A Three-Dimensional Simulation of Coastal Upwelling off Oregon¹

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ABSTRACT

The wind driven, z-y, two-layer β-plane numerical model developed by Hurlburt (1974) is used to investigate the effects of a bottom topography and coastline configuration, like that off Oregon, on the onset and decay of the ocean upwelling circulation. The digitized nearshore Oregon bathymetry is analyzed for dominant scales, and a smoothed version is used in model cases with several different initial states and wind stresses. Cases with topography are compared to cases with plane sea beds. Topographic variations are found to dominate over coastline irregularities in determining the longshore distribution of upwelling. Results indicate that stronger upwelling observed near Cape Blanco is primarily due to the local bottom topography and not the cape itself. Observed variations in the meridional and zonal flow are attributed to the topographic β-effect. In particular, during spin-up with an equatorward wind stress, a nearshore poleward undercurrent is most likely to develop in regions where topographic beta is positive. Upper layer poleward flow is observed during spin-down. The existence of an onshore transport jet south of Cape Blanco is predicted. Zonal mass balance is not observed. Topographic Rossby waves are excited during spin-up. Baroclinic continental shelf waves are observed in time series of the pycnocline height contours.

1. Introduction

Many investigators have recognized the need to understand the three-dimensional aspects of coastal upwelling, and in particular the effects of longshore variations in bottom topography and coastline on the upwelling flow field. The numerical model developed by Hurlburt (1974) allows inclusion of longshore variation in bottom topography and coastline, but has so far been applied only to fairly simple and highly idealized topography and coastline configurations. The present study is the first attempt, using this model, to study the effects of the actual coastline and bottom topography of a mesoscale coastal upwelling region, namely, that of the Oregon coast. We will examine relationships between important longshore scales of geometry and the associated response of the flow field. Also, two recent field programs, Coastal Upwelling Experiments (CUE) I and II have focused on an area near Newport, Ore. These experiments plus a continuing research effort by Oregon State University have made the Oregon coast one of the world's most intensively studied upwelling areas. Thus, we intend to compare results with this extensive observational knowledge.

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In this work, we are guided by several important theoretical studies in the three-dimensional aspects of upwelling and longshore variability. In particular, Arthur (1965) and Yoshida (1967) have discussed the effects of capes on steady-state flows, both concluding that the strongest upwelling should occur on the south side of capes. Sugino-hara (1974) studied the effects of longshore variability in the wind stress in a model with a straight coast and longshore-independent bottom topography. For a flat bottom case he found that after the winds were shut off the upwelled portion of the pycnocline propagated poleward at the speed of an internal Kelvin wave. Also, for a case with y-independent shelf-slope topography, Sugino-hara's model develops a poleward flow in the lower layer after the winds are shut off and exhibits northward-propagating continental shelf waves. Gill and Clarke (1974) also studied the effects of longshore variations in the equatorward wind stress and found Kelvin wave and continental shelf wave dynamics to be important in understanding the local upwelling intensity. The correlation between movements of the thermocline and changes in sea level at the coast has led them to suggest the possibility of a working procedure for the prediction of upwelling. Hurlburt (1974) showed that the β effect simulated by north-south sloping topography plays a fundamental role.
in the dynamics associated with mesoscale longshore topography variations. Hurlburt found that these variations produce barotropic flows which extend far beyond the topographic feature. Variations in coastline geometry were observed to excite large-amplitude internal Kelvin waves. Effects of a longshore-independent shelf-like topography have been investigated in a three-dimensional, continuously stratified \( f \)-plane model for the steady and time-dependent cases by Pedlosky (1974b, c). Shaffer (1974) found that longshore variations in the topography of the Northwest Africa shelf determined the distribution of onshore flow and upwelling for that area. He proposed a homogeneous \( f \)-plane model with a shelf whose edge varies in the longshore direction (but with no sloping topography) to explain the observed “funneling” of onshore transport.

In the present work several cases are examined in order to study the onset, and the initial stage of relaxation, of coastal upwelling in a hydrodynamic model with Oregon-like topography.

2. The model

The model used in this study was developed by Hurlburt (1974). It is a wind-driven, \( x-y-t \) , two-layer primitive equation model on the \( \beta \)-plane. The model is nonlinear and retains the free surface.

a. Model geometry

Fig. 1 depicts a typical section of the appropriate geometry for application of the model to an eastern ocean coastal upwelling case. The origin of the right-handed Cartesian coordinate system is located at the easternmost extent of the fluid, at a specified latitude and at an arbitrary reference level below or level with the greatest depth of the fluid.

The fluid is contained in a channel of horizontal dimensions \( L_x \) and \( L_y \). The western boundary is a straight solid wall, while the eastern boundary is variable. As discussed below, the northern and southern boundaries are open. As shown in Fig. 1 \( D(x, y) \) is the height of the bottom above an arbitrary reference level. The subscripts 1 and 2 refer to the upper and lower layers respectively throughout this paper.

b. Problem formulation

The ocean is modeled using a two-layer incompressible fluid assumed to be hydrostatic and Boussinesq. The effects of atmospheric pressure gradients and tides are neglected.

Layer densities \( \rho_1 \) and \( \rho_2 \) are constants. Thermodynamics and thermohaline mixing are not included. Implicit in these omissions is the assumption that the time scale for vertical advection will be much less than that for vertical eddy diffusion. This assumption is not very realistic, especially under conditions of strong surface turbulence, tidal mixing or shear instability. There is observational evidence that, in the Oregon coastal upwelling region, periods of shear instability do occur, with Richardson numbers <10 having often been observed (Huyer, 1974). In a numerical model incorporating thermodynamics, Thompson (1974) has shown that vertical mixing may be comparable to vertical advection during the week to 10-day upwelling cycle.

Although the omission of thermodynamics precludes the evolution of a realistic steady state, the present model has much to offer, especially owing to the
transient nature of the upwelling event cycle. For small interface displacements, Thompson (1974) found
good agreement between a purely hydrodynamic model and one with thermodynamics. In all of the cases to
be discussed below the interface is always at least 20 m from the surface, thus reducing inaccuracies
due to neglect of thermodynamics.

