The mountain trembled to its very base, and the rock rocked. I threw myself upon my face, and clung to the scant herbage in an excess of nervous agitation.

"This", said I at length to the old man, "this can be nothing else than the great whirlpool of the Maelstrom".

"So it is sometimes termed", said he. "We Norwegians call it the Moshoestrom, from the island of Moskoe in the midway".

EDGAR ALLAN POE

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COASTAL WIND EFFECTS
ON FJORD CIRCULATION

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ABSTRACT

The influence of wind driven, coastal circulation on a narrow fjord is investigated with a linear, two-layer model. The major assumption in this study is that the Coriolis acceleration is important in the coastal regime, but is unimportant in the narrow fjord because the cross-fjord flow vanishes. Diffusion of mass and momentum is also neglected in the present formulation. A thin (~2 m), fresh, surface layer, which appears in most fjords, is not included in this analysis because its dynamics depend on mixing.

For a wide range of model parameters and wind conditions, wind forced, geostrophic, alongshore coastal currents have a strong effect on circulation within the fjord. These currents control the free surface and pycnocline displacements at the fjord mouth, which affect pressure gradients within the fjord. Classical fjord theory assumes hydraulic control by topographic constrictions at the mouth; the present mechanism is an alternative to these classical ideas.

Model results show that the free surface reflects the pycnocline shape, i.e., the pressure gradients in the lower layer almost vanish. Also, alongshore and across-shore winds affect the fjord differently. Alongshore winds, through Ekman flux, cause a net volume change in the fjord; while, across-shore winds simply set up the surface with no net volume change.
We find that strong coastal winds can produce large velocity shears across the pycnocline in the fjord which can have a significant effect on turbulence and diffusion.

Data from Norwegian fjords, Western Canadian fjords and the Strait of Juan de Fuca support this idea of fjord circulation produced by wind driven, coastal currents.

**INTRODUCTION**

Since the Second World War, there has been increased interest in the study of deep silled fjord dynamics. One of the earliest of these studies (Saelen, 1946) indicated the importance of horizontal exchange between the fjord and coastal waters. He found that horizontal exchanges could account for differences between calculated and observed temperature distributions above sill level in Nordfjord. Similar arguments were put forward to explain hydrographic conditions in fjords on the western coast of North America (Picard, 1961; Herlinveaux, 1962).

Recent investigations of fjords with deep sills on the western coast of Norway indicate that coastal waters act as an outer boundary to the intermediate layer [the layer below a thin (~2 m) surface freshwater layer] and that the mean circulations are shown to be closely connected with wind induced fluctuations in the density field of adjacent coastal waters (Svendsen, 1980).

At one time, stratification of coastal water near fjords was thought to be controlled by seasonal runoff variations or by variations in the water properties of offshore currents. However, coastal upwelling studies of the 60's and 70's showed that the major changes in coastal stratification are wind induced (see Smith, 1968). This is confirmed by studies on the coast of Norway (Svendsen, 1971; Devold, 1972; Hermansen, 1974) which demonstrate that the major changes in coastal stratification are in response to winds.

Cannon (1971) found three distinct flow reversals in the Juan de Puca submarine canyon which appeared to be related to coastal wind reversals. Svendsen and Utne (1973) suggest that renewal of deep water in Hardangerfjord is related to prolonged periods of upwelling favorable coastal winds. A more detailed investigation of the Byfjord (Helle, 1975, 1978) clearly confirms this hypothesis.

In addition to coastal influences, processes within the
fjord can produce circulation. Variations of vertical turbulent mass transport within a fjord causes pressure gradients which induce currents. This diffusively driven circulation is usually quite weak and can be dominated by the stronger circulation due to coastal upwelling and downwelling. This overpowering of the diffusive circulation was evident in Josenfjord (Svendsen, 1977) where the mean flow in the upper part of the intermediate layer was opposite to the mean horizontal pressure field.

The above studies indicate that there is considerable interaction between circulation patterns in fjords and on adjacent continental shelves. To investigate the nature of these interactions, we constructed a model which simulates a narrow fjord connected to an adjacent coastal area.

The major assumption of our model is that the Coriolis acceleration is important in the dynamics of coastal circulation, but unimportant for fjord dynamics because the cross fjord flow vanishes in a narrow fjord. The transition in the physics between the coastal and fjord regimes is abrupt; no attempt is made to reduce slowly the Coriolis acceleration in the fjord. For simplicity, nonlinear effects and mixing are neglected but can be included in later models.

