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On the response of the horizontal mean vertical density distribution in a fjord to low-frequency density fluctuations in the coastal water

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ABSTRACT

Baroclinic water exchange through a fjord mouth, driven by a slowly varying density field outside the mouth, is modelled by a simple quasi-steady frictionless model. It is assumed that a certain fraction of the horizontal pressure difference between the coastal water and the fjord is used to accelerate the fluid into the mouth. The continuous vertical density distribution in the fjord, which changes in response to the water exchange, is modelled using a time-dependent, one-dimensional advective-diffusive "filling-box" type of model. The model has been tested against an almost one-year-long time series of salinity and temperature from the Ørsta fjord (horizontal surface area about 15 km^2) on the Norwegian west coast. It is found that for this particular fjord, the mean externally forced baroclinic water exchange is one order of magnitude greater than the mean water exchange driven by the estuarine circulation (600 and 60 $m^3\ s^{-1}$ respectively). Such a vigorous water exchange between a fjord and the external area implies that the time-averaged concentrations of many biological and chemical species above the sill level in the fjord are approximately equal to those in the coastal water outside the fjords.

1. Introduction

The hydrostatic pressure in a fixed point in the sea can be thought of as composed of the air pressure on the sea surface plus a component due to the hydrostatic pressure from a water column of "standard" density (the surface component) plus an internal component due to the deviation of the seawater density from the chosen standard. The surface component changes as a result of sealevel fluctuations and the internal component changes as a result of fluctuations in the vertical density distribution.

Along a sound connecting a semi-enclosed embayment and the sea, there are almost always non-zero horizontal pressure gradients on all depths down to the sill depth (the deepest connection between the sea and the embayment). Such pressure gradients are usually primarily due to changing surface and internal pressure in the sea outside the sound. However, local freshwater supply and mixing in the embayment may as well cause horizontal pressure gradients and an accompanying (estuarine) circulation.

Many aspects of the surface and internal response of an embayment, e.g., a fjord, upon a changing sealevel outside the mouth have been reported in the literature, see, e.g., Stigebrandt (1980) for a review. The present paper is mainly devoted to the internal response to low-frequency internal pressure (density) variations outside the mouth of a fjord. The accompanying circulation is often termed intermediary circulation since it usually is best seen in the intermediary layer between the surface mixed layer and the basin water below the sill level, cf. Fig. 1.

It has been qualitatively known for a long time that low-frequency fluctuations in the density field outside a fjord mouth may cause an internal response in the fjord (Pettersson, 1920). Jerlov

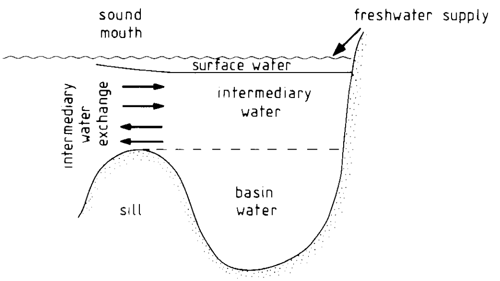


Fig. 1. Sketch explaining some of the nomenclature used in this paper.

(Johnson, 1943) showed a beautiful example of large amplitude isopycnal displacements in the Gullmar fjord, Sweden. The fjord was ice-covered so local wind stress could be ruled out as the driving agent. Although no simultaneous measurements were available from the sea outside the mouth, it was concluded that the isopycnal displacements in the fjord were driven by similar displacements outside the fjord. Later examples of externally imposed vertical isopycnal displacements in fjords may be found in, e.g., Svendsen (1980), Holbrook et al. (1983), Stucchi (1983) and Djurfeldt (1987). In their review paper, Matthews and Heimdal (1980) referred some early examples of biological evidence for intermediary water exchange and Lindahl (1987) showed that in less than a week a plankton community residing in the Gullmar fjord could be advectively washed out and replaced by a different community imported from outside the fjord.

The fjordic response to a slowly varying density field outside the mouth has been modelled only for rather idealized stratifications. Klinck et al. (1981) used a two-layer stratification. The off-fjord variation of the depth of the interface was caused by downwelling/upwelling due to alternating onshore/offshore Ekman transports in the coastal water.