Under the above assumptions the model equations are the vertically-integrated primitive equations for a
rotating, stably stratified fluid on a β-plane (Hurlburt, 1974):
\[
\frac{\partial V_1}{\partial t} + V_1 \cdot \nabla V_1 + k \times \nabla V_1 = -g \nabla (h_1 + h_2 + D) + \tau_S - \tau_B \rho_1 h_1 + A \nabla^2 V_1, \tag{1}
\]
\[
\frac{\partial h_1}{\partial t} + \nabla \cdot (h_1 V_1) = 0, \tag{2}
\]
\[
\frac{\partial V_2}{\partial t} + V_2 \cdot \nabla V_2 + k \times \nabla V_2 = -g \nabla (h_1 + h_2 + D) + \tau_S - \tau_B \rho_2 h_2 + A \nabla^2 V_2, \tag{3}
\]
\[
\frac{\partial h_2}{\partial t} + \nabla \cdot (h_2 V_2) = 0, \tag{4}
\]

where 1 denotes the upper layer, 2 the lower layer,
and other terms are as follows:
\[
\begin{align*}
\rho & = \frac{\partial}{\partial x} - \frac{\partial}{\partial y} \\
V_i & = u_i i + v_i j \\
f & = f_0 + \beta (y - y_0) \\
g' & = g(\rho_2 - \rho_1)/\rho_2 \\
\tau_S & = \tau_S i + \tau_S j \\
\tau_B & = \rho C_B |V_1 - V_2|/(V_1 - V_2) \\
\end{align*}
\tag{5}
\]

The subscript \(i = 1, 2\). Symbols are defined in the
Appendix. Derivation of the layered equations and explanation of the interfacial and bottom stress for-
mulations can be found in O'Brien and Hurlburt (1972).

No-slip conditions apply at the eastern and western boundaries. The quasi-symmetric boundary condition
invented by Hurlburt (1974) is used in the north and
south. In short, the boundary condition, which is based on the \(x \cdot z\) model of Hurlburt and Thompson
(1973), sets \(\partial / \partial y = 0\) in (1)-(4) except for the north-

south (N-S) pressure gradient in the momentum

equations and \(\partial h_1 / \partial y\) in the continuity equations.
The N-S pressure gradients are given by
\[
g' \frac{\partial}{\partial y} (h_1 + h_2 + D) = \int_{-L_2}^{L_2} \beta (v_1 - v_2) dx \\
+ g' \frac{\partial}{\partial y} (h_1 + h_2 + D) \big|_{-L_2}^{L_2}, \tag{6}
\]
\[
g' \frac{\partial}{\partial y} (h_1 + h_2 + D) - g' \frac{\partial h_1}{\partial y} \\
= \int_{-L_2}^{L_2} \beta \left( v_2 + \frac{h_1}{h_2} \right) dx \\
+ g' \frac{\partial}{\partial y} (h_1 + h_2 + D) \big|_{-L_2}^{L_2} - g' \frac{\partial h_1}{\partial y} \big|_{-L_2}^{L_2}, \tag{7}
\]

where
\[
v_4 = v_1 - \left[ v_2 h_1 + \left( v_2 + \frac{h_1}{h_2} \right) \frac{\partial h_1}{\partial x} \right] / (h_1 + h_2). \tag{8}
\]

Hurlburt and Thompson (1973) show that the N-S pressure gradient responds primarily to the geostrophic
N-S flow. Use of (8) removes ageostrophic components from \(v\) in (6) and (7).

This open-basin condition is particularly suited to
our purposes. We are interested in modeling the up-
welling circulation of the Oregon shelf region, a region of
approximately 500 km N-S extent. To model such a
region adequately with a closed basin, it would be
necessary for the basin's horizontal dimensions to be
O(1000 km) for, as shown by Hurlburt and Thompson
(1973), a Sverdrup interior will not develop in a closed
basin of lesser dimensions. A Sverdrup interior is,
however, one of the primary dynamical features of the
upwelling circulation and must be preserved. Use of the quasi-symmetric boundary conditions allows us,
with relative economy, to solve the problem in an open
basin of mesoscale N-S extent and still develop a
Sverdrup interior in the model.

Initial conditions for the model cases to be de-
scribed in Section 4 are either (i) rest or (ii) the quasi-
balanced initial state prescribed by Hurlburt (1974):
\[
u_1 = \frac{H_2}{H_1 + H_2} \left( \frac{\tau_S y_1}{\rho H_1 f} \right) (1 - e^{-r/\lambda_1}), \tag{9}
\]
\[
u_2 = \frac{H_1}{H_2} u_1, \tag{10}
\]
\[
\tau_1 = \frac{1}{\rho \beta (H_1 + H_2)} \text{curl}_S \tau_S, \tag{11}
\]
\[
H_1 = \text{constant}, \tag{12}
\]
\[
H_2 = (H_2 + D) \big|_{-L_2}^{L_2} - \frac{f}{\beta g (H_1 + H_2)} (\tau_S y_1 - \tau_S y_2) \big|_{-L_2}^{L_2}. \tag{13}
\]
where $r$ is the distance from the boundary and $\lambda_r$ is the baroclinic radius of deformation, i.e.,

$$\lambda_r = \left[ \frac{g' h_1 h_2}{f^2 (h_1 + h_2)} \right]^{\frac{1}{2}}. \quad (14)$$

The quasi-balanced initialization prescribes a Sverdrup interior and filters out the inertial oscillations, Rossby waves and gravity waves caused by impulsive application of the wind stress at $t=0$, allowing an easier interpretation of the low-frequency dynamics. Eqs. (1)–(4) with the specified boundary and initial conditions and parameters (as explained in Sections 3 and 4) close the problem. The nonlinear nature of the model equations, and the arbitrary bottom topography and coastline require a numerical approach.

c. Numerical scheme

The numerical model devised by Hurlburt (1974) employs the efficient semi-implicit scheme described by O'Brien and Hurlburt (1972) in the $x$ direction; an explicit scheme is used in the $y$ direction. Diffusive terms are treated implicitly in the $x$ direction using the Crank-Nicholson (1947) scheme. Other frictional terms are lagged in time. Leapfrog time differencing is used for Coriolis and nonlinear terms. Adveective terms are approximated using Scheme F from Grammelvedt (1969). The model uses a variable resolution grid: discrete variations in the $x$ direction and an analytically stretched variable in the $y$ direction. In the present work, a constant $\Delta y = 10$ km was used. Our $\Delta x$ is a discrete function of $x$, equal to 4 km in the easternmost 72 km of the grid. Moving westward, $\Delta x$ changes to 8, 20, 50, 100 and finally 300 km in the western part of the basin. This grid was used for all model cases discussed herein.

The grid resolution in the eastern ocean must be fine enough to resolve boundary layer phenomena and the topographic forcing ($\Delta x < 5$ km). A coarse grid has been used in the western ocean since we are not interested in the solution there. Hurlburt and Thompson (1973) and Hurlburt (1974) have demonstrated that for a simple equatorward wind stress in the upwelling area, the solution for the eastern ocean is independent of that for the western boundary region provided the zonal extent of the model is at least 1000 km.

Since we use an explicit scheme in the $y$ direction, the most severe constraint on the permissible time step is given by the Courant-Friedrichs-Lewy (CFL) linear stability condition

$$\Delta t \leq \min \Delta y / [ g \cdot \max (H_1 + H_2)]^2. \quad (15)$$

Fine resolution of the longshore scales, i.e., small $\Delta y$, and a realistically deep basin, i.e., large $\max (H_1 + H_2)$, are both desired. Thus (15) forces a compromise between economy and fine resolution of longshore scales and puts a practical limit on the total basin depth.