Several important results are obtained from this simple model. Most notable is that the offshore geostrophic circulation strongly controls hydrographic conditions and circulation within the fjord. Classical fjord circulation theory assumes that sills or constrictions at the mouth are the controlling factors (Stommel and Farmer, 1943; Long, 1975; Pearson and Winter, 1978; Stigebrandt, 1979, 1980). The present research shows an alternate control mechanism: offshore geostrophic circulation.

The model responds differently to alongshore and across-shore wind stress. If the model is forced by across-shore (up- and down-fjord) winds, there is an induced tilt in the free surface and pycnocline, but the total volume of water within the fjord remains constant. On the other hand, if the wind is strictly alongshore, a net transport into or out of the fjord produces flooding or emptying of the fjord as a whole.

Two final observations about fjord dynamics are made from the model results. The free surface slope is a reflection of
the pycnocline slope, i.e., it is baroclinic, not barotropic. The velocity shear is large in some cases which can have a strong effect on vertical diffusion.

Section 2 is a discussion of the assumptions used to construct the model. In addition, governing equations and boundary conditions are presented, together with the numerical solution techniques. Model results for a typical deep, narrow fjord forced by a constant wind stress are presented in Section 3. The effect of variable winds and sills are considered in Section 4. A simulation of the Strait of Juan de Fuca is also discussed. Section 5 presents conclusions of the research and points to possible extensions of the model.

MODEL FORMULATION

The principal question investigated in this study is the nature of the dynamical interactions of a rotating, stratified, wind driven coastal regime with a non-rotating, stratified narrow fjord. Our model is constructed by formulating the appropriate governing equations for the coast and fjord independently and matching the solutions at the junction point of the two regimes.

Fjord Model

Typical fjord widths are in the 2-5 km range while a typical baroclinic Rossby radius of deformation \((\lambda = (g'H)^{\frac{1}{2}}f^{-1})\) is about 10 km \((g' = 2 \text{ cm sec}^{-2}, H = 50 \text{ m}, f = 10^{-4} \text{ sec}^{-1})\). In addition, the narrowness of the fjord reduces the cross fjord flow to zero, which removes the influence of the Coriolis acceleration. Therefore, rotational dynamics will have, at best, only secondary importance inside narrow fjords. In wide estuaries, this will not be true.

The influence of density stratification in a silled fjord is determined from a series of density measurements by Svendsen (1977) in Josenfjord on the west coast of Norway. A typical density profile (Fig. 1a) was used to calculate the squared
buoyancy frequency (Fig. 1a) from which vertical modes were obtained. (For a discussion of vertical modes, see Pedlosky, 1979, p. 356 ff). The first five modal functions were calculated; the first and second are shown in Fig. 1b, c. The second eigenvalue (equivalent depth) is approximately half of the first and the rest of the eigenvalues decrease as \( n^{-2} \) where \( n \) is the mode number. Presently, we are comparing current meter records from Josenfjord with the vertical modal functions to determine the number of layers which would represent the fjord circulation.

We have constructed a two layer model as a first approximation. Later models can incorporate more layers if necessary. This analysis ignores a thin (<2 m thick) layer of brackish \( (\sigma_t < 10) \)}
water which appears in most fjords for most of the year. The dynamical importance of this fresh layer is discussed later in light of some model results.

The model is kept as simple as possible by excluding nonlinear and vertical mixing effects at this point. Shear stress is applied at the surface, but no vertical frictional effects are included at the interface or bottom. A small amount of horizontal friction is included (A = 10^5 cm^2 sec^{-1}).

**Shelf Model**

The coastal upwelling dynamics of Hurlburt and Thompson (1973) are used to model the coastal area. Their results indicate that an alongshore pressure gradient is crucial in the dynamics of coastal upwelling. In a vertical plane model, this pressure gradient is retained implicitly by formulating a vorticity equation and allowing f to vary, i.e., the coastal dynamics are on a β-plane.

As in the fjord, nonlinear and mixing processes are ignored except for surface stress, vertical friction is not included. Also, a small horizontal friction effect is included, mainly to bring alongshore velocities to zero at the coast.

