure
$e_{s,m}$ – the pressure of saturated water vapour at the sea surface
C_e – an empirical constant ($=C_h$)
w – the wind velocity

All quantities except e_a and $e_{s,m}$ have been measured. $E_{s,m}$ is a known function of the sea surface temperature, and e_a can approximately be calculated as a function of T_a when the humidity is assumed constant. The humidity varies between 70 and 100% which implies that the variation of R_e is within 30%. The possible error introduced by this assumption is not assumed to be higher than the error of extending land based measurements of evaporation to the entire Kattegat.

The calculations of R_e and R_h assumes similarity theory to be valid in the boundary layer, which can only be done when the atmosphere is stable. Since the water temperature in the summer period almost always will be lower than that of the air in this region (the Kattegat), this assumption is valid.

After having calculated the net supply of energy of the surface, the change of temperature and volume is found from

$$\delta T = \frac{\delta E}{C_W \rho y} \qquad (21)$$

The temperature change produces a density change which is

$$\delta\rho = \frac{\delta E\alpha}{C_W \rho y} \tag{22}$$

α being the coefficient of thermal expansion.

From the above equations the time necessary to erode the newly established thermocline down to the salinity interface is calculated. If more than 1 day is used, there will be no wind generated entrainment across the halocline. If the erosion of the thermocline takes place faster than 1 day, entrainment is assumed to take place for the rest of the day. Calculation starts every day by forming a new thermocline at the surface, and possible remaining thermoclines from the preceding days are kept in the system.

Flow Generated Upwards and Downwards Entrainment

The flow generated entrainments are calculated from

$$\mathbb{R}_F^T = \frac{\frac{1}{2} v_E \, \xi \Delta g y \rho + \text{GAIN OF TURB.KIN.ENERGY}}{\text{PROD}} \tag{23}$$

v_E – the entrainment velocity (m/s)
ξ – a distribution parameter (=1)
y – the layer depth (m)
$\mathbf{R}_F^T$ – the Bulk Flux Richardson Number (=0.04-0.05 for subcritical flow) and = 0.18 for supercritical flow. In this context is used $\bar{\mathbf{R}}_F^T = 0.040$
PROD – the turbulent kinetic energy produced per unit area in the water

For $F_\Delta^2 \rightarrow 0$ the gain in turbulent kinetic energy becomes negligible, while for $F_\Delta^2 \rightarrow \infty$ $v_E/V \rightarrow 0.07$. We therefore use

$$v_E = \frac{\text{PROD} \times \mathbb{R}_T^F}{\frac{1}{2}\Delta g y \rho + \frac{\text{PROD} \times \mathbb{R}_f^T}{V\, 0.07}} \tag{24}$$

The production of turbulent energy PROD is calculated as follows

$$\text{PROD}_{\text{lower}} = V_\ell \tau_b + V_i \tau_i = \rho \frac{f_b}{2} V_\ell^3 + \rho \frac{f_i}{2} V_\ell (V_\ell - v_i)^2 \tag{25}$$

$$\text{PROD}_{\text{upper}} = V_u \tau_i = V_u (V_u - v_i)^2 \rho \frac{f_i}{2} \tag{26}$$

in which V_u and V_ℓ is the velocities in the upper and lower layer respectively. V_i is the interfacial velocity $(=\frac{1}{2}(V_u - V_\ell))$. The tidal current are included in the above velocities and has been taken as a simple harmonic current with an amplitude of 0.12 m/s and a period of ½ day for the Kattegat region.

The Equations of Motion for the Kattegat

The equations of motion have been applied with the following assumptions:

- convective terms can be neglected
- horisontal density effects in the pressure terms are neglected
- the Boussinesq assumption is valid
- They have been expressed as integral equations for the northern and the southern parts of the Kattegat.