Internal pressure changes in the sea may possibly induce a great variety of responses in a fjord. Different kinds of wave motions which may match the boundary conditions imposed by the topography of the mouth might constitute a broad class of possible responses. Proehl and Rattray (1984) studied the wave response of a deep and wide estuary with continuous vertical stratification to off-estuary upwelling/downwelling. Since the estuary considered is wider than the internal Rossby radius of deformation rotational effects are important. The low-frequency response was shown to take the form of internal Kelvin waves propagating up the estuary where they were assumed to dissipate. Finite changes of the (horizontal mean) vertical density distribution in the estuary due to the water exchange with the sea, were not considered by these authors.

The aim of the present paper is to investigate the response of the horizontal mean vertical density distribution in a fjord to low-frequency density fluctuations in the coastal water. No attention will be paid to the causes of the latter. These are considered known, either from measurements or from computations using a model for the coastal dynamics. We confine our analysis to comprise only fjords with narrow mouths implying that rotational effects may be neglected. We focus on the low-frequency part of the forcing spectrum.

We will use a high vertical resolution of the density structure and the flow field in the fjord mouth. We then need to describe the vertical density structure in the fjord with the same high resolution. The one-dimensional (horizontally integrated) model applied to the Baltic Sea in Stigebrandt (1987) is considered adequate for modelling slow changes of the vertical density distribution in the fjord. The model has a high vertical resolution (1 m). It is also capable of modelling the mixed layer at the sea surface and the estuarine circulation caused by local freshwater supply and local wind mixing and heat exchange through the sea surface. The mixed layer dynamics is according to Stigebrandt (1985). In this way, some aspects of the interaction between the intermediary circulation (driven by the coastal internal density fluctuations) and the estuarine circulation may be studied as well. The fact that the fjord model is horizontally integrated of course implies that waves within the fjord cannot be resolved.

The coupling between the coastal zone and the fjord is by a specific dynamical model. The time-dependent horizontal pressure gradient field close to the mouth, caused by different vertical density distributions outside and inside the mouth respectively, drives intermediary currents into and out of the fjord. The transports through the mouth are such that the vertical density distribution in the fjord tends to adapt to the changing density structure outside the mouth. The response of the mouth flow and the vertical density distribution in the fjord to a specified forcing is expected to depend on the topographical properties of the fjord. In particular the width and depth of the mouth (influencing the baroclinic transport capacity of the mouth) and the horizontal surface area of the fjord (determining the storage capacity of the fjord) should be important quantities.

This paper is organized as follows. In Section 2, the hydrodynamical coupling of the fjord and the sea is discussed and a model for the flow in the mouth is presented. In Section 3, the response of the vertical density distribution in the fjord to the flow in the mouth is discussed. Measured time series of salinity and temperature, obtained within and outside the Ørsta fjord, are described in Section 4. The model test against these field data is presented in Section 5. The paper is concluded by some remarks in Section 6.

2. Modelling of the flow in short sounds

The model for the flow in the mouth to be developed here should be simple, if possible, since its main use will be in a physical-biochemical circulation model for fjords. For simplicity we will consider the flow through the mouth as quasi-steady. This excludes the study of the highest frequencies of forcing since quasi-steadiness requires that the time scale T of flow fluctuations is much longer than L/U where U is the maximum speed and L is the length over which the fluid is accelerated towards the mouth. We will also assume that the mouth (sound) is short which implies that frictional effects in the mouth may be neglected. Ideally we may think of the mouth as a slit of variable width in a thin vertical wall.

There are a number of questions about the detailed properties of the flow field in and close to the mouth that cannot be answered at present. Is all of the total horizontal pressure difference between the coastal water and the fjord used to accelerate the fluid into the mouth? In which way do different "flow layers" contract or expand when flowing through the mouth? Are wave properties of the flow and internal hydraulic controls important aspects to consider? For each level, we will assume that there is a direct relationship between the velocity in the mouth and the total horizontal pressure difference at the same level. After having derived an expression for the velocity in the mouth we will apply this to some classical two-layer cases. Hereby we may estimate an empirical coefficient which may be interpreted to be equal to the fraction of the total horizontal pressure difference which appears to be used to accelerate the fluid into the mouth.