**Further Model Considerations**

The present formulation ignores one subtlety which was pointed out by Dr. Martin Mork (personal communication): there is an implicit assumption in the model that the oceanic regime extends to infinity in the alongshore direction while the fjord is narrow in that direction. No attempt is made at this point to address theoretically the dynamical significance of the two different alongshore length scales.

However, since the oceanic model is on a β-plane, the vertically integrated transport at any position is nearly zero (due to the north-south pressure gradient). Therefore, the circulation is confined basically to vertical planes and is independent of the width of each regime.

**Model Equations**

The equations of motion for transport in the oceanic part
of the two-layer model, consistent with the above considerations are

\[ u_{1t} = fu_1 - gh (h_1 + h_2 + D) x + \tau x + Au_{1xx} \]  
\[ -\beta v_1 - fu_{1x} + \tau y + Av_{1xx} \]  
\[ -u_{1x} \]  
\[ u_{2t} = fu_2 - gh (h_1 + h_2 + D) x + g'H h_1 x + Au_{2xx} \]  
\[ v_{2xt} = -\beta v_2 - fu_{2x} + Av_{2xxx} \]  
\[ h_{2t} = -u_{2x} \]

where subscripts indicate partial derivatives. The notation is standard and is explained in the List of Symbols. Note, however, that \( u \) and \( v \) denote vertically integrated transports.

The model geometry is illustrated in Fig. 2. The point

![Figure 2. Model geometry and variables. The \(\beta\)-plane ocean is joined to a nonrotating fjord at \(x = 0\). The ocean basin is 1000 km wide to avoid direct influence of the western boundary.](image)

\( x = 0 \) is the junction between the shelf and the fjord, with the shelf occupying the interval \((-L_o, 0)\) and the fjord occupying \((0, L_f)\). \( L_f \) and \( L_o \) are the lengths of the fjord and ocean, respectively.
The governing equations for the fjord are (1) and (4) with \( f = 0 \), and (3) and (6). Equations (2) and (5) do not apply because there is no \( \nu \) transport in the fjord.

**Matching and Boundary Conditions**

The matching conditions between the two regions are continuity of transport and layer thickness.

The boundary conditions are

\[
\begin{align*}
    u_i(x = -L_o) &= v_i(x = -L_o) = 0 \\
    u_i(x = L_f) &= 0 \\
    v_i(x = L_f) &= 0
\end{align*}
\]  

for \( i = 1, 2 \).

One further boundary condition must be applied to the vorticity equation. We choose to apply the condition at the western oceanic boundary; conditions there are known better than at the eastern boundary. Because the oceanic regime obeys \( \beta \)-plane dynamics, a Munk-type western boundary current is available to balance the \( \nu \) transport due to wind stress curl in the interior.

The transport of the western boundary current is determined by the horizontal gradient of vorticity \( (\nu_{xx}) \) at the western wall \( (x = -L_o) \). For an expanded discussion of this condition, see Hurlburt and Thompson (1973). Rather than adjust the western boundary current to equal the Sverdrup transport exactly, the following approximation is used. The steady Munk boundary layer solution is obtained from

\[
A \hat{\nu}_{XXX} - \beta \hat{\nu} = 0
\]

with the conditions that \( \hat{\nu}(x = -L_o) = 0 \) and \( \hat{\nu}(x \to \infty) \to 0 \). The solution is

\[
\hat{\nu} = C \exp(-K(x+L_o)) \sin(\sqrt{3}K(x+L_o))
\]  

where \( K = \frac{1}{2} \sqrt{\frac{3}{A}} \) and \( C \) is a constant. The value of \( \hat{\nu}_{xx}(x = -L_o) \) and \( \int_{-L_o}^{\infty} \hat{\nu} dx \) are obtained in terms of \( C \) from (8).

The last condition on the two-layer model is that the interior transport across the plane of the model is balanced by
the western boundary transport or
\[ \int_{L_0}^{L_0+\Delta} (v_1 + v_2) \, dx = -\int_{L_0}^{L_0} \nabla \phi \, dx \]
where \( \Delta \) is the thickness of the boundary layer. This relation determines \( C \) which is used to find \( v_{xx}(x = -L_0) \).