3. Bottom topography and coastline

The necessary bathymetry data $D(x, y)$ were digitized from the 1968 Coast and Geodetic Survey charts 1308N-17 and 1308N-22 and from a chart compiled by Dr. John V. Byrne from unpublished Coast and Geodetic Survey soundings (personal communication). The data were digitized on a 2 km grid of 500 km N–S extent and 120 km E–W extent. The northern and southern limits of the data are approximately 46°30'N and 42°N. The coastline was constrained to be a single-valued function of latitude. The chart data are bottom depths (m) below mean lower low water. Points falling in areas of depth $\geq 1000$ m were arbitrarily assigned the value 1000 m. During the digitization phase we felt that this depth was a reasonable upper limit on basin depth for the purposes of our study. The error for depths $< 1000$ m is estimated to be, at most, $\pm 5\%$ (Peffley, 1974).

Bathymetric charts such as Fig. 2 reveal a complex topography. The most prominent coastline feature is Cape Blanco, north of which is an essentially meridional coast. Underwater topographic features include the small width-scale O(10 km) Columbia River canyon ($y = 225$ in Fig. 2), a much larger scale canyon (axis at $y = 120$), and a ridge system comprised of Stonewall and Heceta Banks ($y = 40$ to $y = -40$).

Shelf widths vary from $\sim 15$ km off Cape Sebastian ($y = -212$) to $\sim 70$ km for the area just south of the Columbia River canyon and the area of Heceta Bank. The magnitudes of the shelf slope and continental slope show considerable longshore variability. Smallest shelf slopes are in the Heceta Bank region, $O(\times 10^{-3})$; largest values $[O(1.5 \times 10^{-3})]$ occur on that portion of the shelf between Cape Blanco and Cape Sebastian. Continental-slope values are smallest in the area off Cascade Head ($y = 90$) where they are $O(1.5 \times 10^{-3})$, compared with order of magnitude greater slopes found seaward of the shelf edge off Heceta Bank.

To determine the important scales of variability in the bottom topography, a Fourier analysis of the zonal rows of the digitized bottom heights was done. It showed that for practically every row, at least 90% of the variance in bottom heights is in scales $\geq 8$ km. A Fourier analysis of zonal position of the isobaths and coastline as a function of latitude showed that at least 90% of the variance is in scales $\geq 20$ km.

In designing the numerical model grid we attempted to use resolution fine enough to resolve both the forcing functions and the dependent variable fields in the important coastal region. Theory indicates that the important onshore-offshore scales for the upwelling circulation are the shelf width $L_s$ and the baroclinic radius of deformation $\lambda_r$. Therefore, we needed to use a $\Delta x$ at least small enough to resolve the features
we get $\lambda_I = 9$ km. We have used $\Delta x = 4$ km within the 70 km nearest the coast, an area including almost all the sloping topography.

Longshore scales of the upwelling and current fields are determined primarily by the scales of the wind and the geometry. In recent work Pedlosky (1974a) has analyzed the role of the longshore structure of the wind stress in coastal upwelling. He concluded: "The long-shore currents are dominated by the largest scales of forcing, while the onshore and upwelling flow is sensitive to all scales of forcing with their structure a strong function of scale." Since we use a $y$-independent wind stress, our choice of $\Delta y$ is based on a meridional Fourier analysis of the coastline position and the bottom heights. Since at least 90% of the longshore variability is in scales $\geq 20$ km, we chose $\Delta y = 10$ km for the model cases discussed below.

A smooth representation of the coastline (Fig. 3), ignoring features with scales $\leq 20$ km, was used. The 500 km region of Fig. 2 has been extended by adding two regions of $y$-independent coastline, 150 km in the north and 150 km in the south. The zonal position of the coast in the northern region is a simple repetition of its position at $y = 250$ km, and likewise that in the southern region is a repetition of the coastal position at $y = -250$ km. The topography between $y = 250$ and $y = -250$ in Fig. 3a was chosen as every fifth row of the bathymetry data shown in Fig. 2. $Y$-independent extensions of the isobaths into the north and south regions were made in a manner exactly similar to that described for the coastline. These regions of $y$-independent coastline and topography are used to reduce longshore derivatives of the velocity field near the northern and southern boundaries. The open-basin boundary condition is based on the assumption that these derivatives will be small.

The steep slopes present in the actual bathymetry demand finer resolution than was economically feasible in this study. Thus we chose to smooth the topography of Fig. 3a. The following filter was used: (i) a 9-point smoother designed to eliminate identically $2\Delta x$ and $2\Delta y$ variation was applied four times to the bottom height field; and (ii) a Hanning filter $[X_j = 0.5X_j + 0.25(X_{j-1} + X_{j+1})]$ was applied 12 times to the meridional rows of the bottom height field.

If we compare the zonal position of the coastline in Fig. 3b to that of the actual coast, we find an absolute error with mean of 1.3 km and standard deviation of 1.2 km. Comparing the bottom topography heights of Fig. 3b with those of Fig. 2 shows an absolute error of mean 10.3 m and standard deviation of 23.7 m. The range of local zonal bottom height gradients for the smoothed topography is from zero to $1.9 \times 10^{-2}$. Local meridional height gradients range from $-3 \times 10^{-1}$ to $3 \times 10^{-4}$. Although quantitative aspects of the bottom and coastline configuration have been distorted, we have attempted to use a

Fig. 2. Bathymetric chart of the Oregon coastal region contoured by computer from the 2 km grid digitized bottom topography data. Depths are in meters. Isobaths deeper than 400 m are not shown. Ordinate labels correspond directly to those of the $x$-$y$ plots in subsequent sections. Although a right-handed coordinate system is used, positive abscissa values are used in this and other $x$-$y$ plots for simplicity. References in the text to positions in $x$-$y$ plots imply kilometers.

g' = 2 \times 10^{-2}$ m s$^{-2}$, \quad f = 10^{-4}$ s$^{-1}$,

$h_1 = 50$ m, \quad h_2 = 150$ m,
smoothing scheme which preserves both the qualitative aspects and the most important quantitative scales.

4. Case descriptions

We will now define the specific model parameter space used for the Oregon cases, and describe the structure of three separate cases, the results of which are discussed in Sections 5 and 6. The parameter values (common to all cases to be discussed) are given in Table 1.

Since our interest is in modeling the circulation on the shelf and not abyssal currents, we have chosen to use a maximum basin depth of 400 m. This assumption is not a severe one, for observations (Mooers et al., 1976) indicate that the upwelling circulation is confined to the upper few hundred meters off the Oregon coast. Also, Thompson (1974) has found no significant qualitative differences in comparing the results of upwelling models with basin depths O(200 m) to those from model runs using much deeper basins.

