LABORATORY-NUMERICAL MODEL COMPARISONS OF CANYON FLOWS: A PARAMETER STUDY.

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ABSTRACT

An integrated set of laboratory and numerical model experiments has been conducted to understand the development of residual circulation surrounding a coastal canyon and, further, to explore the degree to which laboratory experiments can provide useful benchmark datasets for numerical models of the coastal ocean. The physical system considered here is the interaction of an oscillatory, along-slope background current with an isolated canyon incised in an otherwise uniform continental slope. The flows considered are laminar and are investigated in a cylindrical test cell mounted on a 1.8 m diameter turntable at Arizona State University. The numerical experiments employ the Spectral Element Ocean Model (SEOM) developed at Rutgers University.

The use of an idealized shear stress boundary condition along the coastal floor in the numerical model gives good quantitative agreement with the laboratory results for the zeroth-order, time-dependent flow and good qualitative agreement for the higher order \([i.e., O(Ro)]\), where \(Ro\), the Rossby number, is small\] time-mean flow. The quantitative agreement for the latter, however, is not within estimates of laboratory uncertainties. It is shown that this owes to the shear stress boundary condition used along the coastal floor. The use of a no-slip condition along the floor improves the model away from the canyon boundaries, but the enhanced viscosities needed to obtain numerical stability give boundary layers that are too wide along the coastline. The numerical model results also suggest that the free surface boundary condition in the laboratory was more than likely not wind stress free, and that the fluid above the shelf for certain experiments, owing to surface cooling by evaporation, did not have a purely linear density profile.

The laboratory experiments and their numerical model counterparts have been examined to investigate the trends of a number of flow diagnostics with changes in some of the governing parameters, the others being held fixed. While the quantitative
agreement of the higher-order flow, as described above, is not exact, the flow tendencies with the various parameter changes are shown to be similar for the two models. A scaling analysis suggests that for the range of parameters considered here there are two natural vertical length scales, one the fluid depth and another a more narrow boundary-layer-like thickness [O(RoBu^{-1/2})], where Bu is the Burger number] centered at the level of the shelf break. A scaling argument for the normalized time-mean horizontal flow at the shelf break level gives 
\[ \frac{\bar{U}_1}{u_0} \sim \left\{ Ro \left( \frac{h_s}{h_D} \right)^{-1} Ro_t^{-1} Bu^{-1/2} E^{-1/2} \right\}, \]
where \( h_s / h_D \) is the ratio of the depth of the shelf to that of the deep ocean, Ro is the temporal Rossby number, and E is the Ekman number. Laboratory and numerical data are shown to support this scaling.

Overall, the utility of a closely coupled laboratory-numerical model study is confirmed with the clear indication that such efforts are mutually supportive. The difficulty of obtaining more precise quantitative agreement in this controlled setting suggests just how much more difficult it will be to produce accurate simulations of the coastal ocean where turbulent mixing, fine-scale bottom roughness, and other complicating dynamical factors are prevalent.

1. Introduction

The shelf break is a transition zone between the shallow shelf and the deep ocean, and their respective property fields. As coastal flows are predominantly along-isobath, the cross-shelf transports of sediments, nutrients and physical tracers are difficult to measure. Consequently, the processes responsible for these transports are difficult to deduce from direct observations.

Compounding this problem is the inherent difficulty in modeling the flow over and around steep topography in the presence of background stratification, rotation, and turbulent mixing (itself poorly understood). In situ measurements having sufficient spatial and temporal resolution to serve as datasets for the development and testing of numerical models are few. The goals of the present research include a better understanding of the effect of sea floor topography -- and in particular submarine canyons -- on coastal flows, and a demonstration of the utility of laboratory datasets for the testing and improvement of coastal ocean models.

Submarine canyons are known to be critical factors in determining the nature of coastal current systems. A background on our current understanding of the effects of canyons from an observational standpoint can be obtained by referring to the papers of Hickey (1995,1997); similarly, the papers of Klinck (1996) and She and Klinck (2000) are good introductions to the theoretical and numerical aspects of the problem.

Hickey’s (1997) study of Astoria canyon reported on datasets obtained from an 18-element moored array and concurrent CTD surveys. She was able to characterize spatial patterns of the three-dimensional velocity and temperature fields and to deduce vorticity patterns resulting from upwelling- and downwelling-favorable winds.
Nevertheless, as pointed out in Haidvogel (2003; this issue), the Astoria data set by itself is insufficient to completely define the important elements of the three-dimensional circulation and dynamics. She and Klinck (2000) apply a hydrostatic primitive equation model to a smoothed version of Astoria Canyon. For conditions of steady upwelling- and downwelling–favorable winds, their results are consistent with the general nature of Hickey’s (1997) observations. Nevertheless, quantitative replication of Hickey’s Astoria dataset using coastal ocean models has thus far proven elusive.

In a manuscript now in press, Allen et al. (2003) consider an upwelling event in both a laboratory and numerical model (the S-Coordinate Rutgers University Model – SCRUM). They find poor agreement between the laboratory and numerical models for the homogeneous case owing to nonhydrostatic effects and, while the agreement is better for the stratified case, the differences are attributed to truncation errors in the terrain-following numerical model.

In a recent paper by the three of us, comparisons of the flow fields obtained by a series of laboratory experiments and a numerical model were made; see Pérenne, Haidvogel and Boyer (2001), henceforth referred to as PHB. The motivations for this study were: (i) the recognition that current numerical models when applied to the same forcing and boundary conditions can provide qualitatively different results when strict one-to-one comparisons are made (see, Haidvogel and Beckmann, 1998); (ii) the difficulty of field programs providing adequate data in space and time to test the models; (iii) the understanding that most of the physical aspects of the coastal ocean can be modeled to some degree in the laboratory; and (iv) the realization that modern data acquisition techniques such as particle tracking (PTV) and particle image velocimetry (PIV) can provide datasets having characteristics similar to those generated by numerical models.

It should be made clear at the outset that there is no substitute for comparing models against oceanic data. The view here is that laboratory experiments can isolate individual physical processes, and can be a cost-effective way of aiding in the early development of numerical models. Laboratory datasets therefore complement datasets obtained under realistic oceanic conditions, and both are valuable for model testing and validation.

Several other of our works offer additional background and perspectives on the problem studied here. Boyer, Zhang and Pérenne (2000), henceforth BZP, report on the theoretical aspects of the physical system being considered, as well as the results from some of the initial experiments. Pérenne et al. (2000) consider impulsively started, upwelling- and downwelling-favorable flows past a submarine canyon; this study depicts clearly some of the fundamental differences between upwelling and downwelling flows (e.g., an upwelling flow spins down far more slowly than its downwelling counterpart). Finally, a companion article by Haidvogel (2003; this issue) examines the issue of cross-shelf transport in this physical system.
One area of emphasis in the present study is the time-mean or rectified flow generated by an isolated canyon in the presence of periodic forcing. There is a rather extensive literature on the general subject of rectified flows. Zimmerman (1981), in a review article, emphasizes the importance of the proper scaling of the problem and in particular points to the importance of the ratio of the tidal excursion to the natural horizontal length scale (see below) of the topography under consideration. Robinson (1981, 1983) advances a number of physical processes that can lead to rectification. Huthnance (1973) and Loder (1980) discuss one such process that is based on frictional effects. As vortex lines are oscillated across bottom relief, the effects of friction are greater where the depth is smaller. This leads to the generation of a residual relative vorticity that in turn develops into a mean or rectified flow.

The plan for the paper is as follows. We first give a brief description of the physical system, and the laboratory and numerical techniques used (Section 2). We next discuss the quantitative differences between the laboratory and numerical model results for the central case as presented in PHB (Section 3). The sensitivity to changes in the system parameters, and a discussion of the physical processes responsible for the changes, are presented in Section 4. Conclusions are given in Section 5.