The final boundary condition in each layer is
\[ v_i(x = -L_0) = H_i (H_1 + H_2)^{-1} \nabla_{xx}(x = -L_0) \quad (9) \]
for \( i = 1, 2 \). The constant in (9) divides the transport between the two layers.

**Numerical Procedure**

The system of equations (1) - (6) with boundary conditions (7) and (9) is solved numerically using a staggered grid, centered in time finite difference scheme; \( u_i, h_i \) are known at the same time on alternating \( x \) nodes, while \( v_i \) is known at the same nodes as \( u_i \) at intermediate times.

In the \( x \) momentum equations, (1) and (4), the acceleration term uses a trapezoidal time integration, the pressure gradient and viscosity terms are averaged over old and new time steps, and the Coriolis term is evaluated explicitly. This setup is similar to the semi-implicit scheme discussed by O'Brien and Hurlburt (1972).

The continuity equations (3) and (6) are also treated implicitly with the divergence term averaged over new and old time steps.

In the vorticity equations, (2) and (5), forward differences are used for both \( x \) and \( t \), the beta term is averaged over both time and space. The viscous term is evaluated by the appropriate central difference in \( x \) at the old time step. This approximation for the viscous term allows explicit calculation of the \( v \) transports at each time step. The alternative to this approximation requires the solution of a system of forced, linear algebraic equations with a quadri-diagonal coefficient matrix, which greatly increases the required calculations at each time step.

The time differenced equations reduce to two coupled, forced, ordinary differential equations in \( x \) for the \( u \) transport at the new time. The two equations can be uncoupled by forming
new variables.

\[ q_j = (u_1 + \lambda_j u_2) \quad j = 1, 2 \]

For a suitable choice of \( \lambda_j \), the equations uncouple into two equations of the form

\[ q_j - \alpha_j q_{jxx} = F_j \quad j = 1, 2 \]  \hspace{1cm} (10)

where \( \alpha_j \) is a constant and \( F_j \) is the known forcing function composed of variables from the old time step. Note that because the Coriolis term is explicit, its effect is contained solely in the forcing function, \( F_j \), and that the \( x \) momentum equations for the fjord also reduce to (10) with a slightly modified forcing function.

Using a central difference form for the second derivative in (10) reduces the differential equations to a set of forced, linear algebraic equations with a tridiagonal coefficient matrix. Fast, direct methods exist for solving such systems of equations (e.g., Isaacson and Keller, 1966, p. 55 ff).

**BASIC CASE**

The simplicity of the model allows us to test large areas of the parameter space. However, to maintain some order in investigating various dynamical situations, the parameters are chosen to correspond to fjords for which appropriate data are available. The basic case is typical of fjords on the south-western coast of Norway. The parameters are listed in Table I.

| \( \Delta x = 2 \text{ km} \) | \( H_1 = 20 \text{ m} \) |
| \( \Delta t = 1 \text{ hr.} \) | \( H_2 = 480 \text{ m} \) |
| \( f = 10^{-8} \text{ sec}^{-1} \) | \( L_f = 30 \text{ km} \) |
| \( \beta = 10^{-13} \text{ cm}^{-1} \text{ sec}^{-1} \) | \( L_o = 970 \text{ km} \) |
| \( g' = 4.5 \text{ cm sec}^{-2} \) | \( A = 10^5 \text{ cm}^2 \text{ sec}^{-1} \) |

The model, started from rest, is forced by wind stress curl over the main ocean basin. The structure of the assumed
wind field is shown in Fig. 3. The sign of the wind at the eastern boundary determines if there is upwelling or downwelling at the mouth of the fjord.

The model results for downwelling conditions are shown in Fig. 4. Ekman pumping causes the free surface to rise, with the largest elevation occurring at the junction point. At the same time, the pycnocline deepens, driving an outward transport in the lower layer of the fjord. Within one day of the start of the wind, the pycnocline is such that the lower layer pressure gradient is nearly zero. As long as the wind blows, this small lower layer flow balances the incoming, upper layer Ekman flux.

A northward coastal jet forms on the oceanic side of the junction point, which is in geostrophic balance with the surface pressure gradient. After a longer time (∼10 days), a southward oceanic undercurrent also forms. This geostrophic, along-shore current system, with its surface and pycnocline displacements, controls the thickness of the two layers at the fjord-shelf junction and thereby, controls the circulation in the fjord.