The vertical coordinate z is counted positive below the geoid while sealevel at time t, $a(t)$, is counted positive above the geoid. The pressure $p(z,t)$ at the level z is then

$$p(z,t) = g\int_{-a(t)}^{z} \rho(z,t)\,dz + p_a(t), \tag{2.1}$$

where $\rho(z,t)$ is the density and $p_a(t)$ is the atmospheric pressure. We introduce a standard density ρ_0 and rewrite eq. (2.1) in the following way

$$p(z,t) = g\int_{-a(t)}^{z} [\rho(z,t)-\rho_0]\,dz + g\rho_0[z+a(t)] + p_a(t). \tag{2.2}$$

Since most sounds of interest are fairly short, we may safely neglect the atmospheric pressure difference along the sound. The instantaneous horizontal pressure difference between the coast (index 1) and the fjord (index 2) then has two components, the internal pressure difference DPI(z) and the surface pressure difference DPS defined by

$$\mathrm{DPI}(z) = g\int_{a_m}^{z} [\rho_1(z)-\rho_2(z)]\,dz, \tag{2.3}$$

$$\mathrm{DPS} = g\rho_0(a_1-a_2). \tag{2.4}$$

In eq. (2.3), $a_m = \min(a_1, a_2)$ and we have neglected a term of magnitude $g(\rho_{1(2)}-\rho_0) \times |a_1-a_2|$, where $\rho_{1(2)}$ is the mean density close to the sea surface in the coastal water (fjord). Since the density difference is two orders of magnitude smaller than the reference density the magnitude of the neglected term is only a few % (at most) of the magnitude of DPS. In this paper, we will for simplicity assume that the sea levels are close to $z=0$ which allows us to use $a_m=0$ in eq. (2.3).

Thus, by vertical integration of the density profiles inside and outside the sound respectively the instantaneous horizontal internal pressure difference along the sound DPI(z) may be computed at each level z above the sill level. We don't know if the vertical density distribution in the fjord (coastal area) is representative of the vertical density distribution just inside (outside) the sound. This would require that there are very efficient mechanisms for horizontal exchange on both sides of the sound. There is generally also a surface pressure difference DPS between the coastal area and the fjord caused by an unknown difference in sea level.

Water is assumed to be accelerated down the horizontal pressure gradient without suffering from frictional losses (Bernoulli flow). The fjord and the area outside the mouth are both assumed to be wide compared to the width of the mouth. The far upstream speed of water flowing towards the mouth may then be assumed to be negligible. Assuming that the fraction α of the total horizontal pressure difference is used to accelerate the fluid into the mouth the (quasi-steady) speed $u(z)$ of the water in the mouth should be given by

$$u(z) = \varepsilon\{\varepsilon\alpha 2[\mathrm{DPI}(z) + \mathrm{DPS}]/\rho_0\}^{1/2}, \tag{2.5}$$

where we have used the Boussinesq approximation, u is taken positive for flow into the fjord and ε is defined in the following way:

$$\varepsilon = \begin{cases} 1 & \text{if } \mathrm{DPI}(z) + \mathrm{DPS} > 0, \\ -1 & \text{if } \mathrm{DPI}(z) + \mathrm{DPS} < 0. \end{cases} \tag{2.6}$$

The instantaneous horizontal surface pressure difference between the coastal area and the fjord DPS is determined from the condition of volume conservation. This requires that the net flow through the mouth is equal to the rate of freshwater supply, Q_f plus the rate of volume change $Q_b = A\,\mathrm{d}a_2/\mathrm{d}t$. Here A is the horizontal surface area of the fjord. Q_b may be derived from measurements or computations of the barotropic response of the fjord to the barotropic forcing. Thus

$$\int_0^{hs} u(z)\,B_m(z)\,\mathrm{d}z = Q_f + Q_b, \tag{2.7}$$

where $B_m(z)$ is the width of the mouth and $z = hs$ is the sill level.

In the derivation above, we have assumed that the rotation of the earth does not influence upon transports through the mouth. This means that, when considering periods greater than the inertial period, the width of the mouth must be at most of the order of the appropriate internal Rossby radius.

It should be noted that strong, fluctuating barotropic currents (e.g., tidal currents) in the fjord mouth may impede the development of a baroclinic current structure during some of the time. For a two-layer system this occurs when a densimetric Froude number exceeds a certain critical value. The interaction between barotropic and baroclinic (two-layer) currents in a fjord mouth was investigated by the present author (Stigebrandt, 1977). In that paper it was demonstrated that weak barotropic currents do not influence the baroclinic mean transports through a mouth. It is not evident if and how these two-layer results on the interaction between barotropic and baroclinic currents in a fjord mouth may be generalized to a continuous density distribution.