2. The Laboratory and Numerical Models

The present study addresses the quantitative differences between the laboratory and numerical experiments in PHB, and considers the system behavior owing to changes in the governing parameters. It is a natural follow-on to PHB, and the reader is thus referred to that paper for details concerning the background for the study and the laboratory and numerical models employed; only a brief synopsis of that study is given here.

The physical system considered is shown schematically in Figure 1. An annular coast (vertical), shelf (horizontal), and continental slope model, incised by a single submarine canyon, is placed in the central portion of a circular test cell; the deep ocean is between the continental slope and the outside wall of the test cell. The test cell is filled with a linearly stratified fluid whilst rotating and the initial background rotation of the tank is set at f/2, where f is the Coriolis parameter.

After the fluid has reached a state of solid body rotation, experiments are initiated by modulating sinusoidally the test cell at an amplitude $\Delta \Omega$ and frequency $\omega_0$ about the background rotation rate f/2. The amplitude $\Delta \Omega$ is chosen so as to obtain the desired amplitude $u_0$ of the oscillatory velocity as defined at the shelf break, while the designation of $\omega_0$ specifies the temporal Rossby number of the background flow. Tables 1 and 2 are expanded versions of those given in PHB and reflect the experiments reported herein. Note that oceanic prototype values are given to indicate that the region of parameter space for the ocean is similar to that being investigated in the laboratory.
The laboratory observations employ particle-tracking velocimetry (PTV) to obtain the horizontal velocity field at selected levels. The particle tracking data are stored on videotapes and then processed using Digimage software (Dalziel, 1992). A level of uncertainty for the laboratory observations is established by conducting a series of experiments for which the same external parameter values are used. It is demonstrated that, based on a normalized kinetic energy difference criterion between the mean of the experiments and each individual realization, measurements of the residual flows are repeatable to within about 5%.

The numerical model used is the Spectral Element Ocean Model (SEOM); see Haidvogel and Beckmann (1999). SEOM solves the hydrostatic primitive equations using a higher-order finite element approximation. The use of the finite element decomposition offers several practical advantages. Here, for example, a spatial finite element grid can be constructed which exactly follows both the sloping bottom and the circular sidewalls of the test cell. This, together with the seventh-order spatial approximation being applied, guarantees an accurate solution to the physical problem. For these experiments, SEOM is configured to exactly match the geometric and physical forcing conditions prevailing in the laboratory. A more detailed description of the SEOM implementation is given in PHB.

It is useful here to note how the viscosity and bottom boundary condition were handled in the numerical analysis. In the vertical direction, a combination of laminar viscosity \( \nu = 0.01 \text{ cm}^2\text{s}^{-1} \) and a bottom stress parameterization to represent the laminar bottom boundary layer are used. The parameterization employed assumes that the shear stress along the boundary is a linear function of the free-stream velocity outside the boundary layer, and is given by:

\[
\tau_0 = -\rho_0 \frac{v_f}{2} u_k = -(0.05).
\] (1.1)

Relation (1.1) is an exact solution for the streamwise shear stress exerted on the fluid by the floor for a laminar flow along a smooth horizontal surface. In the numerical experiments presented in PHB, the horizontal viscosity is enhanced by a factor of 100 above its true (molecular) value. The numerical model is not able to obtain solutions unless enhanced viscosities are used, owing to the finite numerical resolution employed. (The finest horizontal and vertical grid spacings used in the SEOM simulations are of the order of 1-2 mm.)

2. On the Differences Between the PHB Numerical and Laboratory Model Results

While comparisons of the qualitative features of the time-mean or residual flows in the laboratory and numerical models in the PHB study were highly satisfactory, the quantitative differences were too large to be ignored. For example, Table 6 in PHB indicates differences ranging from 33 to 79% in a normalized kinetic energy difference parameter, with the numerical model having the higher energy for all of the cases.
considered (note that the cases referred to here are for the different boundary conditions used in the PHB study – only one set of parameters viz, those of Experiment 01 below, is considered) The numerical runs in PHB set the vertical viscosity at the same value as that of the kinematic viscosity of water and the horizontal viscosity at one hundred times that of water.

Recognizing that the model floor shear stress condition (1.1) used in the numerical simulations is a parameterized boundary condition which depends only on the Coriolis parameter, the flow speed outside the boundary layer, the mean fluid density, and the viscosity (e.g., it does not depend on the background stratification and the slope of the topography), this aspect of the model was considered a strong candidate for contributing to the quantitative model-model differences.

To explore this matter further, the SEOM model was modified to provide for the application of a true no-slip condition along the solid boundaries of the canyon, instead of the parameterized shear stress condition (1.1). As was the case for the stress condition, the use of the no-slip condition also required enhanced viscosities for numerical stability, and again the vertical and horizontal viscosities employed were the same as and one hundred times that of water, respectively. Figures 2a,b,c are the residual vorticity and divergence fields at the shelf break level for (a) the PHB laboratory experiment, (b) a numerical run using the parameterized shear stress boundary condition, and (c) the numerical run for the no-slip condition, respectively.

Comparing the vorticity and divergence plots for the laboratory experiment and the stress law and no-slip numerical calculations, one notes the following: (i) the no-slip calculation better captures the symmetry across the canyon axis of the laboratory experiments, (ii) the no-slip run more accurately predicts the maximum of the vorticity and the horizontal divergence, and (iii) the no-slip model better represents the anticyclonic pattern in the upper left side of the canyon.

The principal difficulty with the no-slip run is the thick boundary layer of anticyclonic vorticity along the canyon walls. This layer owes to the enhanced horizontal viscosity. In non-rotating boundary layer theory (e.g., aerodynamics) the boundary thickness scales as \( \frac{\delta_H}{L} \sim Re^{-1/2} \), where \( Re = \frac{Lu_0}{\nu} \) is the Reynolds number, \( \delta_H \) is the dimensional boundary layer thickness and \( L \) a streamwise length scale. Taking \( L = W = 20 \text{ cm} \), \( u_0 = 1.0 \text{ cms}^{-1} \) and the enhanced viscosity \( \nu = 1.0 \text{ cm}^2\text{s}^{-1} \), one obtains \( \frac{\delta_H}{W} \sim 0.22 \), which is indeed in good accord with the noted boundary layer thickness in the numerical model.

Owing to conservation of mass constraints, these thick boundary layers in the no-slip case force the maxima of the vorticity and divergence further away from the boundaries than would be the case for solutions that might be anticipated should the model yield results using the viscosity of water. Nevertheless the numerical simulations away from the boundaries for the no-slip model agree considerably better with the laboratory experiments than do the parameterized shear stress runs. Unfortunately the use of the no-slip condition requires far more computer time than
does the use of the parameterized shear stress law. The need for an improved parameterization for the laminar Ekman layer for unsteady stratified flows along sloping surfaces is thus clearly evident.

At the smallest spatial resolution employed in PHB, it was not possible to use the SEOM model to reach a stable result for the case in which both the horizontal and vertical viscosities had the value of the kinematic viscosity of water. To test the model sensitivity to changes in the horizontal and vertical viscosities, a number of experiments were run using the shear stress condition and different combinations of these viscosities. Figures 3a,b are the model results for the velocity and vorticity fields of the residual flow at the shelf break for two such cases; in 3a the vertical viscosity is that of water and the horizontal viscosity one hundred times greater, and in 3b both were ten times greater. While the flow patterns are certainly qualitatively similar, one notes substantial quantitative differences in the velocity and vorticity fields. Furthermore, it is not clear which of the numerical runs 3a,b is better in simulating the laboratory experiment. For example, the vorticity field of Figure 3a (3b) is weaker (stronger) than that of Figure 2b. The intent, of course, in introducing eddy viscosities is to stabilize the numerical calculation, with the view that the flow patterns obtained are relatively insensitive to the exact choice of the coefficients. It is concluded that the use of enhanced eddy viscosities must be done with caution since, as exemplified here, such viscosities can change the basic flow patterns significantly.1

In summary, owing to the use of enhanced viscosities and a rather simplified shear stress boundary condition along the model floor, the model-model comparisons for the residual flow in the present study, while demonstrating good qualitative agreement, showed significant quantitative differences. It is emphasized that these comparisons were for the first-order residual velocity field (the oscillatory forcing is considered the zeroth order), and thus are a severe test.