This offshore control is seen explicitly if the basic case is forced by the wind for 48 hrs., after which the wind is turned off. An x-t contour plot (Fig. 5a,b) of the free surface and pycnocline anomaly show clearly that the slopes...
Figure 4. Typical fjord simulation. Total depth is 500 m with no bottom topography. Other parameters are given in Table I. $U^Y$ is northward at the eastern boundary, impulsively started at $t = 0$. a) Free surface, b) pycnocline and c,d) alongshore velocity in each layer contoured in x-t plane. Dashed lines indicate negative values. Contour intervals are a) 2 cm, b) 2 m, c) 5 cm/sec, d) .5 cm/sec.

inside the fjord vanish a day and a half after cessation of the wind, but offshore, the slopes of the free surface and pycnocline remain. These offshore slopes prevent the layer thicknesses inside the fjord from returning to their initial values. Fig. 5c shows why the flow does not relax. The geostrophic coastal currents maintain the offshore surface and pycnocline slopes, even in the absence of the wind.
Figure 5. Basic case forced for 48 hours. a) Free surface and b) pycnocline anomaly are contoured in x and time. Dashed lines show negative anomalies. Contour intervals are a) 2 cm, b) 2 m.
Two additional features of this simulation need to be emphasized. First of all, notice in Fig. 4a that the free surface in the fjord continues to rise with time. After ten days of constant winds, water at the head has risen 20 cm from its initial level. This flooding is the result of Ekman transport from the alongshore wind pushing water into the fjord. Such dynamics are not available within the fjord, so the water "piles up" at the junction point until the pressure gradient can transport the water into the fjord.

The second notable result is that the free surface slope is a baroclinic effect. If the flow were barotropic, there would be a tilt of the free surface but not of the pycnocline. Fig. 4b shows that the fjord flow is baroclinic to some extent. For a purely baroclinic flow, the surface and pycnocline slopes are proportional with $\Delta \rho / \rho$ as the proportionality constant. A short calculation shows that the fjord circulation is almost entirely baroclinic. The pycnocline slope at some time is obtained from Fig. 4b (e.g., day six when the slope is $4.5$ m/30 km). The density difference is $4.5 \times 10^{-3}$. Therefore,
for the flow to be purely baroclinic, the free surface slope must be 2 cm/30 km. Fig. 4a shows that this is the model result. Thus, the free surface slope is essentially a reflection of the pycnocline displacement, i.e., it is baroclinic.

The basic case results were calculated for a southward coastal wind stress which gives upwelling at the coast. The model dynamics are the same for either direction of the wind, only the response is reversed. However, the surface layer becomes thinner with time causing very large velocities due to a venturi effect. The pycnocline eventually reaches the free surface and the model breaks down.

The results of this simple calculation are supported by wind and current measurements from a fjord system on the south-western coast of Norway (Fig. 6). Winds were measured off the coast on Utsira Island (W2) and in Josenfjord (W1) and Hylsfjord (W3). Current meter records from station B in Josenfjord indicate the flow patterns. Sea level slope is obtained from tide gauges at Stavanger and Sand. All of these measurements are smoothed with a 24 hour sliding mean.

The close correlation of the surface (1.5 m) fjord currents (Fig. 7c) with the up- and down-fjord winds (Fig. 7a) is evident. There are times, however, when the surface flow is opposite to the surface wind stress, such as June 12, June 20 and June 26, 27. These times correspond to periods of strong, southward coastal wind stress (Fig. 7b). Such southward stress gives offshore Ekman transport of the surface water, thereby pulling water out of the fjord. Support for this calculation is provided by current records at 1.5 m, 5 m, and 10 m (Fig. 7c d, e) which show downfjord currents. There is a small compensating up-fjord flow in the lower layers during these wind events.

The model response to southward wind is obtained from Fig. 4 by changing the sign of the results (because of model linearity). The response is then a falling of the free surface and a rising of the pycnocline. The free surface slopes downward from the head to the mouth of the fjord while the pycnocline slopes downward from the mouth to the head. These
Figure 6. Southwestern coast of Norway. The inset locates various fjord stations discussed in the text.

Pressure gradients drive outward flow in the upper layer of the fjord and inward in the lower layer. This flow pattern is in agreement with the current meter records in Fig. 7. Note that during each of the three events, the circulation in the fjord is opposite to that expected from the direction of the wind in the fjord.