In order to solve these equations, a number of simplifications are necessary. Most important is the assumptions of the form of the surface and the interface. For this model it has been assumed that they are hyperboloids (surfaces with straight line generatrices). The solution of the equations are described in a later chapter.

The Southern Boundary Condition

The southern boundary conditons is the following:

- The in- and outflow through the Great Belt is found from a relation which was earlier established (Pedersen 1980).
  The water exchange in the upper layer, Q_{SB}, was expressed simply as

$$\Delta H = K\,Q_{SB}^{\;2} \qquad (27)$$

  $\triangle H$ is the sea level difference between Gedser and Hornbæk and K is the specific resistance of the Great Belt and Fehmarn Belt. The specific resistance was found to depend on the interface position (Pedersen 1980).
- The model boundary condition is placed in the southern Langelands Belt, the salinity of the upper layer assumed to be 11 ppt.
- The model boundary condition for the lower layer is placed at the section Røsnæs-Fyns Hoved, and has two components:
  a) The total water exchange generated by wind- and current mixing in the Great Belt, Kiel Bight and Mecklenburg Bight.
  b) An extra contribution at high velocities from the Kattegat to the Great Belt, because the velocity of the lower layer on such occasions is increased to 1/3-1/2 of the velocity of the upper layer. This has been demonstrated by the Belt Project measurements at that section.
  c) A horizontal exchange (dispersion) of 15% within each layer.
  With these assumptions the flow of the lower layer will almost always go south, except if large changes of the interface takes place. This is in agreement with current observations close to the bottom (The Swedish Gøteborg-Frederikshavn Project).

– The flow through the Sound and the Little Belt are put at 43% and 14% of the Great Belt exchange, respectively.

The calculated flows will not necessarily satisfy the equation of continuity of the Baltic Sea, but this is of no consequence for the Kattegat, since the water balance at this location is determined from a state of equilibrium between the outflow at the northern control section and the vertical mixing and the net outflow from the Sound and the Belts.

Solution of the Equations

In order to obtain a stable solution when using a time step of one day, it becomes necessary to utilize a rather specialized algoritm. It is carried out as follows:

1) The stationary solution to the equations of motion is determined for the time step in question. This is the solution which would apply, if conditions remained unchanged.
2) Since the system due to the changes in meteorology undergoes a change all the time from one intermediate state to a new one without ever reaching the steady state condition, it is necessary to determine the proper response functions for the system. This is done approximately by assuming that the response on an impulse is an exponential function with the time scale determined by the velocity of the internal waves and reservoir time constant for Kattegat (e.g. the time to tilt and move the interface). The change from one situation to a new one is carried out by means of a convolution integral technique.
3) After these steps the new salinity distribution is calculated from the mass continuity equation.

The solutions are finally transformed to a number of areas of the Kattegat, as shown in Fig. 1. If the time scale for the tilt of the interface is long, the interface will almost remain horizontal. On the other hand, if the time scale is small the interface tilting will be quick which results in very large volume exchanges in the coastal zone.

Verification

To verify the model comparisons with existing measurements in nature have been performed. The checks have been performed for salinity and interface depth for certain periods. The comparisons are shown in Figs. 6 and 7.

It is realised that high frequency motions and local pertubations are not covered by the model. However, the longer term effects (effects with time scales longer than 3 days) are fairly well described (Figs. 6 and 7).

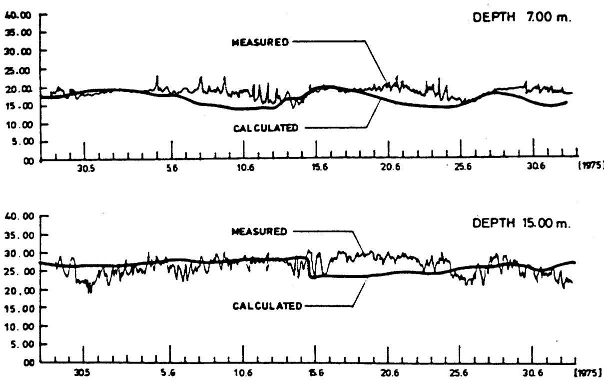


Fig. 6.
Comparison between measured and calculated salinities in the Northern Great Belt (National Agency of Environmental Protection 1979).