3. A Sensitivity Study of Parameter Changes

We now show how the residual flow fields found for the central case of PHB are altered as selected parameters are varied (holding the remaining ones fixed). This process, in conjunction with a scaling analysis and comparisons with the results of the numerical model, will lead to a better understanding of the physical mechanisms responsible for the observed velocity, density, and pressure fields.

3.1 Scaling Arguments

Boyer, Zhang and Pérenne (2000; i.e., BZP) considered certain aspects of the theory underlying the physical system studied here. The reader is referred to that article

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1 Although, the SEOM model could not precisely model the residual flows found in the laboratory experiments for the present physical system, the laboratory-numerical model comparisons presented below support the fact that the trends, as evidenced by the sensitivity of the residual flow field to changes in various system parameters, are nonetheless in agreement.
for details. It is important, however, to address the underlying assumptions made by BZP because they are important in understanding the scaling arguments for the horizontal components of the residual flow field that follow.

A cylindrical coordinate system \((r, \theta, z)\) is employed, with \(z = 0\) being at the tank center at what would be the level of the free surface were it not for the presence of the coastal region. Further we introduce the perturbation density,

\[
\rho'(r, \theta, z, t) = \rho(r, \theta, z, t) - \rho_b(z),
\]  

(3.1)

where, \(\rho(r, \theta, z, t)\) is the density field,

\[
\rho_b = \rho_0 - \Delta \rho \frac{z}{h_D}
\]  

(3.2)

is the background density distribution before the turntable modulation is initiated, \(\rho_0\) is the density along the free surface, the dependence of \(\rho_b\) on the radial coordinate is ignored, and \(\Delta \rho\) is the density difference from the free surface to the tank floor in the deep water. Similarly, the perturbation pressure is defined as

\[
p'(r, \theta, z, t) = p(r, \theta, z, t) - p_b(z),
\]  

(3.3)

where, \(p(r, \theta, z, t)\) is the pressure field,

\[
p_b(z) = p_{atm} - \rho_0 g z + \frac{1}{2} \Delta \rho g z^2,
\]  

(3.4)

\(p_{atm}\) is the background pressure field and \(g\) is the acceleration due to gravity.

The dimensionless variables are defined as

\[
t^* = \omega_0 t, \quad r^* = \frac{r}{R_s}, \quad \delta r^* = \frac{\delta r}{W}, \quad \theta^* = \frac{\theta}{W}, \quad z^* = \frac{z}{h_D}, \quad v_r^* = \frac{v_r}{u_0}, \quad v_\theta^* = \frac{v_\theta}{u_0}, \quad w^* = \frac{w}{W},
\]  

\[
\rho_b^* = \frac{\rho_b}{\Delta \rho}, \quad \rho^* = \frac{\rho'}{\rho_0}, \quad \Delta \rho = \frac{\Delta \rho}{\rho_0}, \quad p^* = \frac{p'}{p_{atm}}, \quad \rho_{atm} = \frac{\rho_0 f}{u_0} W
\]  

(3.5)

where asterisks designate dimensionless quantities, \(\delta r\) is the characteristic scale of variations in the dependent parameters in the \(r\) direction, \((v_r, v_\theta, w)\) are the Eulerian velocity components and the remaining terms are defined in Table 1.
The dimensionless parameters governing the system are the

Temporal Rossby Number, \( Ro_t = \frac{\omega_0}{f} \),

Rossby Number, \( Ro = \frac{u_0}{fW} \),

Burger Number, \( Bu = \frac{N^2 h_D^2}{f^2 W^2} \),

Ekman Number, \( Ek = \frac{V}{f h_s} \) and

Geometrical, \( h_D, h_S, L \).

The resulting governing equations are given in BZP and are not reproduced here.

To address the question of scaling and, in keeping with the parameter values given in Table 2, we assume \( a \) priori that

\[ \text{Ro} \ll 1, \ Ro_t \sim O(1), \ Bu \sim O(1), \ Ek \ll 1, \ \frac{h_D}{W} \sim O(1), \ \frac{h_S}{h_D} \sim O(1), \ \frac{L}{W} \sim O(1). \quad (3.7) \]

On the basis of (3.7), as in BZP, it is reasonable to assume that the dependent parameters, including the three velocity components and the perturbation density and pressure, can be expressed as power series in the Rossby number. To the lowest order \((i.e., \ Ro^0)\), BZP show that the laboratory flow exhibits a balance between the unsteady acceleration, the Coriolis “force”, the pressure gradient and the periodic forcing which owes to the time-dependent rotation rate of the turntable \((i.e., \ \text{the turntable’s angular acceleration})\). This last term can be interpreted as a time-varying body force. The lowest order flow is periodic, with the time mean of the velocity components and the perturbation density and pressure being zero. The residual or time-mean flow thus appears as a first order term \((i.e., \ \sim Ro^1)\).

It is helpful to define a control volume for the canyon and its surroundings as follows, where one recalls that a control volume is an imaginary surface in space through which a fluid can flow freely. The control volume used here is given schematically in Figure 4 and is composed of two parts. The upper part is the volume between the shelf break level and the surface, and whose boundaries follow the shelf
break as defined by the intersection of the shelf with the canyon; the boundary at the mouth of the canyon is given by the straight line connecting the two extremes of the canyon mouth. The lower part is defined by the canyon walls on the shoreward side and the surface formed by horizontal straight lines connecting the extremes of the canyon along its mouth. The other symbols on the figure will be discussed below.

The preliminary experiments of BZP include time series measurements of the horizontal velocity components at various elevations along vertical traverses at selected horizontal locations within and in the vicinity of the canyon. The data show that the velocity components \( v_r \) and \( v_\theta \): (i) reach an approximate periodic state almost immediately (i.e., roughly within the first oscillation); (ii) scale as the forcing amplitude \( u_0 \) with, as might be expected, \( v_\theta > v_r \); (iii) exhibit harmonics which are most readily apparent in the \( v_r \) component; and (iv) lead to a mean or rectified horizontal motion field whose scale is significantly smaller than \( u_0 \). While the vertical velocity field is not measured directly, it clearly is very weak.

Owing to the nature of the topography and the stratification, consider first the expected magnitude of the vertical velocity component within and in the vicinity of the canyon. A study of the canyon topography leads one to conclude that the maximum vertical velocities would most likely occur in the region neighboring the line of intersection of the shelf and canyon. Along this line fluid parcels advect over the shelf break with the expectation that vortex lines passing into the canyon will be stretched, leading to the generation of cyclonic vorticity. Fluid parcels advecting along the continental slope and then encountering the canyon are able to move along the canyon isobaths and are not strongly forced to move vertically against the buoyancy forces.

The kinetic energy per unit mass of the fluid passing over the rim has a maximum value of \( u_0^2/2 \). If we assume that the maximum depth \( \Delta z \) to which fluid parcels with this kinetic energy can penetrate is the distance over which all of the kinetic energy is transformed into potential energy, \( (N\Delta z)^2 \), one can readily show that

\[
\Delta z^* = \frac{\Delta z}{h_d} \sim \frac{Ro}{(Bu)^{1/2}},
\]  

(3.8)

where the numerical factor has been absorbed; note that this is a maximum value and the actual vertical penetration might be expected to be smaller. In this regard, and owing to the absorbing of some of the numerical coefficients here and below, one might expect to find the coefficient(s) for the scaling of the characteristic rectified flow to be different than unity, as is indeed the case.