Further evidence of this offshore control of the fjord circulation is shown in Fig. 8. The mean up- and down-fjord current is shown to be strongly affected by the coastal winds. Observe, especially, June 18 and June 27.

Sea level slope (Fig. 9) provides some support for the model results, although not as clear cut as the previous current data. The calculated difference of water level at the two stations indicates a net downward slope of the free
Figure 7. Fjord and coastal winds with current measurements at Station B. Up- and down-fjord winds were measured at Station W1 (Josenfjord) while alongshore coastal winds were measured at W2 (Utsira). The wind and current records are smoothed with a 24 hour sliding mean.

surface from the head to the mouth. Nevertheless, there is some indication of increasing (decreasing) coastal free surface elevation with northward (southward) coastal wind stress.
Figure 8. Mean current profiles from Station B with wind roses from Station W2 on the coast. Currents are 24 hour means evaluated at noon on the day indicated. The wind roses are six hourly coastal wind measurements for the day indicated and the previous two days.
Figure 9. a) Coastal and fjord sea level and winds. Fjord winds are measured at W3 in Hylsfjord and the coastal winds are from W2. Stavanger and Sand are the coastal and fjord sea level stations, respectively. WL curves are 24 hour sliding means of the water level records. The fifth plot in each set is the difference in water level between the coastal and fjord stations.

**ADDITIONAL REMARKS**

**Sills**

Sills are present in most fjords; hence, their effect on the circulation must be considered. We ran the basic case with a parabolic sill 400 m high and 20 km wide, centered at the junction point (Fig. 10). Even though this sill takes up 80% of the total depth, its dynamical effect on the transport in either layer is indiscernible. The velocities were much larger in the vicinity of the sill but this is caused by
thinning of the layers, a venturi effect. Several other cases with sills were calculated; none showed any major effect due to the sill.

60 km Long Fjord

One case was calculated with a fjord length of 60 km but with all other parameters equal to the basic case values. Details of the flow development are clearer in this simulation (Fig. 11). First of all, notice that the pycnocline at the head of the fjord rises during the first day. This arises from the faster set up of the free surface compared to the pycnocline. Therefore, for the first day, both layers flow into the fjord. By day two, the pycnocline slope is sufficient to balance the free surface slope and an outward flow develops.

The upper layer flow takes longer to develop fully as is evident in Fig. 11a. About six days are required for the initial
Figure 10. The basic case with northward winds is calculated with a 400 m high by 20 km wide sill centered at the junction between the fjord and the ocean. a) u velocity (cm/sec) in the upper layer. b) u velocity (cm/sec) in the lower layer. The contour interval in both figures is .1 cm/sec. Dashed contours indicate negative values.

flow to develop. A comparison of Fig. 4b and 11d shows the different times required for the pycnocline to adjust to the wind forcing in the two cases.
Figure 10. (continued). Basic downwelling with a 400 m high sill. c) Free surface anomaly in cm and d) pycnocline anomaly in m. The contour intervals are c) 1 cm and d) 1 m. Dashed contours indicate negative values.

Five-day Periodic Winds

Rarely do winds persist for many days, so the effect of time varying wind forcing must be considered. As a first approximation, the winds are given a sinusoidal temporal variation with a five day
Figure 10 (continued). Basic downwelling with a 400 m high sill. e) Upper layer and f) lower layer alongshore velocity in cm/sec. The contour intervals are e) 2 cm/sec and f) .5 cm/sec. Dashed contours indicate negative values.

period. This choice approximates the synoptic variation of the winds

The response is indicated in x-t contour plots (Fig. 12). As before, the largest anomaly in the free surface and interface occurs
Figure 11. 60 km long fjord forced by northward coastal winds. All other parameters are the same as the basic case. The alongshore velocities are similar to Figs. 4c and 4d and will not be shown. a) upper layer u and b) lower layer u in cm/sec. The contour intervals are a) .5 cm/sec and b) .1 cm/sec. Dashed contours indicate negative values.

at the junction of the two regimes. The asymmetry of the response as the wind reverses is due to the time required to reverse the circulation pattern. During the first 2½ days, the circulation
Figure 11 (continued) 60 km long fjord. c) Free surface anomaly in cm and d) pycnocline anomaly in m. The contour intervals are c) 1 cm and d) 1 m. Dashed contours indicate negative values.
Figure 12. Synoptic wind simulation. a) Free surface and b) pycnocline anomaly for a 5-day period sinusoidal wind stress. Physical parameters are the same as the basic case. Dashed lines indicate negative values. Contour intervals are a) 2 cm, b) 2 m.
spins up from rest. During the second 2½ days, the wind blows against the prevailing circulation which must be stopped before being reversed. The response is always smaller during the second half period than during the first. Circulation near the fjord has a short enough response time that it keeps up with variations in the wind.