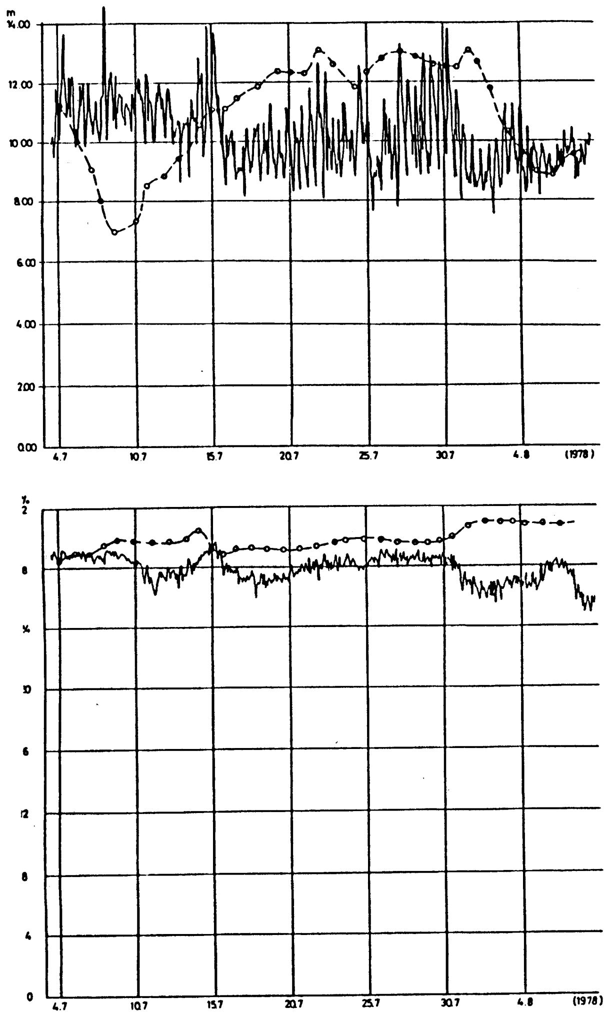