Let the characteristic time-mean or residual horizontal flow speed be \( \overline{U_1} \); i.e., \( \overline{U_1} \sim v_{r1}, v_{\theta1} \). Here \( (v_{r1}, v_{\theta1}) \) are assumed to be of order unity but because of the assumed expansion in Ro, represent an order \( Ro^1 \) term in the expansion. Consider the system behavior after the flow has reached an approximate periodic state. Further,
consider fluid parcels advecting into the canyon beginning with the zero velocity time in the cycle and continuing for one-half a flow cycle. During this time interval, fluid columns, before passing over the shelf break, have weak relative vorticity but gain cyclonic vorticity by vortex stretching as they advect across the shelf break (during the next half cycle, cyclonic vorticity is formed by the same process on the opposite side of the canyon). This vorticity is produced along a line of length \( \sim L \) characterized by the length of the canyon. Owing to stratification effects, it is assumed that the stretching will be limited to the vertical scale given by (3.8). This length is smaller than the drop in elevation of the topography after the column has advected only a short distance into the canyon. The stretching can thus be assumed almost instantaneous as the fluid column passes across the break.

If one then assumes that the potential vorticity for the short columns passing over the shelf is conserved, we can write

\[
\frac{f'}{h_S} \sim \frac{\zeta + f}{h_S + \alpha \frac{Ro}{Bu^{1/2}} h_D}
\]

where \( \alpha \) is a scaling factor and \( \zeta \) is the dimensional vorticity. After simplification (3.9) can be written as

\[
\zeta = \alpha \frac{h_D}{h_S} \frac{Ro}{Bu^{1/2}} f.
\]  

(3.10)

The scaling for the total cyclonic vorticity \( \Omega \) generated within the control volume of Figure 4 for the left-to-right and right-to-left phases of one oscillation cycle, and for the full length of the canyon, can then be written as

\[
\Omega \sim 2L \alpha u_0 \int_0^\pi \sin \omega_0 t dt.
\]  

(3.11)

Using (3.10) and integrating (3.11), one obtains

\[
\Omega \sim \frac{h_D}{h_S} \frac{Ro}{Ro, Bu^{1/2}} u_0 L.
\]  

(3.12)

At the time a periodic state has been obtained there must be a balance between the vorticity produced during one oscillation cycle and that dissipated. The former is given by (3.12) and the latter can be estimated by the dissipation in the boundary layers along the surface where the rectified flow directly encounters the lower bounding surface; see articles on Ekman boundary layers on flat and sloping surfaces in Pedlosky (1979). We take the total area over which this so-called Ekman suction is in effect to be scaled by the product of the length and width of the canyon. Owing to stratification
The total dissipation of vorticity $D$ for a single cycle for the time-mean flow thus scales as

$$D \sim E^{1/2} \frac{\varsigma_B}{W} L,$$

where $\varsigma_B \sim \overline{U}_1 / W$ is the basin-scale vorticity. Equating (3.12) and (3.13) and simplifying, one obtains

$$\frac{\overline{U}_1}{u_0} \sim \frac{Ro}{(h_S / h_D) Ro Bu^{1/2} E^{1/2}} = \lambda,$$

where $\lambda$ is defined as the non-dimensional scaling parameter. In the present experiments the parameters $h_S / h_D$ and $E$ are not varied and thus these aspects of the scaling cannot be tested. Finally it is instructive to solve (3.14) for the dimensional value of the characteristic rectified flow; one obtains

$$\frac{\overline{U}_1}{u_0} = \frac{u_0 f^{3/2}}{\omega N v^{1/2}}.$$

Relation (3.15) nicely demonstrates the influence of the wide range of parameters on the time-mean flow. The laboratory and numerical model results and their comparisons will next be discussed.

### 3.2. The Laboratory and Numerical Model Experiments and the Flow Diagnostics Employed

Laboratory experiments were conducted for five different sets of dimensionless parameters. These are listed in Table 3 and are labeled Experiments 01-05. Experiment 01 is the same as that reported in PHB and is henceforth designated the “central experiment.” Three of the laboratory experiments (02, 03 and 04) have the same geometrical parameters as the central experiment, but each differs in one of the dynamical parameters from the central case. The fifth (05), while having the same dynamical parameters as the central experiment, has a canyon of the same depth but only half as wide; this experiment will be referred to as the “narrow canyon” experiment.

All of the laboratory data reported here were obtained using particle tracking techniques in horizontal planes at four vertical levels. Define the origin of the
coordinate system to be at the free surface directly above the shelf break and centered in the canyon with \( x \) being toward the right facing the deep water, \( y \) being toward the deep water and \( z \) being vertically upward; see Figure 1. The dimensionless heights of the particle tracking measurements were at \( z^* = \frac{z}{h_D} = -0.1 \), the mid-level of the fluid on the shelf; \( z^* = -0.2 \), the level of the shelf break; and \( z^* = -0.3 \), -0.4, levels below the canyon rim. Only three of these were used during each experiment. All of them, however, were investigated in the numerical simulations.

While there are aspects of the numerical model that make it impossible to obtain precise simulations of the laboratory experiments, the numerical model is nevertheless used to compare with the laboratory data with the aim of determining whether the flow trends with changes in the dynamical parameters are in consonance with the changes observed in the laboratory. In addition to simulating four of the five experimental cases (\( i.e., \) except the narrow canyon) a number of other numerical experiments were conducted as listed in Table 3.

With the aim of delineating clearly the observed differences in the various laboratory experiments and numerical simulations, a number of quantifiable and objective measures of the respective flows are defined; see the schematic diagram of the control volume of Figure 4. The following derived quantities that describe aspects of the large-scale motions resulting from the interaction of the background current with the canyon are presented in dimensionless form. Conservation of mass requires that the net mass transport across the control surface of Figure 4 be zero. The relative locations of inflow and outflow, and of upwelling and downwelling, are of interest, however. In the laboratory experiments the flow across the canyon rim was measured only at the \( z^* = -0.1 \) level. The flow into or out of the control volume along the canyon mouth was determined at all of the observation levels.

The following integrals define the normalized volume transport per unit depth into the control volume through the canyon mouth, the flow over the shelf break on the left side of the canyon facing toward the deep water, and the flow over the shelf break on the right side of the canyon, respectively (see Figure 4 where all quantities are positive for flow into the canyon and negative for flow out); \( i.e., \)

\[
Q_1^* = -\frac{1}{u_0W} \int_{-W/2}^{W/2} v(x, y) dx. \quad (3.16)
\]

\[
Q_2^* = -\frac{1}{u_0W} \left\{ \int_{-W/2}^{0} v dx + \int_{y_m}^{y_h} u dy \right\} \quad (3.17)
\]

\[
Q_3^* = -\frac{1}{u_0W} \left\{ \int_{0}^{W/2} v dx + \int_{y_m}^{y_h} u dy \right\} \quad (3.18)
\]

where \( y_m, y_h \) are the \( y \) coordinates of the mouth and the head of the canyon at the calculation level, respectively.
While the vertical velocity cannot be measured directly in the laboratory, it can readily be calculated from the numerical model results. As such $Q_z^*$, the vertical volume flux through the respective levels of the control volume, was calculated at each level for all of the numerical experiments; in particular

$$Q_z^* = \frac{1}{u_0 Wh_D A_{cv}} \int \int w dx dy.$$  (3.19)

While it is not possible to obtain $Q_z^*$ directly in the laboratory, it is possible to determine the horizontal divergence. The area integral of the horizontal divergence at each observation level can be determined as $Q_z^*$ is then

$$D^* = \frac{1}{u_0 W A_{cv}} \int \int (u_x + v_y) dx dy = -\frac{\partial Q_z^*}{\partial z^*},$$  (3.20)

where $A_{cv}$ is the cross-sectional area of the control volume at the elevation in question. It than follows that

$$D^* = -\frac{\partial Q_z^*}{\partial z^*}.$$  (3.21)

Because $Q_z^*$ is zero along the free surface, it is possible in principle to obtain the vertical volume flux through each level of the control volume of Figure 4 by integrating $D^*$ with respect to $z$ from the free surface. In the present experiments, however, the vertical spacing is too large, and, as described above, the vertical velocity gradients too steep to make such integrations.