**Sea Breeze**

The effects of sea breeze forcing on the model is also considered. The up- and down-fjord wind is purely sinusoidal with a 24 hr. period and an amplitude of .4 dynes cm$^{-2}$. It is applied from the head of the fjord to 20 km offshore of the junction point. There is no alongshore wind.

This simulation (Fig. 13) again shows the strong control that the geostrophic coastal currents have over the fjord circulation. After two days, the major surface and internal slopes are set up and are balanced by the coastal currents. The major sea breeze effect in the fjord is within 2 km of the head. Here the layer thicknesses oscillate slightly with a 24 hr. period. A similar small oscillation is found on the seaward side of the strong coastal current.

The fjord responds differently to up- and down-fjord winds than it does to alongshore coastal winds. The free surface is set up such that there is no net change in the volume of water in the fjord. The Ekman transport for across-shore winds is along the coast, so there is no transport to fill or empty the fjord. Note also that the free surface is baroclinic as it was for alongshore wind forcing.

The largest onshore-offshore currents in the upper and lower layers are 3 cm/sec and .1 cm/sec, respectively. The surface current reverses in phase with the wind, while the lower layer current is out of phase with the wind.

This model result is contrary to the findings of Svendsen and Thompson (1978, Fig. 10). Their averages for the diurnal variation in Josenfjord show that only the surface (1.5 m)
Figure 13. Sea breeze simulation. Up and down fjord winds (.4 dynes/cm$^2$) are applied over the fjord and 20 km offshore. Basic case parameters are used. a) upper layer $u$ and b) lower layer $u$, both in cm/sec. The contour intervals are a) 2 cm/sec and b) .1 cm/sec. Dashed contours indicate negative values.
Figure 13 (continued) Sea breeze simulation. c) Free surface anomaly in cm and d) pycnocline anomaly in m. Contour intervals are c) 1 cm and d) 1 m. Dashed contours indicate negative values.
Figure 13 (continued) Sea breeze simulation. e) upper layer and f) lower layer alongshore velocity in cm/sec. Contour intervals are e) 2 cm/sec and f) .5 cm/sec. Dashed contours indicate negative values.
current responds directly to the wind. At 5 m and 10 m, the current is out of phase with the wind and from 20 m on down, the currents are in phase.

These data, in conjunction with the model results, show the dynamical importance of the thin, fresh surface layer which is present in the fjord. Any surface forcing within the fjord acts on this thin surface layer, setting up a pressure gradient. The density contrast between this layer and the next is so large that the layers remain distinct. The pressure gradient in the top layer forces the next layer against the wind stress. Therefore, circulation models of a fjord with a fresh surface layer should include this thin layer in the dynamics.

Strait of Juan de Fuca

One final calculation is presented which considers Strait of Juan de Fuca. The parameters (Table II) for case are quite different from those of a Norwegian fjord

Table II. Parameters for Strait of Juan de Fuca simulation (parameters not given here are the same as Table I).

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
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<tr>
<td>$g'$</td>
<td>1 cm sec$^{-2}$</td>
</tr>
<tr>
<td>$H_1$</td>
<td>30 m</td>
</tr>
<tr>
<td>$H_2$</td>
<td>120 m</td>
</tr>
<tr>
<td>$L_f$</td>
<td>100 km</td>
</tr>
<tr>
<td>$L_o$</td>
<td>900 km</td>
</tr>
</tbody>
</table>

The major difference is a change of aspect ratio. The wind forcing has the structure given by Fig. 3 and is northward over the eastern ocean.

The interface anomaly is presented in Fig. 14. Despite the change in geometry, the results are quite similar to the basic case. The major displacement of the interface occurs at the junction point, again controlled by the geostrophically balanced coastal currents. Note also that the majority of the fjord is only slightly affected by the disturbance at the coast, even after five days of steady forcing.