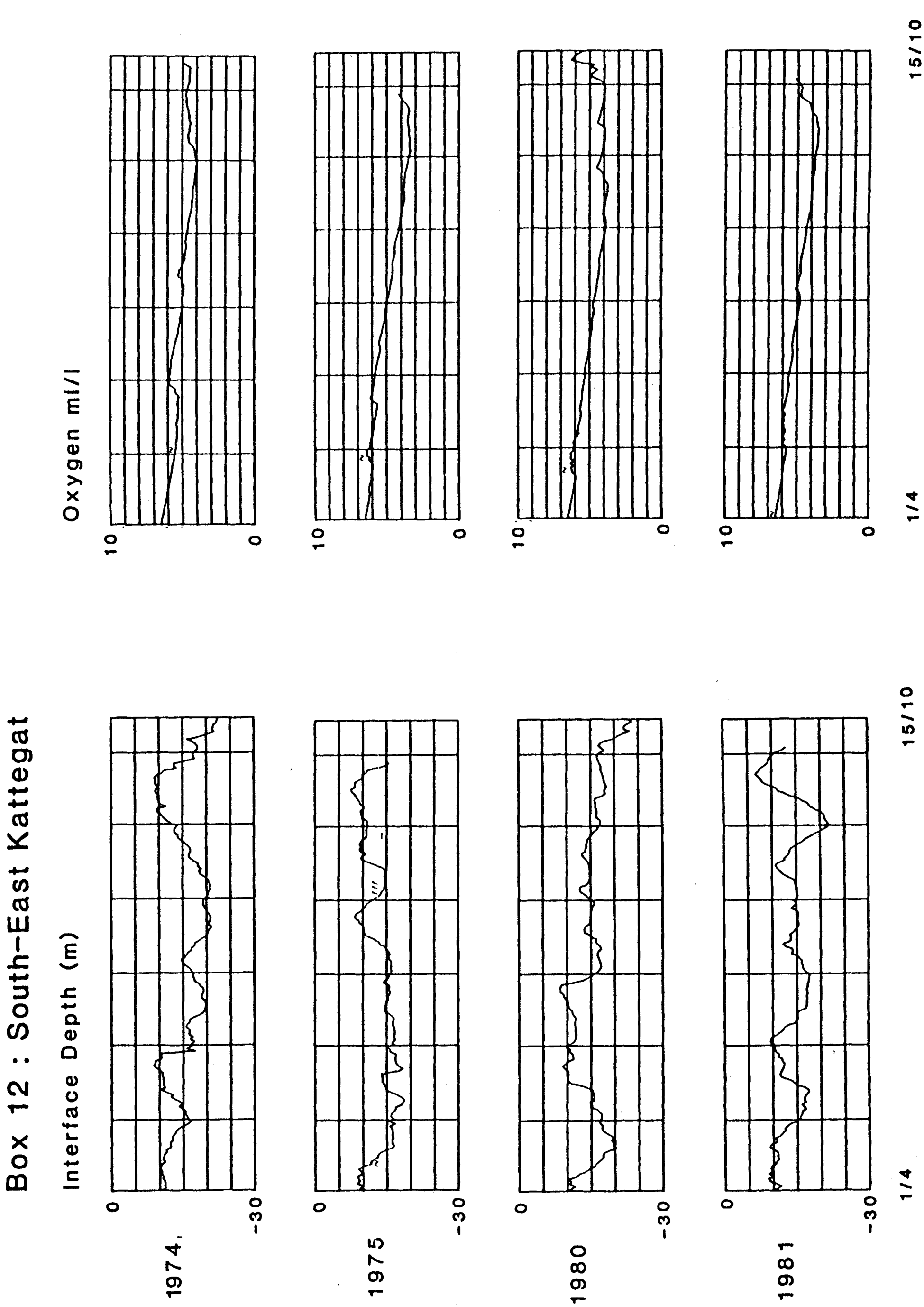