Because the normalized kinetic energy per unit mass $KE^*$ gives some sense of the strength of the residual current, it is found to be a good measure to test the scaling prediction of relation (3.14). $KE^*$ is defined as

$$KE^* = \frac{u^2 + v^2}{u_0^2},$$  (3.22)

where $(u, v)$ are the velocity components in the $(x, y)$ directions and $u_0$ is the amplitude of the oscillatory flow.

3.3 Results – Overview of Measurements

The measurements (laboratory) and calculations (numerical model) of the normalized horizontal fluxes ($Q_1^*$, $Q_2^*$, $Q_3^*$), the integral of the horizontal divergence ($D^*$, laboratory only), the vertical volume flux ($Q_z^*$, numerical model only) and the
maximum and average kinetic energy per unit mass \((KE_{\text{max}}, KE_{\text{ave}})\) for the horizontal levels depicted in the control volume of Figure 4 are given for the laboratory experiments and the numerical model calculations in Tables 4 and 5, respectively.

Measurements from the laboratory experiments and calculations from the numerical model, except for the superinertial case (Experiment 04), have similar large-scale characteristics. These show that relatively strong upwelling occurs from the upper levels of the canyon to the surface. Both models also show that weak downwelling may be found deep in the canyon. This pattern can be seen qualitatively in the data for \(D^*\). Note that for all of the laboratory runs \(D^*\) is positive at the mid-depth level on the shelf and strongly negative at the shelf break level. Taking the vertical volume flux at the free surface to be zero and assuming no reversals in the sign of \(D^*\) between \(z^* = 0.0\) and \(z^* = -0.1\), this implies an upwelling in the shelf water which then sharply decreases below the shelf break.

It is not possible to determine the distribution of the vertical volume flux with depth from the laboratory experiments because the distance between measurement levels and the vertical gradients in the velocity field are too large for reasons cited earlier. Because the dimensionless distance between the surface and the first measurement level \(z^* = -0.1\) can be considered small and, more importantly, that this level is sufficiently far from the shelf break (where vertical gradients in the velocity field are large), an estimate of \(Q_{z^*}\) in this region can be obtained from (3.15). If one takes the value of \(D^*\) at \(z^* = -0.1\) to be representative of this quantity throughout this upper layer and that the horizontal fluxes \((Q_1^*, Q_2^*, Q_3^*)\) at \(z^* = -0.1\) are approximately the same from this level to the surface, then conservation of mass requires that

\[
(0.1)(Q_1^* + Q_2^* + Q_3^*) = -D^*\left(z^* = -0.1\right) = Q(z^* = -0.1). \tag{3.23}
\]

The \((0.1)\) on the left side represents the distance from the measurement level to the free surface and the \(z^*\) values on the right of the equals sign indicate that the evaluation of these terms should be at \(z^*\), which in the present case is \((-0.1)\). Consideration of the three experiments (01, 03, 04) for which measurements were made above the shelf show that each is characterized by upwelling and, furthermore, that the volume fluxes with regard to relation (3.23) are satisfied within the error limits of the experiments; see Table 4.

A review of the vertical volume flux calculations for the numerical experiments given in Table 5 confirms the upwelling nature of the flow in the vicinity of the shelf break. All of the numerical runs show the strongest upwelling at the shelf break with the large \(Bu = 40\) (Experiment 10), as might be expected, showing the weakest, in fact negligible, upwelling. One notes that these same calculations show weak downwelling \((i.e., Q_{4^*} < 0)\) only at a few deep canyon locations and that the magnitudes relative to those at the higher levels for upwelling are very weak.
One also notes from Table 5 that $Q_1$, the flow into the canyon through its mouth, is always positive at the shelf break and except for Experiment 08 has its maximum value there. The flow is found to be away from the canyon at only the deepest fluid depths for the numerical experiments; $Q_1$ is always positive and has its maximum value at the shelf break. The data also show that the flow on the shelf is always into the canyon from the left, facing the deep water (i.e., $Q_2^* < 0$) and, except for the strongly stratified case, Experiment 10, the flow is also into the canyon on right side (i.e., $Q_3^* < 0$).

Let us now focus on the normalized kinetic energy measurements. In comparing the maximum and average data at the various levels for the control volume for the laboratory (Table 4) and for the numerical model (Table 5), one concludes that a strict quantitative comparison of the laboratory and numerical data will not give satisfactory results because the respective values vary widely. This finding, however, should not be too surprising. As discussed in Section 2, the numerical runs do not perfectly represent the laboratory because of the enhanced viscosity that has to be used to stabilize the numerical computations. Furthermore, the velocity data obtained in the laboratory is not really point data. The sampling volume used in the particle-tracking analysis has a size of $\sim 1$ cm in the horizontal and 3-4 mm in the vertical. Additionally, owing to diffraction effects, the light beam used to illuminate the tracer particles, has a slight curvature over the distance through which the measurements are made; i.e., the thin fluid layer in which measurements are obtained is not perfectly horizontal. It is concluded that the present laboratory techniques do not adequately resolve the vertical structure of the boundary layer-like fluid motion in the vicinity of the shelf break.

As might be expected, a comparison of the normalized average kinetic energy per unit volume ($KE^*_\text{ave}$) data gives more consistent results than do those for $KE^*_\text{max}$. That is, in comparing values of $KE^*_\text{ave}$ between a laboratory experiment and a given run at a given level one finds that the specific values differ widely. On the other hand, if one compares the variation of $KE^*_\text{ave}$ with the normalized vertical coordinate for a given experiment one finds that the qualitative nature of the variation to be similar for both the laboratory and numerical runs. Furthermore, the data also show that the kinetic energy of the mean flow is strongest in the vicinity of the shelf break and tends to fall off quite quickly deep in the canyon and toward the free surface; see Tables 4 and 5.

In order to test the scaling argument leading to relation (3.14), it is hypothesized that an objective measure of a characteristic $\overline{U_1}$ is the square root of the average kinetic energy per unit mass of the horizontal time-mean flow within the canyon at the shelf break level with the outer boundary being a straight line across the mouth of the canyon and the inner boundary being the intersection of the canyon with the shelf break. Figure 5 is a plot of the measured $\overline{U_1}/u_0$ against the scaling parameter, $\lambda \sim \{Ro(h_S/h_B)^{-1}Ro^{-1}Bu^{-1/2}E^{-1/2}\}$ for the laboratory experiments (except for 05) and the numerical simulations (except for 08 where different viscosities were used than for the other runs). The numerical simulations are obtained using the shear stress bottom
boundary condition discussed earlier. The numerical simulations 06 and 07 were conducted at Rossby numbers 0.2 and 0.3, respectively, and had as their principal purpose to investigate the system behavior for varying Ro, the remaining parameters held fixed. Similar experiments were not conducted in the laboratory. The least squares fit to a straight line of the remaining data is plotted on Figure 5; i.e.,

\[
\overline{U_1} = (0.9\lambda + 12.7) \times 10^{-2}. \tag{3.24}
\]

While there is some scatter in the data, both the laboratory results and the numerical simulations support the scaling arguments advanced. Some scatter of the data should be expected because the problem being addressed is inherently 3-D and the measure adopted for determining \( \overline{U_1} \) is at only one plane in the vertical.