This result is compared to measurements in Juan de Fuca and offshore by Cannon, et al. (1971) which shows an inclination
Figure 14. Straits of Juan de Fuca simulation. X-t contour of the pycnocline anomaly. The impulsively started wind has the structure given in Fig. 3 and is northward on the eastern boundary. Physical parameters given in Table II Contour interval is 4 m.

of the internal density field of .5 m per km after sustained southerly offshore winds. The model shows an internal inclination of .3 m per km.

No doubt, other dynamical processes are important in Juan de Fuca. The Coriolis acceleration is important because the straits are not narrow compared to the internal deformation radius. Nevertheless, available data indicate that offshore rotational dynamics affect conditions in the strait.

SUMMARY AND CONCLUSIONS

We have constructed a simple, linear transport model for a two-layer fjord connected with a two-layer ocean. The oceanic dynamics are rotational (β-plane) while the fjord is not affected by the Coriolis acceleration. This simple model neglects vertical mixing, nonlinear effects and geometric
effects such as changes in the fjord width. Nevertheless, several interesting and previously unexpected features appear in the model simulations. They point out the need to reconsider some aspects of fjord circulation theory. These ideas can be tested by extending observational programs to include wind, currents and hydrography from nearby coastal areas.

The major conclusions of this research are the following. 1) Alongshore geostrophic coastal currents strongly control circulation in the fjord. This current has the effect of elevating (or depressing) the free surface and pycnocline, thereby controlling the displacement of these surfaces at the fjord mouth. The resulting pressure gradients within the fjord drive the circulation. Previous theories (Long, 1975; Pearson and Winter, 1978; Stigebrandt, 1979, 1980) use topographic conditions at the mouth to control fjord circulation. The interplay of these two control mechanisms will be considered with a nonlinear version of this model.

Field studies in Josenfjord, Norway (Svendsen and Thompson, 1978); Alberni Inlet, British Columbia (Stucchi and Bell, 1980) and the Strait of Juan de Fuca, Washington (Cannon, et al., 1971) support our conclusions that wind driven coastal currents influence fjord circulation. Similar interactions are found between the Chesapeake Bay and coastal waters (Wang and Elliot, 1978) and the Potomac River and Chesapeake Bay (Elliot and Wang, 1978).

2) The free surface response to alongshore and across-shore wind stress is markedly different. Across-shore winds produce free surface slopes without changing the volume of water in the fjord. Alongshore winds, on the other hand, cause large changes in the volume of water in the fjord.

3) The surface tilt in the fjord is mainly a baroclinic effect. As the calculation in Section 3 shows, the surface basically mirrors the pycnocline.

4) Most simulations were made with and without sills. There was never any major effect from sills, due mainly to the linearity of the model. We expect a nonlinear model to respond more strongly to bottom topography.
5) The velocity change between the two layers in the fjord was as much as 10 cm/sec for some cases. Similar shear is obtained from diffusively driven models (Officer, 1976, p. 116 ff) but they require diffusivities of 50 cm²/sec, which is quite large, especially for fjords. These large velocity shears, due to coastal wind effects, will affect turbulence levels and diffusion within the fjord.

Future work with more complicated models and expanded field data sets will refine the various ideas presented here. However, we expect coastal dynamics to be an important factor in fjord circulation physics.

LIST OF SYMBOLS

- horizontal eddy viscosity
- constants used to uncouple the x transport equations
- northward variation of the Coriolis parameter
- bottom topography variation from z = 0
- density difference between the two layers
- Coriolis parameter
- forcing function for uncoupled x transport equations
- acceleration of gravity
- reduced gravity
- initial thickness of the two layers
- variation of thickness from initial values
- length of ocean and fjord, respectively
- internal Rossby radius of deformation
- buoyancy frequency squared
- combinations of u₁, u₂ to uncouple x transport equation
- surface wind stress in the East and North directions, respectively
\begin{align*}
u_1, u_2 & \quad - x \text{ transport in each layer} \\
v_1, v_2 & \quad - y \text{ transport in each layer} \\
x, y, z & \quad - \text{cartesian coordinates. Positive values are East, North and upward, respectively.}
\end{align*}

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