In this study, we model the permanent pycnocline with the layer interface, and model upwelling by displacements of that interface from its initial position. If the convention of using the 25.5–26.0 σt band to specify the permanent pycnocline is used, observations show that σt surface to be at a depth of 50–100 m off Oregon in the early summer. Since the interface is not allowed to intersect the bottom, we have chosen the initial depth of the interface to be 50 m in order to include more of the shelf topography. As indicated in Fig. 1, the topography is everywhere at least 5 m below the initial interface position. A density difference of 2 kg m⁻³ is used to approximate that between the upper wind-mixed layer and lower layer. This value agrees with typical Oregon observations, e.g., Huyer (1974).

O'Brien and Hurlburt (1972) and Hurlburt and

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Symbol</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>E–W basin extent</td>
<td>L_e</td>
<td>5000 km</td>
</tr>
<tr>
<td>N–S basin extent</td>
<td>L_n</td>
<td>800 km</td>
</tr>
<tr>
<td>Initial upper layer thickness</td>
<td>h_1</td>
<td>50 m</td>
</tr>
<tr>
<td>Initial lower layer thickness</td>
<td>h_2</td>
<td>variable</td>
</tr>
<tr>
<td>Maximum total basin depth</td>
<td>max(h_1+h_2)</td>
<td>400 m</td>
</tr>
<tr>
<td>Coriolis parameter at y=0 (latitude 44°15')</td>
<td>f_s</td>
<td>1.02×10⁻⁴ s⁻¹</td>
</tr>
<tr>
<td>dβ/du</td>
<td>β</td>
<td>2×10⁻¹⁰ m⁻¹ s⁻¹</td>
</tr>
<tr>
<td>Gravitational acceleration</td>
<td>g</td>
<td>9.80 m s⁻²</td>
</tr>
<tr>
<td>Reduced gravity</td>
<td>g'</td>
<td>2×10⁻⁴ m s⁻²</td>
</tr>
<tr>
<td>Horizontal eddy viscosity</td>
<td>A</td>
<td>5×10⁶ m² s⁻¹</td>
</tr>
<tr>
<td>Upper layer density</td>
<td>ρ₁</td>
<td>1.000×10³ kg m⁻³</td>
</tr>
<tr>
<td>Lower layer density</td>
<td>ρ₂</td>
<td>1.002×10³ kg m⁻³</td>
</tr>
<tr>
<td>Interfacial frictional drag</td>
<td>C₁</td>
<td>10⁻⁶</td>
</tr>
<tr>
<td>Bottom frictional drag coefficient</td>
<td>C₂</td>
<td>10⁻⁶</td>
</tr>
<tr>
<td>Time step</td>
<td>Δt</td>
<td>120 s</td>
</tr>
<tr>
<td>Grid increment in x direction</td>
<td>Δx</td>
<td>variable</td>
</tr>
<tr>
<td>Grid increment in y direction</td>
<td>Δy</td>
<td>10 km</td>
</tr>
</tbody>
</table>
described above except that the wind forcing was "shut off" at $t=2.5$ days and the integration continued for another 2.5 days with zero wind stress. The wind forcing of the Oregon upwelling regime is highly variable and even changes direction during the upwelling season. Therefore, we feel that a study of the onset of upwelling and initial stages of decay is important to understanding the Oregon case.

5. Spin-up

a. Introduction

As a review, we will summarize the dynamics of coastal upwelling for a $\beta$-plane case with equatorward wind stress, a meridional coastline, and a flat bottom. The longshore flow is nearly geostrophic. A Sverdrup balance exists in the interior. Offshore flow in the upper layer is Ekman drift reduced by a longshore pressure gradient; the onshore flow $u_z$ in the lower layer is geostrophic. Near the coast, the zonal flow must go to zero to satisfy the boundary condition, and geostrophy breaks down resulting in an equatorward jet in the upper layer and a poleward jet in the lower layer.

The longshore pressure gradient reduces the barotropic mode near the coast and permits the poleward undercurrent to develop. The cause of the longshore pressure gradient has been a subject of controversy. Hurlburt and Thompson (1973) considered an $x-z$ model which neglected $y$ derivatives except those of the pressure gradient terms and the Coriolis parameter. They derived the result

$$\int_{-L_z}^{x} \frac{\partial}{\partial y} (k_1 + h_2 + D) \, dx = \frac{g}{\beta} \frac{\partial}{\partial y} \left[ \frac{\partial}{\partial x} (h_1 + h_2 + D) \right]_{-L_z}^{x},$$

which shows how a N-S sea surface slope can be induced by the $\beta$-effect.

Longshore variations in bottom topography can also contribute to the N-S pressure gradient through the topographic $\beta(\beta_T)$ effect. How $\beta_T$ becomes important can be illustrated by considering a barotropic vorticity equation (Hurlburt, 1974):

$$\frac{\partial}{\partial t} \left( \frac{\partial \tau_S}{\partial x} \right) + \frac{\partial}{\partial y} \left( \frac{\partial \tau_S}{\partial y} \right) = -f \left( \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right) - \frac{1}{\rho} \text{curl} \left( \frac{\tau_S - \tau_H}{h} \right),$$

where the advective terms and diffusive terms have been neglected. Substituting from the continuity
equation

$$\left( \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right) = -\frac{1}{h} \left( \frac{\partial h}{\partial t} + u \frac{\partial h}{\partial x} + v \frac{\partial h}{\partial y} \right)$$  \hspace{1cm} (18)$$

and defining

$$\beta_T = -\frac{1}{h} \frac{\partial h}{\partial y}$$  \hspace{1cm} (19)$$

yields

$$\frac{\partial}{\partial t} \left( \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y} \right) - \frac{f}{h} \frac{\partial h}{\partial x} + \frac{1}{\rho} \frac{\partial}{\partial x} \left( \frac{\tau_x - \tau_B}{h} \right)$$

$$+ (\beta_T + \beta) v = - \frac{\partial}{\partial y} \left( \frac{\tau_x - \tau_B}{h} \right).$$  \hspace{1cm} (20)$$

Thus the N–S sloping topography yields an analogous term to the planetary vorticity advection in the integral of Eq. (16), and can therefore exert an effect on the longshore pressure gradient. Hurlburt found the $\beta_T$ effect to be dynamically important even for mesoscale features of the order of 100 km.

b. Upwelling

Longshore variability in the upwelling pattern for the case with topography is in sharp contrast to the flat bottom case of Fig. 5. As an aid to understanding, the reader is encouraged to use the attached transparency of Fig. 3b as an overlay in interpreting Fig. 5b and other figures portraying topography case solutions. The striking conclusion from comparing Figs. 5a and 5b is the reduction in upwelling at the shore everywhere except at the canyon ($y=120$) and at Cape Blanco. Generally weaker upwelling near the coast in the topography case is due to upwelling being induced further offshore by E–W bottom slope. The layer interface of Case III exhibits vertical velocities at the coast of $\sim6 \times 10^{-6}$ m s$^{-1}$ for the cape region and $4 \times 10^{-5}$ m s$^{-1}$ for the CUE study area ($y=100$). Halpern (1974a), estimating vertical velocities in the CUE area during two periods of southward winds, found values of $6.6 \times 10^{-5}$ and $1.25 \times 10^{-4}$ m s$^{-1}$. Downwelling is seen in areas west of the zero contour.