In summary, those features of the flow found common in all of the experiments, (i.e., for the range of parameters studied) are that (i) the shelf break level is clearly one of enhanced fluid dynamical activity, (ii) the horizontally integrated residual currents in the shelf water and upper levels of the canyon for all of the experiments are upwelling zones, (iii) deeper in the canyon there is a tendency to have zones of weak downwelling, (iv) the flow from the deep ocean through the canyon mouth is into the canyon for the upper levels of the canyon and for the fluid on the shelf, and (v) on the shelf the residual flow was associated with a transport into the canyon from the deep water through its mouth and away from the canyon for both the left and right sides of the canyon facing the deep water.

4. Results – Case Studies

Figures 6a,b,c are plots of the laboratory and numerical model residual velocity fields for Experiment 01 (the central case of PHB) for the observation levels \( z^* = \{-0.1, -0.2, -0.4\} \), respectively. In these and the like figures to follow, the velocity vectors, and vorticity and horizontal divergence contours, to the extent possible, have the same scale. Those cases with different scales are identified as such.

The large-scale residual flow field for the laboratory experiment for the central case is given by the left-side illustrations of Figures 6a,b,c. For the mid-depth level on the shelf, Figure 6a, one notes a large anticyclonic vortex structure having a canyon-scale horizontal dimension centered near the left side of the canyon near the mouth. This vortex is associated with a significant flow into the canyon across the canyon mouth; see Figure 6a. The rectified flow at the shelf break (Figure 6b) is strong relative to the layer at mid-depth above the shelf and to the flow deeper in the canyon, as exemplified by Figure 6c. This is a ubiquitous finding of the present study. Furthermore, the residual flow is noted to be strongest along the upstream side of the canyon near the canyon head. Below the canyon rim the rectified flow is considerably
weaker but it is organized in such a fashion as to establish a downstream-rectified current that follows the canyon/continental slope isobaths (here and in the following, downstream will be designated as the flow direction with the shoreline on the right). The mechanism establishing this rectified flow deep in the canyon is in consonance with that owing to an oscillatory current up and down the canyon along the canyon isobaths as discussed by Zhang et al. (1994).

Thus for the down-canyon part of the flow cycle, with the boundary on the left facing downstream, the transport in the bottom boundary layer is upward toward the shelf and owing to buoyancy effects and conservation of mass then drifts toward the canyon axis carrying low momentum fluid into the interior. During the up-canyon portion of the cycle the boundary layer flow is toward the canyon floor which draws high momentum fluid into the boundary layer. Averaging over a cycle, the time-mean flow is in the direction with the canyon wall on the right facing downstream; i.e., in the directions along the canyon wall indicated in Figure 6c. A similar argument for the right side of the canyon also holds for Figure 6c with the flow again being along the wall with the boundary on the left.

Comparing the numerical runs with those of the laboratory, one notes quite different qualitative features in the upper layer. While the flow velocities are weak here, the numerical run clearly feels the edges of the canyon more than does its laboratory counterpart; e.g., note the regions of relatively high cyclonic vorticity found adjacent to the canyon walls. These differences may owe to such matters as the enhanced viscosity in the numerical calculations or to surface effects in the laboratory run including wind shear (although the tank was covered with a canopy) or more likely the background stratification not being linear in the upper level owing to evaporation from the free surface and the attendant mixing in that region.

The laboratory-numerical model comparison at the shelf break level is much better; i.e., compare the two illustrations of Figure 6b. Here it is clear that all of the major features of the residual flow such as the strong inflow from the deep water and the strong current on the downstream side of the canyon can be identified on both plots. From Tables 4 and 5 we note that the numerical run is somewhat more energetic than the laboratory experiment, as evidenced by the kinetic energy measures. The qualitative nature of the flow patterns and the kinetic energy level of the residual flow in the observation level below the canyon rim are in reasonable accord as indicated by comparing the illustrations of Figure 6c and the data in Tables 4 and 5.

Figures 7a,b,c are the velocity fields obtained from the laboratory experiments and the numerical model for Experiment 02 in which the dimensionless parameters have all been held fixed with the exception of the Burger number which has been decreased by a factor of four from the central case. It can be shown that this experiment is one for which the only change is that of the buoyancy so that any model results can be construed in terms of a change in buoyancy.
While the qualitative nature of these flow fields at their common levels (i.e., compare Figure 6b with 7a and 6c with 7b, respectively), it is evident that a decrease in buoyancy has substantially enhanced the energetics of the rectified flow; compare the figures or note that the maximum and average kinetic energies are substantially larger at all levels for the lower buoyancy experiment. From the scaling argument given by (3.14), the characteristic mean flow velocity should scale as $Bu^{1/2}$. The laboratory observation is thus in agreement with the scaling.

The qualitative agreement between the laboratory and the numerical runs at the $z^* = -0.4$ level is excellent with all of the major flow features captured by the numerical simulation. As can be seen in Tables 4 and 5, the volume transport per unit depth across the mouth of the canyon is small, but slightly negative (-0.007) for the laboratory experiment and small but positive (0.013) for the numerical run.

Figures 8a,b are the residual velocity and vorticity fields at the shelf break level for the laboratory experiments and the numerical model for Experiments (03, 04), respectively. These experiments have the same parameters as the central case with the exception that 03 has a smaller temporal Rossby number, $R_0t = 0.25$, and 04 has a superinertial $R_0t = 1.25$. Let us first discuss the low $R_0t$ experiment. The qualitative nature of the laboratory results of Figures 6b and 8a are quite similar, suggesting the physical processes at work are similar. One notes from Table 4 that these observations support the scaling argument that $101 \sim U/\omega_0$. 

Figure 8b depicts the velocity and vorticity fields at the shelf break level for the superinertial case. Figures 6b and 8b are the velocity and vorticity fields at the shelf break level for Experiments 01 and 04, respectively. The most apparent difference between the superinertial Experiment 04 and the central case Experiment 01 (and 02 and 03) is that the qualitative nature of the residual flow fields for the laboratory experiment are inherently different; i.e., the superinertial case has a multi-cell structure which differs markedly form the subinertial cases which are dominated by a single cyclonic cell. The multi-cell structure owes to the very short tidal excursions in these experiments. That is, from Table 3, one notes that $X^* = 0.018$ for Experiment 04. Thus on each one-half cycle only small patches of cyclonic vorticity are formed on opposite sides of the upper levels of the canyon. These patches are advected along the canyon walls with the shallow fluid on the right and reach an equilibrium state with the stronger cyclone in the central portion of the canyon and the weaker one downstream of the canyon along the shelf. For larger excursion lengths, a cyclonic circulation dominates the shelf break level of the canyon.

No laboratory experiments were performed for varying Rossby numbers, with the remaining parameters being held fixed. The SEOM model however explored this aspect and the results for the velocity field at two separate Rossby numbers with the remaining parameters being those of the central case are given in Figures 9a,b; see also Experiments 06 and 07 in Table 5. Figures 6b (left), 9a, and 9b qualitatively show the significant increase in the kinetic energy of the residual flow as the Rossby number is increased. The final experiment (05) was conducted using a narrower canyon; in fact
the streamwise scale was made one-half the size used for the other experiments. The various dimensional parameters were adjusted so that all of the dynamical dimensionless parameters were held fixed. Figures 10a,b,c are plots of the velocity and vorticity fields obtained from the laboratory experiments for the \( z^*/h_D = -0.2 \) (shelf break), \(-0.4\) and \(-0.6\) levels. We note first that this experiment should not satisfy the scaling arguments advanced in obtaining (3.14) because the geometrical parameters have been changed and the experiment thus does not satisfy the requirement of geometrical similarity implied by relation (3.10).