2.1. Comparison with classical two-layer flows

Using $Q_b = 0$, eqs. (2.3)–(2.7) constitute a complete model for the baroclinically forced, quasi-stationary (intermediary) water exchange between the fjord and the coastal area. However, we have to determine which value should be assigned to the empirical constant α. This is done below by comparisons with classical two-layer results (Stommel and Farmer, 1953) for a rectangular (constant width) mouth.

First, we consider a case with a thin (compared to the sill depth) light layer (density ρ) of thickness h in the fjord resting upon sea water. Outside the fjord there is only seawater (density $\rho + \Delta\rho$). We then obtain $\mathrm{DPI}(z) = g\Delta\rho z$ ($z \leqslant h$). Since the thickness of the lower layer in the mouth is great compared to h the speed of this layer can be neglected. We then require zero horizontal pressure gradients below the depth h (this condition replaces the use of eq. (2.7)), i.e., $\mathrm{DPI}(h) + \mathrm{DPS} = 0$, and we obtain $\mathrm{DPS} = -g\Delta\rho h$. From eq. (2.6), $\varepsilon = -1$ for $z \leqslant h$ and from eq. (2.5) we obtain for the velocity in the mouth:

$$u(z) = -(2\alpha g'(h - z))^{1/2}, \quad (z \leqslant h)$$

where $g' = g\Delta\rho/\rho_0$. The minus sign means that the flow is out of the fjord. The mean velocity of the light layer of thickness h is $u = -\frac{2}{3}(2\alpha g' h)^{1/2}$. The transport computed by the model is uh. It is known that for a narrow mouth the outflowing layer will contract upon accelerating towards the mouth and $h_m = \frac{2}{3}h$ where h_m is the thickness of the surface layer in the mouth (see e.g., Stigebrandt, 1980b). According to the Stommel and Farmer theory the squared densimetric Froude number in the mouth, $F^2 = u^2/(g' h_m)$, should be equal to 1 for this case. Thus $u = -(g' h_m)^{1/2}$ and the transport through the mouth is $-h_m(g' h_m)^{1/2}$. To obtain this result from our model α has to be equal to $\frac{1}{3}$.

Next, we consider the case when there are two vertically homogeneous fluids above the sill level ($z = H$) on the two sides of the mouth. The density difference between the fluids is $\Delta\rho$ with the denser water in the sea. This is equivalent to Stommel and Farmers so-called overmixed case. For DPI(z) we obtain $\mathrm{DPI}(z) = g\Delta\rho z$ ($z \leqslant H$). We require no net transport through the mouth. (This is equivalent to the condition expressed by eq. (2.7).) By reasons of symmetry it is easily realized that this is equivalent to the requirements of no current and zero horizontal pressure difference at mid-depth of the mouth (i.e., at $z = H/2$). Thus $\mathrm{DPI}(H/2) + \mathrm{DPS} = 0$ which gives us $\mathrm{DPS} = -g\Delta\rho H/2$. The total horizontal pressure gradient hence switches at half the sill depth. For $u(z)$ we then obtain

$$u(z) = \varepsilon(\varepsilon 2\alpha g'(z - H/2))^{1/2}.$$

The mean velocity above the depth $H/2$ in the mouth (where $\varepsilon = -1$) then is $u = -\frac{2}{3}(2\alpha g' H/2)^{1/2}$ and the transport through the upper half of the mouth is $uH/2$. According to the Stommel and Farmer theory the densimetric Froude number for the upper layer in the mouth should be equal to $\frac{1}{2}$. Thus $u^2/(g' H/2) = \frac{1}{2}$ and the transport in the upper layer is equal to $-(\frac{1}{2}g' H/2)^{1/2} H/2$. Equating the two transports gives $\alpha = \frac{9}{16}$.

Thus we obtain different values of α for the two cases above.

For our pragmatic approach, we simply adopt the value $\alpha = \frac{1}{2}$, which is not far from the average of α for the two cases. This α-value implies that half of the horizontal pressure gradient is used to accelerate the fluid into the mouth. The current speed is reduced by the factor $\alpha^{1/2} = 0.71$ compared to the case where all the horizontal pressure difference is used to accelerate the fluid into the mouth and there are no effects of contraction or expansion.

The flows in the cases considered above are both driven by large horizontal pressure gradients which give rise to large velocities. These in turn are determined by internal hydraulic controls. Thus the wave aspect of the flows is important for these cases. For weak horizontal pressure gradients, however, the flow will not approach the speed of internal waves and it is possible that it would be more correct to use $\alpha = 1$ for such cases. To further investigate this problem is outside the scope of the present paper and we use $\alpha = \frac{1}{2}$ in this work.