Favored areas of upwelling in Fig. 5b are the cape and the head of the mesoscale canyon (axis at $y=120$). An interesting comparison can be made with Fig. 1 of Smith et al. (1971), a synoptic mapping of sea surface temperatures for a large part of the Oregon coast for a typical day of the coastal upwelling season. It shows pronounced upwelling south of Cape Blanco and another area of relatively strong upwelling between 44$^\circ$40' and 45$^\circ$20' N ($y=50$ to 120). Holladay and O'Brien (1975) in their Fig. 8 show coldest sea surface temperatures south of the mesoscale canyon. They also conclude that the mean isotherms tended to parallel isobaths. The pattern of upwelling (Fig. 5b) exhibits significant departures from the simple picture of upwelling parallelizing isobaths.

Hurlburt (1974) studied effects of a symmetric mesoscale (half-width $\sim$100 km) canyon on the upwelling pattern. He found (i) decreased upwelling north of the canyon axis, (ii) increased upwelling south of the axis, and (iii) enhanced upwelling on the axis near the shore. For a symmetric ridge, he found effects opposite to those of the canyon. In both cases, the dominant longshore scale appeared to be that of the topographic disturbance. By using the transparency overlay on Fig. 5b, these basic results are
seen to be consistent with the observed pattern of upwelling.

The relative importance of coastline and topography variation is dramatically shown in Fig. 6, a comparison of y-t plots of upwelling at the coast for the flat and topography cases. Figs. 5b and 6b show that Cape Blanco is an area of maximum upwelling. But Fig. 6a clearly demonstrates that the coastline variation, i.e., the cape itself, is not the cause! We conclude that the relatively narrow shelf in the cape region does not diminish the upwelling in the offshore direction as happens along other parts of the shelf, thus forcing the upwelling to occur intensely in the region nearest the one-sided divergence. Fig. 6b also shows the main canyon to be a region of strong upwelling, with decreased upwelling associated with the adjoining ridges. Satellite photos (Meneely and Merritt, 1973) and airborne radiometer measurements (Elliot and O'Brien, 1976) indicate that the region of Cascade Head (y=85) is an area of relatively strong upwelling during the first days of southward winds, and Fig. 6b is in agreement.

c. Free surface

Fig. 7 compares the free surface anomaly for the flat bottom case to that of a case with topography. Except for the deflection due to coastline irregularities, Fig. 7a closely resembles the free surface for a $\beta$-plane case with flat bottom and a meridional coast-

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Fig. 6. Contours of pycnocline height anomaly (m) at the coast as a function of time and N-S position for Case I (a) and Case II (b).

Fig. 7. Free surface anomaly ($m \times 10^{-5}$) at day 5 for Case I (a) and Case III (b).
line. Fig. 7b, however, indicates that topography exerts a strong influence on the free surface. A significantly greater depression of sea level occurs in the case with topography, part of which (≈0.20 m in Case III) can be accounted for by lack of mass conservation in the numerical solution (0.05% of the original mass is lost by day 5). In the flat case mass loss amounts to approximately 0.0035 m of sea level change. Also, at t = 0 the sea surface in Case III is already depressed ≈0.13 m due to the quasi-balanced initialization. The remaining difference is attributed to a stronger barotropic mode induced by the topography. Ridging of the free surface near x = 115 is due to the topographic Rossby wave.

During the period 23–27 August 1972, a period of strong southward wind, a barometrically adjusted sea level decrease of 0.188 m was recorded at Depoe Bay, Ore. (Smith, 1974). Since steric effects are not significant on the time scale of days, the decrease is due to hydrodynamic effects and provides a base of comparison for the present study.

A N–S sea surface slope in the upwelling region (sloping downward toward the pole) is evident in both cases of Fig. 7, although it is more prominent in Fig. 7a. For the flat case, the slope is approximately −0.25 × 10^−2 m (100 km)^−1, in good agreement with theory (Hurhurt and Thompson, 1973) which predicts the N–S slope for a flat-bottomed ocean to be τ_{sg}/[ρg(h_1 + h_2)]. The sea surface slope in the upwelling zone of the topography case is more complex, and is not well delineated near shore in Fig. 7b, but is approximately −0.2 × 10^−2 for the outer shelf.

As discussed earlier, N–S sloping topography can introduce a term in (16) analogous to β and thus can affect the N–S sea surface slope. However, it is the zonal integral of β_1 that is important, a basin width of O(1000 km) being necessary for the β effect to realistically affect the longshore pressure gradient. Values of β_1 for the topography of Fig. 3b are large enough however (values of β_1 of ≈1 × 10^−2 m^−1 s^−1 are typical) to make the β_T effect important even on the mesoscale. Several regions of the actual Oregon bottom topography have β_1 values O(10^−7 to 10^−8 m^−1 s^−1). Thus, the topographic β-effect is probably quite important in Oregon coastal upwelling circulation.

d. Upper layer flow

Close quantitative similarity exists in the contours of the zonally component of upper layer velocity for

The terms "zonally" and "onshore-offshore" are used interchangeably herein to mean motion in the x or E–W direction. Flow perpendicular to isobaths is referred to as "cross-isobath" flow. The terms "meridional" and "longshore" are used interchangeably to mean motion in the y or N–S direction. Flow parallel to isobaths is referred to as "along-isobath" flow. Huyer et al. (1975) found this convention for longshore flow to simplify the analysis of observational data without altering the qualitative results in any way.

the flat bottom and topography cases indicating that topography has little effect on u_1. For an upper layer thickness of 50 m, Ekman dynamics predicts u_1 = τ_{sg}/(ρv hf) ≈ −0.02 m s^−1. Zonal velocities for the topography case are in good agreement with this. Halpern (1976) reported offshore velocities at CSE-II Station B to be 0.15 m s^−1 in the surface Ekman layer which is apparently restricted to the upper 15 m. Thus the model predicts transports in good agreement with the observations, but the artificially thick upper layer of the model causes a decrease in zonal velocity. The longshore dependence of u_1 appears to be dominated by coastline variations, and their associated length scales.

Contours of v_1 (Fig. 8) are essentially similar to the flat bottom case (not shown) but exhibit a stronger poleward flow west of x = 120 in response to the barotropic mode of Fig. 7b. The equatorward surface jet in the topography case has somewhat greater longshore variability, and the jet maximum is slightly further offshore and stronger than in the flat bottom case. The existence of the jet is well documented (e.g., Mooers et al., 1976; Stevenson et al., 1974; Huyer et al., 1975) with equatorward speeds 0.20–0.30 m s^−1 in the jet maximum located 15–20 km offshore. Theory predicts an e-folding width of λ_1 for the equatorward jet. The offshore side of the jet in Fig. 8 has an e-folding width of 12–14 km, in good agreement with the observations of Huyer (1974).