Fig. 7.
Comparison between measured and calculated salinities and interface depth between Samsø and Helgenæs (Danish Hydraulic Institute 1980).


Box 12 : South-East Kattegat
Interface Depth (m)
Oxygen ml/l
1974
1975
1980
1981
0
-30
10
0
1/4
15/10

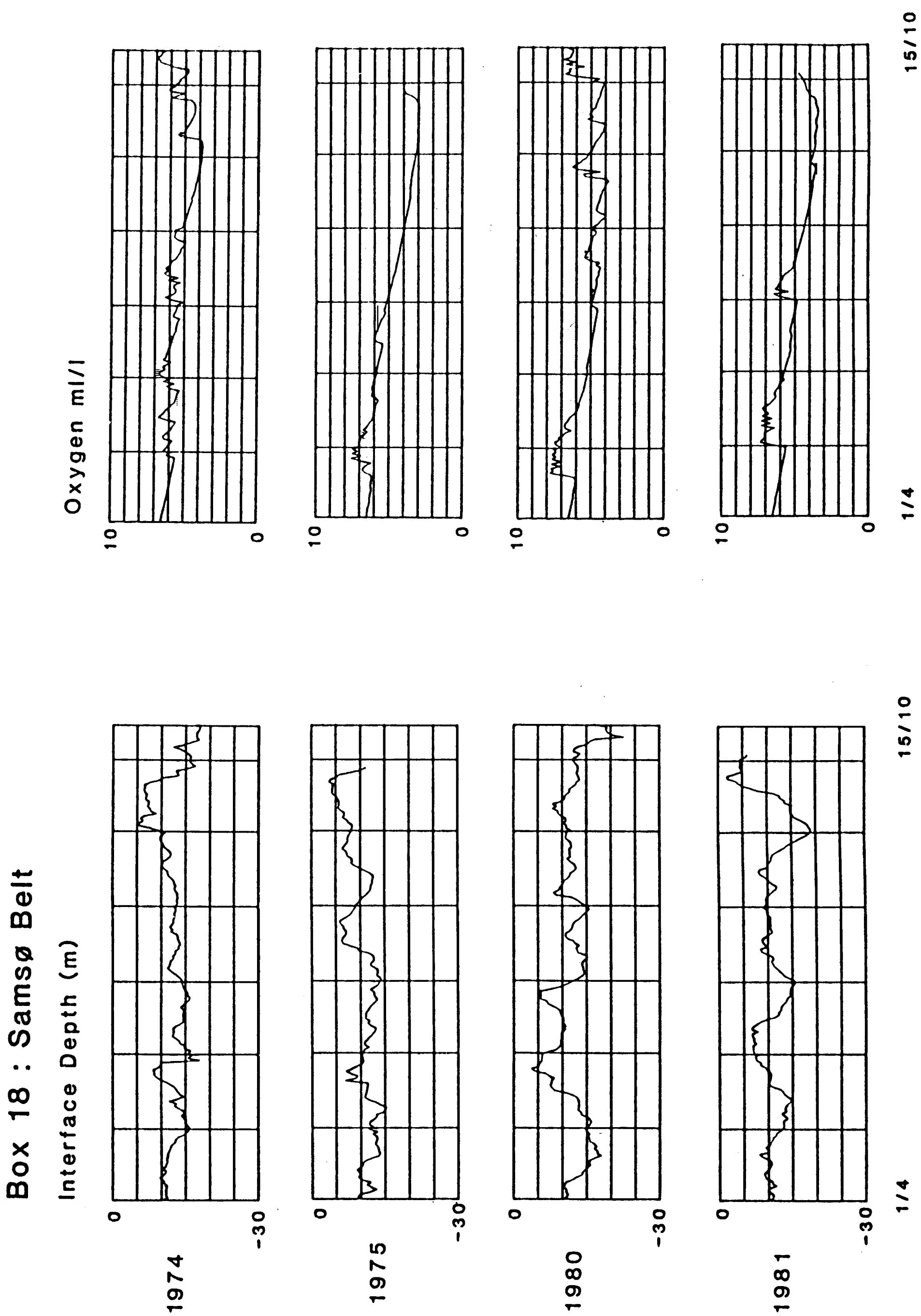


Fig. 8. Time series for calculated oxygen contents for the lower layers in South East Kattegat (area 12) and Samsø Belt (area 18) for the years 1974, 1975, 1980 and 1981.

Modelling of Oxygen Concentration by a first Order Process

In order to determine the effect of meteorological activity on the oxygen content in the Kattegat, it is necessary to express the concentration of oxygen. For this purpose a first order process has been assumed. Admittedly this is a rough approximation. However, comparing different years it will give the relative effect of the meteorology in the years considered.

The following assumptions have been made:

- The vertical mixing during the winter is so efficient, that a concentration of 6.5 ml/l is found in the bottom layer, and the surface layer is saturated.
- The biological and chemical oxygen consumption at the bottom layer is assumed to be proportional to the concentration (a 1st order process), and the inverse timescale λ for the decay is given by

$$\lambda \equiv \lambda_0 2^{(T/10-1)}$$

where λ_0 is the inverse time constant at 10° C (=0.009 day^{-1}). This expression implies that the oxygen consumption is doubled for a temperature increase of 10° C.