3. The fjordic response to the mouth flow

The water flowing into the basin is assumed to be interleaved immediately at the level where it is neutrally buoyant. Thus the depth of interleaving is usually different from the depth at which the water enters the fjord through the mouth. Very dense water intermittently intruding over the sill may feed an entraining dense bottom current which occasionally may reach down to the greatest depths in the fjord. For a simple model for dense bottom currents the reader is referred to Stigebrandt (1987). The fjord model described below is largely adopted from that paper.

The response of the vertical density distribution in the fjord upon the flow in the mouth is computed as described below. We consider a horizontal slab of the basin of thickness dz at the depth z. The surface area of the basin is $A = A(z)$. The salinity $S = S(z)$ of the slab changes according to

$$VS_t = (wAS)_z \,\mathrm{d}z + q \,\mathrm{d}z S_{\mathrm{in}}, \tag{3.1}$$

where subscripts $_t$ and $_z$ denote differentiation with respect to time and depth, respectively. $V = A(z)\,\mathrm{d}z$ is the volume of the slab, $w = w(z)$ is the vertical velocity induced by the in- and outflow through the mouth, $q = q(z)$ is the flow rate per unit height (dimension: $\mathrm{L^2\,T^{-1}}$) of water flowing into the slab {of salinity $S_{\mathrm{in}} = S_{\mathrm{in}}(z)$}. For negative q (horizontal flow out of the slab), S_{in} is equal to S.

Continuity demands that

$$q + (wA)_z = 0. \quad (3.2)$$

Eqs. (3.1) and (3.2) then give:

$$S_t = (1/A)\{wAS_z + q(S_{in} - S)\}. \quad (3.3)$$

A similar equation may be derived for the temperature T. Eq. (3.3) does not account for effects of vertical diffusion of salt within the basin. However, in the model vertical diffusion below the surface mixed layer is computed, see Stigebrandt (1987) for details. The vertical diffusivity below the surface mixed layer in the present model is according to Stigebrandt and Aure (1989) who showed that the diffusivity may be computed from the energy flux to internal tides at the sill and the local vertical stratification. For details about the model computations of the properties of the mixed surface layer the reader is referred to Stigebrandt (1985).

In summary, the computations of the fjordic response to the mouth flow are performed in the following way. From eqs. (2.3)–(2.7), we compute $q_m(z) = u(z)B_m(z)$ in the mouth. The depth of interleaving, at the level of neutral buoyancy, is thereafter computed for positive q_m whereby $q(z)$ is determined. Utilizing eq. (3.2), we next compute w. Finally, S_t is computed from eq. (3.3) {and T_t from a similar equation}. Below we compare model predictions with field data for a specific fjord. For simplicity we have used $Q_b = 0$ in the computations displayed here.

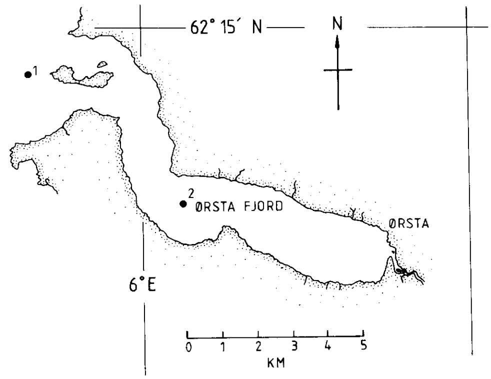


Fig. 2. Map over the area. The islands in the mouth are situated on the sill. The filled circles denoted by 1 and 2 show the observational sites.

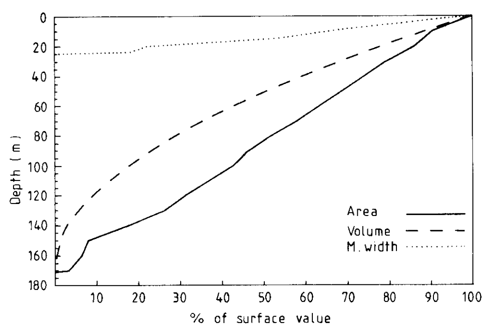


Fig. 3. The vertical distributions of the horizontal surface area, the volume and the width of the mouth in the Ørsta fjord normalized by the values at the sea surface. The absolute values at the sea surface are given in the text.