Huyer (1974) compared geostrophic calculations from hydrographic observations to v-component time series from current meters off Newport. She concluded that "... the observed currents were mainly geostrophic during the period 5 to 20 July [1972] even though the winds were variable." Comparing Fig. 8...
with Figs. 5b and 7b demonstrates the geostrophic nature of the baroclinic jet. The stronger barotropic mode of Case III produces somewhat larger equatorward speeds than in the flat bottom case.

Similarity in the longshore flow between the flat and topography cases does not, in this case, imply a lack of influence of topography on the longshore flow. In the topography case, a weaker E–W baroclinic pressure gradient will cause a weaker baroclinic equatorward flow. However, this effect is somewhat masked in the topography case by the stronger barotropic mode. Comparison of Fig. 8 with the upwelling pattern of Fig. 5b shows how areas of stronger equatorward flow coincide with favored areas of upwelling.

One aspect of the upper layer flow, although not depicted here, is noteworthy. Huyer (1974), in comparing currents at station DB-7 (x=18, y=70) to those on the line off Yaquina Head (y=50), has observed that the longshore current seems to diverge as it moves onto the widening shelf. The model solutions also show a divergence in the upper layer velocity vectors in the same vicinity.

e. Lower layer flow

Lower layer zonal flow in the flat bottom case is weak (<0.01 m s⁻¹) everywhere. In contrast, Fig. 9 clearly demonstrates that topography has a strong effect on the offshore and longshore structure of \( u₂ \). In particular, note the region of strong onshore flow of \( \sim 0.03 \) m s⁻¹ just south of the cape. Although this feature of the flow appears to be associated with the coastline variation, the flat bottom case exhibits no such feature. In fact, flow south of the cape in Case I is directed offshore! In the topography case onshore flow is relatively strong, between 0.01 and 0.02 m s⁻¹, nearshore for most of the N–S extent of the basin. There is a region of strong offshore flow with the axis at \( y \approx 50 \).

Zonal flow in the lower layer can be understood in terms of the linear \( \gamma \)-momentum equation (Hurlburt and Thompson, 1973)

\[
\frac{\partial u₂}{\partial t} + f u₂ = -g \frac{\partial}{\partial y}(h₁ + h₂ + D) + g \frac{\partial h₁}{\partial y} \frac{\tau_{by}}{\rho h₂},
\]

In the interior a geostrophic balance holds. The kinematic boundary condition forces \( u₂ \) to zero in the upwelling region and, for the flat bottom case, \( \partial u₂/\partial y \geq 0 \). For the case with E–W sloping topography, offshore transport over the shelf is approximately the same as in the interior. Thus, mass continuity requires the onshore flow over the shelf to become supergeostrophic, forcing \( \partial u₂/\partial t < 0 \) until a frictional balance is achieved in (21).

The generally stronger onshore flow near shore noted above can thus be understood as a consequence of mass continuity and the rising topography. If the position of the onshore maximum south of the cape in Fig. 9b is compared to that of the net \( x \) transport in Fig. 11b (which closely resembles the lower layer zonal transport), the following can be noted. The velocity maximum is inshore of the transport maximum, demonstrating how \( u₂ \) becomes supergeostrophic meeting the continuity constraint.

Fig. 9 demonstrates that topographic variations affect the pattern of onshore flow well beyond the region of sloping topography, producing barotropic flows which seem to be governed by the barotropic radius of deformation.

Longshore variations in deep onshore flow have been observed (Shafer, 1974) in the Northwest Africa upwelling region. For canyons of approximately half the width of those in the Oregon area, Shafer observed neutrally buoyant floats to move shoreward along the canyon axis. He concluded: “The main ‘onshore’ compensation flow takes place along the axis of the canyon.” This is not observed in Fig. 9. Instead, a pattern of offshore transport north of a ridge, and onshore transport south of a ridge is seen, with the exception of the ridge at \( y \approx -150 \). This pattern agrees with the results of Hurlburt (1974) for a simple symmetric ridge. The Northwest Africa observations and the area of onshore flow north of the ridge at \( y \approx -150 \) in Fig. 9 lead us to speculate that the longshore distribution of deep zonal flow is quite sensitive to the longshore scale of the topographic variation.

Fig. 10 presents the lower layer velocity for the topography case at day 5. For the flat bottom case (not shown) meridional flow at day 5 is everywhere weakly poleward in the lower layer, with a jet structure developing in the upwelling zone similar to that
seen by Hurlburt (1974) for a β-plane flat bottom, meridional coast case. But the picture in Fig. 10a is quite different. The equatorward jet structure of the y-independent boundary regions implies the dominance of a strong barotropic mode there. In the region of longshore varying topography, the barotropic mode is much reduced; in fact, two regions of poleward flow develop. Poleward undercurrent, coincident with the upwelling season, has been observed in Oregon (Mooers et al., 1976; Huyer, 1974).

In the upwelling boundary layer the E–W barotropic and baroclinic pressure gradients compete in driving the geostrophic longshore flow [Eq. (4)]. Comparison of Fig. 10a with the upwelling contours of Fig. 5b demonstrates this tendency for areas of poleward (or weaker equatorward) flow to coincide with areas of stronger upwelling, i.e., an increased baroclinic mode.

Hurlburt and Thompson (1973) have reviewed theoretical explanations of the poleward undercurrent and explained how the β effect can produce a longshore pressure gradient sufficient to reduce the barotropic mode, allowing poleward flow to develop. The poleward undercurrent observed in the present study is similar to that which developed for sharp shelf cases in the x–z models of Hurlburt and Thompson (1973) and Thompson (1974), being restricted to the region immediately adjacent to the coast. As explained in the discussion of Eq. (21), \( \nu_z \) becomes supergeostrophic over an E–W sloping bottom in response to requirements of continuity, forcing \( \partial \nu_z / \partial z < 0 \). Over the flat topography, a moderately increasing N–S pressure gradient in the lower layer coupled with a decrease in other terms of the y-momentum equation in the viscous boundary layer can allow the N–S pressure gradient to dominate very near the coast and \( \nu_y > 0 \).

Hurlburt (1974) has investigated the effects of N–S sloping topography on the development of the poleward undercurrent. Two cases, using wedge-shaped topography with N–S slopes of opposite sign showed the following. Topography sloping upward toward the north (\( \beta_T > 0 \)) resulted in an enhanced poleward flow near the shore. The opposite was true for \( \beta_T < 0 \). As previously discussed, the topographic beta effect can act in (16) to affect the longshore pressure gradient—and thus affect the strength of the poleward undercurrent. Overlaying Fig. 10a with the bottom topography, we see general agreement with the predictions of the \( \beta_T \) effect. Also, the undercurrent is more likely to develop in areas where the zonal width of the sloping topography is greater. The equatorward barotropic mode is more efficiently reduced by \( \int \beta y \nu_z \, dx \) in these areas.