- The biological and chemical oxygen consumption of the upper layer is calculated in the same way as for the lower layer if a thermocline has been formed.
- The temperature of the upper and the lower layers have been taken as the mean annual variation measured at Anholt Knob Lightvessel (0 and 30 metres observations used). For the lower layer this was approximated by

$$T = 8.5 + 3.8 \cos \left(\frac{u}{182.5} (j-270) \right) \ {}^{\circ}\mathrm{C}$$

j being the number of the day in the year.

Having thus constructed a sink for the oxygen, the calculation is carried out by means of the equation of continuity for oxygen.

The outcome of the modelling is shown for four selected years 1974, 1975, 1980 and 1981 in Fig. 8. It depicts the oxygen variation in the bottom of area 12 (the south-eastern Kattegat) and in area 18 (Samsø Belt). The lowest oxygen concentrations reached each year in the bottom of the areas 12 and 18 are shown in Fig. 9.

The maximum oxygen content in the end of the summer period varies considerably from year to year and is closely related to meteorological forcing. The bottom layer receives continuously oxygen in the summers of 1961, 1962, 1964, 1969, 1970, 1974, 1977, 1978, 1979 and 1980 (by the end of August) while long periods without renewal until late autumn are seen in 1963, 1966, 1967, 1968, 1969, 1973, 1975, 1976 and 1981. 1971 and 1972 have not been calculated. The differences between the years are pronounced and show that meteorological activity during some years

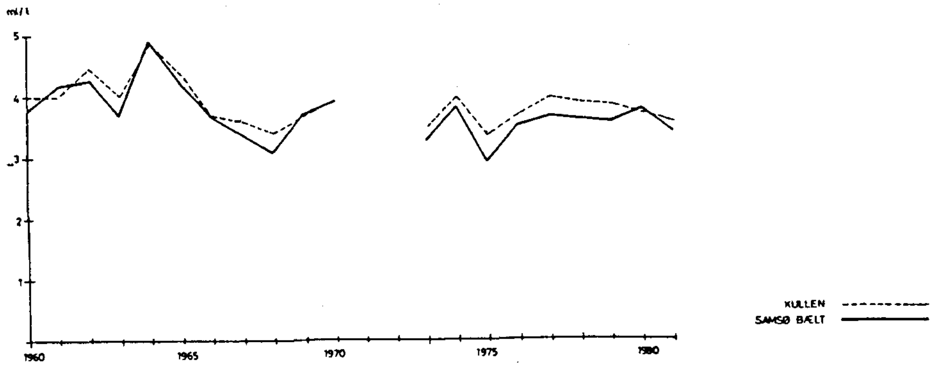


Fig. 9. Calculated minimum oxygen content in South East Kattegat (area 12) and Samsø Belt (area 18) for the period 1961-1981.

in the summer time prevents the oxygen level to become seriously low. 1981 is among the years with unfavourable development in oxygen level.

It is emphasized that the performed calculation are based on a first order oxygen decay process.

According to the present results 1975 reaches a lower oxygen content that 1981. This is in disagreement with observations and the occurrence of fish deaths in 1981 which triggered the entire investigation. However, the above mentioned biological model changes the role of these two years, and the explanation is that nutrients were abundant in 1981 after a heavy run off from land, thus enhancing the biological production. This effect could of course not be described by the simple 1st order model of the oxygen consumption, and this demonstrates the importance of biological modelling with the actual (and time varying) nutritional load.

The following conclusion of the investigation can be drawn:

- the vertical exchange between the upper, brackish and the lower, saline water is strongly dependent on the meteorology.
- A small meteorological activity results in small exchange flows and vice versa for a strong activity.
- It is in years with small meteorological activity that the low concentrations of oxygen appears, because in these years the oxygen supply to the deep layer of Kattegat is smaller due to the smaller exchange flow.

In the model the temperature variation through the year have only been considered on a average basis. A more sophisticated calculation including the correct temperature development and biology will be presented later (Malmgren, Hansen and Niilonen to be published).

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