A review of the data in Tables 4 and 5 suggests that the narrow canyon case has some features akin to that found for the lower Burger number experiment (02). These include stronger integrated divergence fields and higher average mean kinetic energy levels than the remaining experiments. The vorticity fields show a strong region of cyclonic vorticity at the shelf break level (Figure 10a) which follows the slope of the canyon head and can still be clearly identified at the \( z^* = -0.4 \) level deeper in the canyon. Similarly, regions of anticyclonic and cyclonic vorticity are found near the canyon mouth along the left and right sides of the canyon, respectively.

The narrow canyon case simultaneously demonstrates the two vorticity generating mechanisms discussed in this communication. At the shelf break level the oscillating forcing flow continuously injects cyclonic vorticity into the region, which at large times is balanced by dissipation as discussed earlier. This cyclonic vortex core is seen to be strong at the shelf break level and, while weaker, can be identified as well at the \( z^* = -0.4 \) level. Thus the vortex-stretching mode of mean flow continues to come into play at the shelf break level.

One also notes at the \( z^* = -0.4 \) level that cyclonic (anticyclonic) regions of roughly the same strength are found on the right (left) of the canyon facing the deep water. These vortices are also in evidence at the \( z^* = -0.6 \) level on Figure 10c. At this level deep in the canyon, the vortex stretching mechanism is no longer at work. The vortices are present because on each stroke of the forcing current, cyclonic (anticyclonic) vorticity is advected into the right (left) sides of the canyon mouth; this vorticity is not advected out of the canyon on the return stroke, but rather gives a mean cyclonic (anticyclonic) region which remains near the mouth of the canyon; this vorticity advection into the canyon at equilibrium is balanced by dissipation.

No strong upstream influence of the canyon is noted at any of the levels investigated. The downstream influence, however, is especially strong below the shelf break level as evidenced by the along-slope mean flow that persists several canyon width dimensions downstream of the canyon. The numerical model results are in good accord with the laboratory results on this point.

5. Summary and Concluding Remarks
This study has addressed the question of the degree to which laboratory models might be used in the development of geophysical models and in particular those involving coastal currents and their interaction with bottom topography such as the shelf, continental slope and isolated submarine canyons. The study is a precursor to the study of the effects of turbulence on these flow-topography interactions. The results have been both encouraging and at the same time somewhat worrisome. Encouraging in the sense that the SEOM model was able to obtain good qualitative representations of the laboratory flows and that the combined laboratory-numerical approach provided a desired level of checks and balances on the other. For example the laboratory results encouraged the numerical side of the effort to seek a better understanding of why some of the numerical results did not better simulate aspects of the laboratory measurements and furthermore that the use of such basic numerical tools as the introduction of enhanced viscosities had to be done with care. The numerical results brought into question the nature of the laboratory flows in the surface layer where wind shear stress effects and surface cooling seemed to be distorting the near surface flow.

On the discouraging side was the matter of given the “friendly environment of the laboratory”, the fact that the flows considered were laminar, that the initial state and the fluid forcing and boundary conditions in the laboratory were well defined and understood, that, nevertheless, the agreement between the laboratory and numerical results was not at the level anticipated at the outset. One must ask what this portends for the use of numerical models in the prediction of the ocean environment? In the ocean such physical processes as (i) the nature, distribution and role of turbulence; (ii) the parameterization of smaller scales into the modeling process; and (iii) the lack of extensive in situ data in space and time, are all formidable issues that must be addressed before one can have confidence in any numerical model results offered as predictors of that environment.

The principal conclusions of the present study are:

(a) The time mean flow scaling $\frac{U_1}{u_0} \sim \{Ro(h_s / h_D)^{-1} Ro^{-1} Bu^{-1/2} E^{-1/2}\}$ is well supported by the laboratory and numerical model analysis and has potentially important oceanic implications.

(b) The upper regions of submarine canyons in the presence of unsteady forcing are projected to be upwelling zones for a wide range of parameters. Deeper in the canyon, weak downwelling is predicted.

(c) Lower in the canyon there is a tendency to have some downwelling flow but it is weak compared to the upwelling flow above.

(d) In the upper regions of submarine canyons and extending to the free surface a net flow into the canyon across the mouth and a net flow away from the canyon along its flanks should be anticipated.
Coastal regions having canyons with a sharp shelf break are regions in which the vertical gradient in flow properties such as the rectified current are predicted to be large over a narrow (compared with the fluid depth) vertical distance that scales as $RoBu^{-1/2}h_D$.

Superinertial flows generate residual motions which scale as predicted for subinertial ones. The superinertial forcing does not produce mean flows either far upstream or far downstream of the canyon. Furthermore, residual superinertial flows differ from subinertial ones in being characterized by multiple eddy fields (i.e., not the canyon scale single cyclonic eddy patterns of the latter).

Careful comparisons between the laboratory and numerical models have shown the importance of having a good parameterization of the boundary layer along the coastal floor. While this was shown for the laminar flow considered herein, one finds it difficult to believe that the presence of boundary and interior turbulence will make the problem easier.

The use of enhanced viscosities may cause difficulties in certain regions of the flow field especially along coastlines where large horizontal shears may be expected.

Laboratory experiments can be a source of data for testing numerical ocean models; these experiments are limited to those for which the basic physics is isolated.

The laboratory cannot be a substitute for the ocean in the modeling process for in fact it can only isolate certain physical processes; i.e., the complexity of the ocean cannot be simulated.

Even though the study dealt with an idealized background stratification (i.e., linear from top to bottom of the water column) and that the flow was laminar and the topography was highly simplified, the laboratory and numerical results, while quantitatively different, support each other in suggesting that certain phenomena in the vicinity of submarine canyons should be anticipated.

Acknowledgements

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sabbatical leave during the period January-April 2002 during which time some of the work on this research was being conducted. The kind hospitality and stimulating discussions with Dr. Joel Sommeria, the Director of the Laboratoire des Ecoulements Geophysiques et Industriels (LEGI), and his staff is also appreciated. The authors are indebted to Dr. Sergey Smirnov who prepared the final figures. The efforts of Mr. Chris Robichaud in fabricating the shelf-continental slope-canyon model are also acknowledged.

References


Captions:

Figure 1: Physical system.

Figure 2: Vorticity (left) and horizontal divergence (right) fields for the central experiment discussed by PHB (Experiment 01 in the present study) as obtained from (a) the laboratory, (b) the SEOM model using a parameterized shear stress condition along the model floor and (c) the SEOM model using a no-slip condition, including a highly resolved Ekman layer, along the model floor.

Figure 3: Velocity fields for SEOM model at the shelf break level corresponding to the central experiment of PHB (Experiment 01 – see Figure 6b) for horizontal and vertical viscosities having the respective factors (a) (100, 1), (b) (10,10) and (c) (1,100) times that of water.

Figure 4: Control volume.

Figure 5. Characteristic speed of the normalized time-mean flow $\frac{U_1}{u_0}$ (i.e., defined as the square root of the average kinetic energy per unit mass as found for the area at the shelf break level as sketched in Figure 4a) at the shelf break level as obtained from the lab experiments and the numerical model against the scaling relation $\lambda = \left(\frac{h_D}{h_S}\right)RoRe^{-1/2}Bu^{-1/2}E^{-1/2}$.

Figure 6: Velocity and vorticity fields for laboratory and SEOM models for Experiment 01 at the observation levels $z^* = (a) -0.1$, (b) –0.2 and (c) –0.4.

Figure 8. Velocity and vorticity fields for laboratory (left) and SEOM (right) models for the level at the shelf break for (a) Experiment 03 and (b) Experiment 04.