The vectors of Fig. 10b generally follow the isobaths, but regions of cross-isobath flow do occur. Oregon observations (Huyer, 1974) indicate that the major axis of flow is oriented along the local bottom contours. However, mean cross-isobath flow has been observed at all levels at the DB-7 current meter string (\( x = 18, \ y = 70 \)), and the model results indicate cross-isobath flow there. An anticyclonic gyre (axis at \( y = 100 \)) with scale governed by the barotropic radius seems to be associated with the main canyon. A weaker gyre exists at \( x = 130, \ y = 240 \). Their positions remain stationary during spin-up.

f. Net transports

Two-dimensional (x–z) models of the Oregon coast predict mass conservation in an E–W plane, but the
observations do not agree; the x-y model shows why. In contrast to the flat case (Fig. 11a), the case with N-S varying bottom topography (Fig. 11b) displays a drastic departure from mass balance in the x-z plane. The following patterns is always seen in Fig. 11b: offshore net transport north of a ridge and onshore net transport south of a ridge. These variations in the zonal barotropic transport respond to fluctuations in the longshore flow caused by the $\beta_r$ effect. Fig. 11 conclusively shows for the Oregon region that a three-dimensional representation of the current field is essential to explain the observed features.

Net N-S transport contours and transport vectors for Case III at day 5 are very similar to those of the lower layer (Fig. 10). However, there is no net poleward mass transport nearshore; and only one region of net onshore transport exists. This intense barotropic onshore transport just south of the cape is clearly seen to be a consequence of a divergence in the net N-S flow. Barotropic vectors generally follow the bottom contours within the 70 km nearest the coast independent of time during spin-up. Features of the flow further offshore, however, develop with time. Notice that the center of the gyre (axis at $y=100$) corresponds to the closed contour of free surface anomaly in Fig. 7b.

All cases exhibit strong barotropic Rossby waves excited by variations in depth and bottom stress as in the right side of (20). Such Rossby waves are evidenced by poleward flow in the western portion of Fig. 10. In this $\beta$-plane model the topographic Rossby waves and consequent westward intensification appear primarily responsible for the return flow of the topographically-induced barotropic coastal current in the eastern ocean (Hurlburt, personal communication).

6. Spin-down

To investigate the relaxation phase of upwelling, the winds have been varied in the simplest manner possible. In Cases I and II, the wind stress of Fig. 4 has been shut off at day 2.5 and the integration continued to day 5. The emphasis of our analysis will be on the pattern of decay of the longshore current and the upwelling, using comparisons of the picture at day 2.5 to that at subsequent times.

a. Zonal flow

The model solutions show greater upper layer onshore transport in areas where the pycnocline is relaxing faster, as would be expected. The area of strong onshore upper layer flow south of the cape observed during spin-up persists during spin-down, suggesting a testable hypothesis: the existence of an onshore jet during the upwelling season just south of Cape Blanco.

In the lower layer at day 5, both the flat bottom and topography cases are characterized by weak $u_2$ flow, $\sim 0.01$ m s$^{-1}$ compared to $\sim 0.03$ m s$^{-1}$ at day 2.5. The pattern of longshore variability has reversed, now being onshore flow north of ridges, and offshore flow south of ridges (the N-S lower layer flow has also reversed, now being generally poleward).

b. Meridional flow

The meridional component of upper layer flow for Case I, days 2.5 and 4.25, is shown in Fig. 12. By day 4.25 the equatorward jet has decayed, and its maximum has shifted offshore. By comparison, in the flat bottom case (not shown), the upper layer flow at
day 4.25 is equatorward everywhere and the jet has decayed uniformly longshore. During spin-down the area of stronger southward flow associated with the head of the main canyon diminishes and moves northward. The reason for this northward movement is discussed in Section 6c. Of particular interest are two areas of upper layer poleward flow nearshore in Fig. 12b. Huyer (1974) has reported shallow poleward flow at the NH-3 site (x=10, y=45), which coincided with a barotropic reversal in the currents. Although the E–W sea surface slope in the upwelling zone is still negative at day 4.25, the areas of poleward flow in Fig. 12b coincide with two areas of relatively small E–W slope. We believe inertial oscillations are the dominant mechanism acting here, for at day 5 one of the regions is again showing equatorward, but relatively weak equatorward flow.

The lower layer poleward flow develops during spin-down. At day 2.5 a barotropic southward jet exists in the northern and southern regions, with two areas of poleward flow near the coast. During spin-down, these two areas develop until, from day 3.75 on, there is poleward flow nearshore everywhere from the cape northward. This agrees with observations (Smith, 1974) which indicate the poleward undercurrent over the shelf occurs more frequently after upwelling events relax. A time series of $\psi_{2}$ shows that the equatorward jet in the northern boundary region, with axis at $x=20$ at day 2.5, shifts offshore as it weakens. At day 5 its axis is at $x=40$. During this time a poleward jet has developed inshore of the equatorward jet.

During spin-up the poleward undercurrent is found only in certain areas favored by topography. A $y-t$ plot of free surface height at the coast (not shown) reveals that the barotropic mode begins to relax immediately when the wind is shut off at day 2.5. This, plus slower relaxation of the baroclinic mode, produces the deep poleward flow.

Persistent equatorward flow south of the cape can be understood in terms of Fig. 10. Since the undercurrent is a baroclinic flow, a sufficiently strong barotropic mode can mask the poleward current. Fig. 10a shows that the area south of the cape is such an area. This interesting contrast in the flow direction north and south of the cape leads us to predict the following. If two current meters were moored at the 100 m depth during upwelling season, one north and one south of Cape Blanco, the southern current meter would show a mean flow equatorward relative to that north of the cape.

c. Upwelling

In Fig. 13, the pattern of pycnocline displacement at day 2.5 is compared to that at day 5 for the topography case. The upwelling relaxes slightly during spin-down, but as noted by Huyer (1974) "... the effects of upwelling are easier to create than to destroy...[they] do not disappear during several day periods when the wind is unfavorable to upwelling."

No strong internal waves are noticeable in the flat bottom case during spin-down. However, northward propagation of the upwelling disturbance is quite apparent in the case with topography. The area of 12 m positive upwelling south of the cape in Fig. 13a propagates northward, and at day 5 is located just north of the cape (Fig. 13b). Also, the area of relatively strong nearshore upwelling at the head of the main canyon in Fig. 13a has propagated northward explaining the northward movement of the region of

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**Fig. 12. N-S component of upper layer velocity in m s$^{-1} \times 10^4$ for Case II at day 2.5 (a) and at day 4.25 (b). The wind was shut off at day 2.5.**
relatively strong equatorward flow in Fig. 12. This poleward propagation of the disturbance is interpreted as a baroclinic continental shelf wave. This is not an internal Kelvin wave since it does not occur for the flat bottom case (calculated but not illustrated).