4. Field observations from the Ørsta fjord

The model will be applied to the Ørsta fjord (Fig. 2). The head of the fjord is situated at about 62°12′N and 6°7′E. This fjord has the horizontal surface area 15.52 km^2, volume of 1.3 km^3, sill depth 24 m and maximum depth 172 m. The width of the mouth at the sea surface is 1.14 km. The vertical distributions of the fjord horizontal surface area, the fjord volume and the width of the mouth respectively are shown in Fig. 3. The semidiurnal tide has an amplitude of about 0.9 m. This gives a barotropic r.m.s. speed in the mouth approximately equal to 0.07 m s^{-1}. Meteorological observations (used to force the mixed layer submodel) from the meteorological station Hjelvik (in another fjord system about 80 km NE of the Ørsta fjord) were obtained from the Norwegian Meteorological Institute. The daily mean freshwater supply to the fjord was obtained from the Norwegian State Electricity Board (NVE). It may be mentioned that the annual mean freshwater supply to the fjord is about 15 m^3 s^{-1}.

During the period 11 July 1986–10 June 1987, vertical temperature and salinity profiles, down to 50 m below the sea surface, were obtained in three fixed positions approximately once a week (Fig. 2). The measurements were made with an "Electronic Switchgear" (see Aure and Stigebrandt, 1989b) which was calibrated weekly in connection with the field measurements, giving an error of less than 0.1 units in salinity (ppt) and temperature (K). The density field in vertical 1 (outside the mouth) was used to force the model. The time evolutions of temperature and salinity in this vertical are displayed in Figs. 4a and 5a respectively. From these time series we constructed, by linear interpolation, salinity and temperature time series, with time and vertical spacings of 1 day and 1 m respectively. These series of course lack variability on time scales less than one week. The measured time evolutions of temperature and salinity in vertical 2, in the fjord, are displayed in Figs. 4b and 5b respectively.

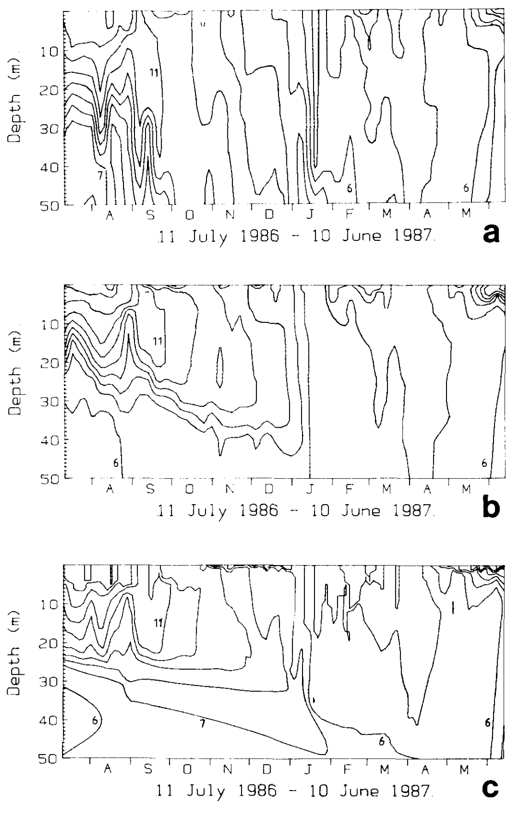


Fig. 4. The time evolution (11 July 1986–10 June 1987) of the vertical temperature distributions: (a) measured outside the fjord (vertical 1), (b) measured inside the fjord (vertical 2) and (c) computed for the fjord from the model (contour interval 1 °C).