Figure 9: Velocity and vorticity fields for SEOM model at shelf break level for same parameters as for the central case Experiment 01 but for Rossby numbers (a) 0.2 and (b) 0.3; no laboratory experiments were performed for these parameters.

Figure 10: Velocity and vorticity fields for laboratory models for Experiment 05, the narrow canyon, at the observation levels $z^* = (a) -0.2$, (b) –0.4 and (c) –0.6.
### Table 1. Dimensional parameters; quantities with asterisks are not independent of the other parameters.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
<th>Ocean</th>
</tr>
</thead>
<tbody>
<tr>
<td>( R_C ) (radius of the coast)</td>
<td>35 cm</td>
<td>-</td>
</tr>
<tr>
<td>( R_S ) (radius at the shelf-break)</td>
<td>55 cm</td>
<td>-</td>
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<tr>
<td>( R_T ) (radius of the test tank)</td>
<td>90 cm</td>
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<td>( h_S ) (depth over the shelf)</td>
<td>2.5 cm</td>
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<td>( h_D ) (maximum depth)</td>
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<td>L (length of the canyon)</td>
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<td>15 km</td>
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<tr>
<td>( f ) (Coriolis parameter)</td>
<td>0.50 s(^{-1})</td>
<td>(10(^{-4})) s(^{-1})</td>
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<tr>
<td>( N ) (Brunt-Väisälä frequency)</td>
<td>1.25 s(^{-1}), 2.5 s(^{-1})</td>
<td>(10(^{-3})) s(^{-1})</td>
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<tr>
<td>( \omega_0 ) (forcing frequency)</td>
<td>0.126 s(^{-1}), 0.262 s(^{-1}), 0.628</td>
<td>(10(^{-2})) s(^{-1}) (wind)~1.4(10(^{-4})) s(^{-1}) (tidal)</td>
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<td>( T = \frac{2\pi}{\omega_0} ) (time-scale of forcing)</td>
<td>10~25 s</td>
<td>From a day to several months.</td>
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<tr>
<td>( \nu ) (viscosity)</td>
<td>0.01 cm(^2)s(^{-1}) (kinematic)</td>
<td>10-80 cm(^2)s(^{-1}) (vertical)</td>
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<tr>
<td>( u_0 ) (amplitude of forcing, shelf break)</td>
<td>0.5 cms(^{-1}), 1.0 cms(^{-1})</td>
<td>20 cms(^{-1})</td>
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<tr>
<td>( X = \frac{2u_0}{\omega_0} ) (excursion of forcing current)*</td>
<td>1.6 cm ~ 15.9cm</td>
<td>3~40 km</td>
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<tr>
<td>( L_D = \frac{Nh_D}{f} ) (Rossby radius of deformation)*</td>
<td>31.2 cm, 62.5 cm</td>
<td>15 km</td>
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</table>
Table 2. Dimensionless parameters.

<table>
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<td>$R_S/R_T$ (shelf to tank radius ratio)</td>
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<td>$W/L$ (horizontal aspect ratio)</td>
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<td>$W/R_T$ (canyon width to tank radius ratio)</td>
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<tr>
<td>$Bu = (N^2h_D^2)/(f^2W^2)$ (Burger number)</td>
<td>2.5, 10.0</td>
<td>4.6</td>
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<td>$E = \nu/(fh_S^2)$ (Ekman number)</td>
<td>$(3.2) \times 10^{-3}$</td>
<td>$4.4(10)^{-6}$</td>
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<td>$Ro_t = \omega_0/f$ (temporal Rossby number)</td>
<td>1.4-10</td>
<td>0.25, 0.52, 1.25</td>
</tr>
<tr>
<td>$X^* = X/W$ (normalized excursion)</td>
<td>0.16~0.80</td>
<td>0.2~2.7</td>
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Table 3. Laboratory experiments and numerical simulations; the Ekman number for all runs is 3.2$(10)^3$

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<th>Boundary Condition</th>
<th>Viscosity Numerical</th>
<th>L/W</th>
<th>Ro</th>
<th>Ro_t</th>
<th>Bu</th>
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<td>L/N</td>
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<td>1.33</td>
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<td>0.52</td>
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<tr>
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<td>L/N</td>
<td>Shear Stress</td>
<td>1</td>
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<td>1.33</td>
<td>0.1</td>
<td>0.52</td>
</tr>
<tr>
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<td>L/N</td>
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<td>100</td>
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<td>1.33</td>
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Table 4. Non-dimensional volume fluxes per unit depth ($Q_1^*, Q_2^*, Q_3^*$), horizontal integral of divergence ($D^*$) and maximum and average kinetic energy per unit mass ($KE_{\text{max}}^*, KE_{\text{ave}}^*$) obtained from laboratory experiments for the designated experiments and observation levels; see Figure 4 and text.

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<th>Level (-z*)</th>
<th>$Q_1^*$</th>
<th>$Q_2^*$</th>
<th>$Q_3^*$</th>
<th>$D^*$</th>
<th>$KE_{\text{max}}^*$</th>
<th>$KE_{\text{ave}}^*$</th>
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<td>-0.026</td>
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<tr>
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<td>---</td>
<td>0.002</td>
<td>0.045</td>
<td>0.006</td>
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<tr>
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Table 5. Non-dimensional volume fluxes per unit depth (Q1*, Q2*, Q3*), area integral of vertical velocity (Q4*, positive upwards), and the maximum and average kinetic energy per unit mass (KE*_max, KE*_ave) obtained from the numerical experiments.

<table>
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<th>Experiment</th>
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<th>Q2*</th>
<th>Q3*</th>
<th>Q4*</th>
<th>KE*_max</th>
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Figure 1: Physical system.
Figure 2: Vorticity (left) and horizontal divergence (right) fields for the central experiment discussed by PHB (Experiment 01 in the present study) as obtained from (a) the laboratory, (b) the SEOM model using a parameterized shear stress condition along the model floor and (c) the SEOM model using a no-slip condition, including a highly resolved Ekman layer, along the model floor.
Figure 3. Velocity fields for SEOM model at the shelf break level corresponding to the central experiment of PHB for horizontal and vertical viscosities having the respective factors (a) (100, 1), and (b) (10,10).
Figure 4 Control volume
Figure 5. Characteristic speed of the normalized time-mean flow $\ddot{U}/u_0$ (i.e., defined as the square root of the average kinetic energy per unit mass as found for the area at the shelf break level as sketched in Figure 4a) at the shelf break level as obtained from the laboratory experiments and the numerical model against the scaling relation $\lambda = (h_o / h_s) RoRe^{1/2} Bu^{-1/2} E^{-1/2}$. The symbols near the data points correspond to either laboratory (L) or numerical (N) experiments numbered in Table 3. The dashed line is the best fit $\ddot{U}/u_0 = (0.9\lambda + 12.7) \times 10^{-2}$. 
Figure 6: Velocity and vorticity fields for laboratory (left) and SEOM (right) models for Experiment 01 at the observation levels $z^*$ equals (a) –0.1, (b) –0.2 and (c) –0.4.
Figure 7. Velocity and vorticity fields for laboratory (left) and SEOM (right) models for Experiment 02 at the observation levels $z^*$ equals (a) –0.2, (b) –0.4 and (c) –0.6.
Figure 8. Velocity and vorticity fields for laboratory (left) and SEOM (right) models for the level at the shelf break for (a) Experiments 03, and (b) Experiments 04.
Figure 9. Velocity and vorticity fields for SEOM model at shelf break level for same parameters as for the central case Experiment 01 but for Rossby numbers (a) 0.2 and (b) 0.3; no laboratory experiments were performed for these parameters.
Figure 10. Velocity and vorticity fields for laboratory models for Experiment 05, the narrow canyon, at the observation levels $z^*$ equals (a) –0.2, and (b) –0.4 and (c) –0.6.