An estimate of the phase speed for the wave observed in Fig. 13 is 0.25 m s$^{-1}$. By comparison, the speed of an internal gravity wave (at the coast) for $h_1 = 10$ m is $\sim 0.40$ m s$^{-1}$. Observations of baroclinic waves for Oregon are sparse, although barotropic continental shelf waves have been observed (e.g., Mooers and Smith, 1968; Cutchin and Smith, 1973).

For cases including both stratification and shelf-like bottom topography, Gill and Clarke (1974) find the free modes have features of both shelf waves and internal Kelvin waves. Allen (1975), in an analytic two-layer model with $\gamma$-independent shelf-slope topography, has found that offshore depth variation leads to a coupling of the baroclinic internal Kelvin waves and barotropic shelf waves. The strength of the coupling depends on a coupling parameter $\lambda$, where $\lambda = \lambda_l / \delta_B$ and $\delta_B$ is a length scale typical of the offshore bottom topography profile. Allen has investigated the problem for weak coupling, i.e., $\lambda < 1$.

The mean $\lambda$ for the sloping topography of Fig. 3b is 0.26. Within 25 km of the coast, a typical $\lambda$ value is 0.4. At Cape Blanco, $\lambda = 1.1$ and the largest amplitude barotropic response is excited (Fig. 9).

An interesting feature of the solutions (visible in Fig. 6) is the apparent longshore variability in shelf wave propagation during spin-up. Although no northward propagation of the upwelling is noticeable during spin-up south of $y = 0$, the pattern of upwelling near the canyon is obviously propagating northward even under forced conditions, at a speed of $\sim 0.25$ m s$^{-1}$.

When the winds are shut off, northward propagation is observed along the entire N–S extent of the basin.

7. Summary

The nonlinear numerical model developed by Hurlburt (1974) has been used to investigate the effects of Oregon-like bottom topography on the onset and decay of the coastal upwelling circulation. The model is a wind-driven, $x$-$y$-$t$, two-layer $\beta$-plane model which neglects thermodynamics. The actual nearshore Oregon bathymetry was digitized and a Fourier analysis done to determine the dominant longshore scales of variation. Using a smoothed version of the topography, and a simple $\gamma$-independent, time-independent wind stress, the model was run with physical parameters characteristic to the Oregon region.

Cases with varying initial states, topographies and wind stresses were compared using synoptic contours and vector plots of model solutions. Variations in bottom topography are found to play a very important role in explaining observed mesoscale features of the Oregon upwelling circulation. The region near Cape Kiwanda ($y = 110$ in Fig. 2) is shown to be an area of favored upwelling due to presence of an underwater mesoscale canyon. In general, for an equatorward wind stress, the observed pattern of longshore variation in upwelling intensity shows greater upwelling equatorward of (and lesser upwelling poleward of) a canyon axis. Topographic variations are found to dominate over coastline irregularities in determining the longshore structure of upwelling. Specifically, our study indicates that the stronger upwelling observed near Cape Blanco is mainly due to local bottom topography rather than the cape itself.
Topography is seen to exert little influence on the pattern of upper layer offshore transport, but does influence the longshore structure of upper layer meridional flow by affecting the baroclinic mode. A great degree of variability in the N–S lower layer flow, including two regions with a nearshore poleward undercurrent, is explained by variations in the longshore pressure gradient induced by the topographic \( \beta \) effect. Topography-induced variations in the long-shore flow excite irregularities in the barotropic onshore-offshore flow on scales exceeding that of the sloping topography. Zonal mass balance is not seen for the Oregon upwelling regime, but net zonal transport is always seen to follow the pattern of an offshore transport north of a ridge and onshore transport south of a ridge under forcing by southward blowing wind.

Topographic Rossby waves are seen in the model solutions and their importance in the return flow of the topographically induced barotropic coastal current of the eastern ocean is discussed. The existence of a relative onshore jet just south of Cape Blanco and relative northward deep flow north of Cape Blanco is predicted. During spin-down (and for some regions during spin-up) baroclinic continental shelf waves are observed in the pattern of the pycnocline height anomaly.

We conclude that the inclusion of realistic bottom topography is essential to an understanding of local upwelling and the longshore current structure of the Oregon region.

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APPENDIX

List of Symbols

\[
\begin{align*}
A & \quad \text{horizontal eddy viscosity} \\
C_l, C_B & \quad \text{drag coefficients for interfacial and bottom friction} \\
D(x,y) & \quad \text{height of the bottom topography above a reference level} \\
f & \quad \text{Coriolis parameter} \\
f_0 & \quad \text{Coriolis parameter at } y = 0 \\
g & \quad \text{acceleration due to gravity} \\
g' & \quad \text{reduced gravity } [-g(\rho_2 - \rho_1)/\rho_2] \\
h & \quad \text{total depth} \\
h_1, h_2 & \quad \text{instantaneous local thickness of the layers} \\
H_1, H_2 & \quad \text{initial thickness of the layers} \\
L_x, L_y & \quad \text{width of the viscous boundary layer in the } y\text{-momentum equation} \\
L_z & \quad \text{width of the continental shelf} \\
L_{xy}, L_{yz} & \quad \text{total extent of the model region in the } x \text{ and } y \text{ directions} \\
t & \quad \text{time} \\
u, v & \quad \text{barotropic components of velocity in the } x \text{ and } y \text{ directions} \\
u_1, u_2 & \quad \text{x-directed components of current velocity} \\
u_1, v_2 & \quad \text{y-directed components of current velocity} \\
x, y, z & \quad \text{tangent plane Cartesian coordinates: } \\
x & \quad \text{positive eastward, } y \text{ positive northward, } z \text{ positive upward} \\
\beta & \quad \text{value of } \beta \text{ simulated by N–S sloping topography} \\
\beta_T & \quad \text{time increment in the numerical integration} \\
\Delta t & \quad \text{horizontal grid increments in the } x \text{ and } y \text{ directions} \\
\lambda_1 & \quad \text{baroclinic radius of deformation} \\
\lambda_E & \quad \text{barotropic radius of deformation} \\
\rho, \rho_1, \rho_2 & \quad \text{densities of sea water} \\
\sigma & \quad (\rho_{s,1,0} - 1) \times 10^5, \text{ where } \rho_{s,1,0} \text{ is the sea water density corrected to atmospheric pressure} \\
\tau_{sx}, \tau_{tx}, \tau_{px} & \quad x\text{-directed tangential stresses at the surface, interface, and bottom} \\
\tau_{sy}, \tau_{ty}, \tau_{py} & \quad y\text{-directed tangential stresses at the surface, interface, and bottom.}
\end{align*}
\]

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