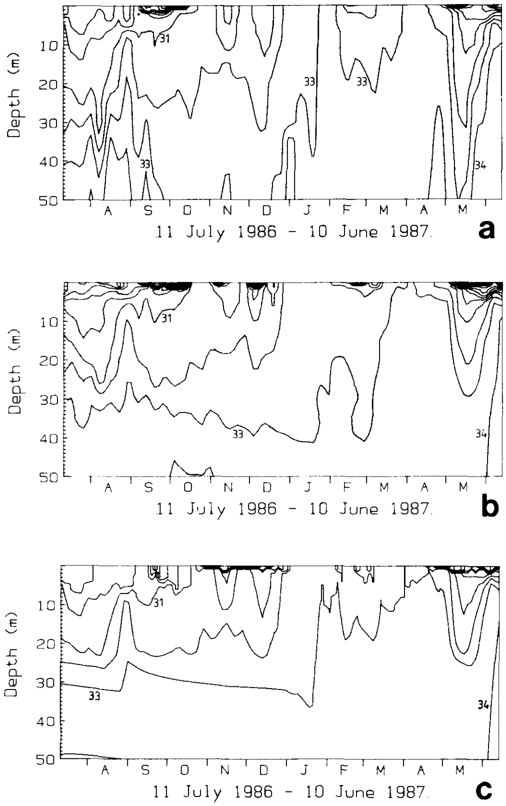
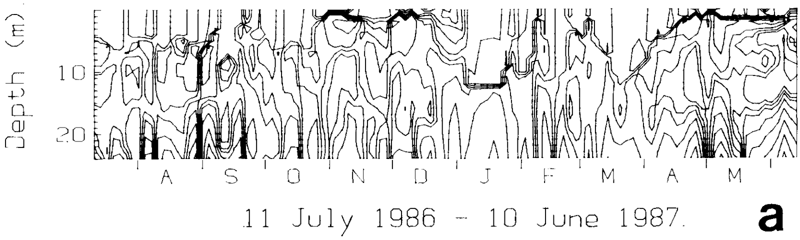


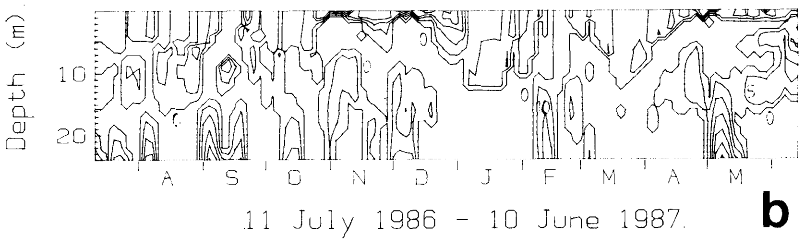
Fig. 5. The time evolution (11 July 1986–10 June 1987) of the vertical salinity distributions: (a) measured outside the fjord (vertical 1), (b) measured inside the fjord (vertical 2) and (c) computed for the fjord from the model (contour interval 1‰).

5. Model results

The computed evolution of the vertical distributions of temperature and salinity in the fjord (Figs. 4c and 5c) is quite similar to the measured evolution (Figs. 4b and 5b). Most major events of intermediary water exchange are apparently well captured by the model. The similarity between the measured and computed fields of salinity and temperature is a strong indication that the water exchange between the fjord and the coast is reasonably well modelled.

The computed daily mean current speeds in the mouth are shown in Fig. 6a. In Figs. 6b and 6c are shown the negative (out of the fjord) and the





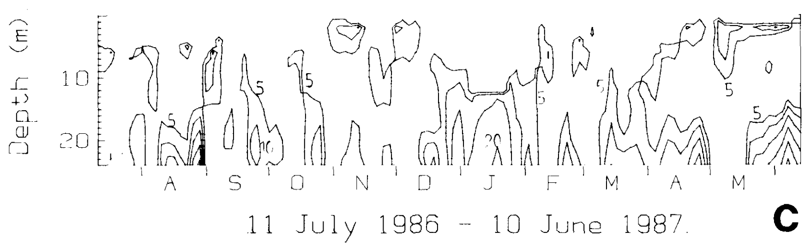


Fig. 6. The computed daily mean currents in the mouth of the Ørsta fjord in the period 11 July 1986–10 June 1987: (a) shows the complete current field, (b) shows the outflow part (isolines start at 0 cm/s) and (c) the inflow part (isolines start at 5 cm/s) (contour interval 5 cm/s).

positive (into the fjord) current components respectively. All contour plots showing model predictions were constructed using output from the model for each third day. It may be seen that the current in the mouth often has several "layers" with respect to the direction of the current. Two and three layers seem to be most common. In periods of strong downwelling at the coast the flow into the fjord near the sea surface may be so strong that the weak estuarine circulation may be arrested (in the middle of September 1986). In periods with large freshwater supply (in November and December 1986 and May 1987 when the monthly means are about 30 $m^3 s^{-1}$) the estuarine circulation is well developed and a so-called compensation current (directed into the fjord) just beneath the outflowing brackish water can be clearly seen in the depth interval 2–5 m (Fig. 6c). No current measurements were undertaken in the fjord mouth during this field program.

The strongest predicted currents in the mouth occur close to the greatest depth (the sill depth) in periods of strong upwelling (at the end of August 1986, at the end of April 1987 and in the beginning of June 1987—see Fig. 6c) or strong downwelling (in the beginning of September 1986 and in May 1987—see Fig. 6b).

The amplitude of the barotropic semidiurnal tidal current in this particular fjord mouth is for most of the time relatively small compared to the computed amplitude of the baroclinic currents.

Hence we do not expect the baroclinic mean transports through the mouth to be much affected by the barotropic tide.

A comparison between the variabilities of the measured (Figs. 4b and 5b) and the computed fields in the fjord (Figs. 4c and 5c) shows that the variability of the measured fields is greater than that of the computed fields. The main reason for this should be that there really is variability in the fjord on time scales less than 1 week (the measurement interval). These short time scale variations cannot, however, be captured by the model which is forced by the weekly measurements outside the fjord. Some of the measured variability may as well be due to measurement errors caused by ship drift and the presence of internal waves. It is not hard to find deviations which one would like to understand. For instance in late January the fjord is approximately homothermal down to 50 m (Fig. 4b) but there is a vertical salinity stratification (Fig. 5b). The model however shows in addition to a salinity stratification also a thermal stratification (Figs. 4c and 5c). This might be due to several reasons, the most likely is perhaps that the meteorological forcing is not very exact since it is derived from meteorological observations made in another fjord system.

The computed average intermediary inflow to the fjord is about 600 $m^3 s^{-1}$ (i.e., the normalized inflow is about 40 $m^3 s^{-1}$ per km^2 of the horizontal surface area of the fjord). This figure also includes the inflow due to the estuarine circulation. However, the average freshwater supply to the Ørsta fjord is about 15 $m^3 s^{-1}$ and with a mean salinity of the outflowing brackish water of about 25‰ and a mean salinity of about 31‰ of the inflowing seawater (participating in the estuarine circulation) the mean estuarine inflow would be of the order of 60 $m^3 s^{-1}$, i.e., one order of magnitude less than the intermediary circulation, driven by the density fluctuations in the coastal water.

6. Concluding remarks

Computations presented here indicate that the intermediary circulation, driven by the density fluctuations in the coastal water, might be the dominating mode of water exchange in many fjords. The application of the model to a large number of fjords in the same region of the Norwegian west coast shows that the normalized mean intermediary water exchange usually is in the range 30–100 $m^3 s^{-1}$ per km^2 of horizontal surface area of the fjord (Aure and Stigebrandt, 1989b). This is usually one or two orders of magnitude greater than the estuarine circulation.

The vigorous intermediary circulation above the sill level implies that the vertical distributions of different state variables in the fjord almost always is quite similar to those in the coastal water (allowing for a certain phase delay). This is also true for the distributions of many biological species, cf. Lindahl (1987). Aure and Stigebrandt (1989a) found that the vertical flow (sedimentation) of organic matter down into the basin waters of a large number of fjords in the present area is a strong function of the sill depth. We concluded that this implies that the (presumably fairly horizontally homogeneous) conditions in the coastal water determines the vertical distribution of the concentration of organic matter in the fjords and thereby the flow rate of organic matter into the sill basins. Local supplies of plant nutrients normally seem to have little effect upon the rate of sedimentation of organic matter below the sill level and thereby also little effect upon the oxygen consumption in the sill basins.

The present model considers only short sounds whereby frictional effects may be explicitly neglected. For long sounds it might be necessary to include frictional effects in the model. In extreme cases, the turbulence generated by friction against the seabed in the sound may homogenize the water in the sound. Long sounds may also buffer the water exchange and thereby reduce the fjordic response. If λ is the length and v is a typical speed in the sound the latter will not be efficiently flushed if λ is greater than or of the same magnitude as vT, where T is the period of the fluctuating current.

The dynamics of the time varying flow between two continuously stratified water bodies were in the present paper treated in a simplified but probably physically reasonably sound way. Being interesting in its own right, and being quite common in nature and therefore of great practical interest, the detailed mechanics of this kind of flow certainly merit further investigations. Among other things one would like to understand where the excess pressure gradient goes (i.e., that part not used to accelerate the fluid into the mouth). Since α was determined by comparison with frictionless flows of large amplitudes we guess that the excess pressure gradient for these cases is used for circulation and mixing downstream of the mouth. One also would like to understand how the wave aspects of the flow influences α. The flow regime with small to moderate velocities in the mouth (compared to the speed of internal waves) seems to be particularly interesting.

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