Wind-Induced Exchange in Semi-Enclosed Basins

Rosario Sanay
Old Dominion University

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WIND-INDUCED EXCHANGE IN SEMI-ENCLOSED BASINS

by

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A Dissertation Submitted to the Faculty of Old Dominion University in Partial Fulfillment of the Requirement for the Degree of

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ABSTRACT

WIND-INDUCED EXCHANGE IN SEMI-ENCLOSED BASINS

Rosario Sanay
Old Dominion University, 2003
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The wind-induced circulation over laterally varying bathymetry was investigated in homogeneous and in stratified systems using the three-dimensional Regional Ocean Model (ROMS). For homogeneous systems, the focus was to describe the influence of the earth's rotation on the lateral distribution of the flow with particular emphasis on the transverse circulation. Along-basin wind-stress with no rotation caused a circulation dominated by an axially symmetric transverse structure consisting of downwind flow over the shoals and upwind flow in the channel along the whole domain. Transverse circulation was important only at the head of the system where the water sank and reversed direction to move toward the mouth. The wind-induced flow pattern under the effects of the earth's rotation depended on the ratio of Ekman depth \((d)\) to the maximum basin's depth \((h)\). The solution tended to that described in a non-rotating system as \(h/d\) remained equal to or below one. For higher values of \(h/d\), the longitudinal flow was axially asymmetric. Maximum downwind flow was located over the right shoal (looking downwind). The transverse component of velocity described three gyres. The main gyre was clockwise (looking downwind) and occupied the entire basin cross-section, as expected from the earth's rotation and the presence of channel walls. The other two gyres were small and localized, and were linked to the lateral distribution of the along-channel velocity component, which in turn, was dictated by bathymetry. In stratified systems, the main focus was to study the interaction between the wind-driven and buoyancy-induced flow over laterally varying bathymetry.
In particular the influence of the earth's rotation and the transverse circulation were examined. The interaction between the wind-induced and buoyancy-driven flow was characterized by the Wedderburn number \((W_e)\) which compares wind stress accelerations to baroclinic pressure gradient accelerations. The influence of the earth's rotation was characterized by the inverse of \(h/d\), i.e., the Ekman number. For both rotating and non-rotating systems under strong winds \((W_e > 1)\), the wind-induced pattern of downwind flow over the shoals and up-wind flow in the channel masked any effects of buoyancy driven flows as the water column remained nearly vertically homogeneous. For weak up-estuary winds \((W_e < 1)\), the gravitational circulation (with or without rotation) remained almost unaltered. Only the upper part of the water column was modified by the wind-stress, showing an increased surface mixed layer and reduction (increment) of the seaward flow with up-estuary (down-estuary) wind. The non-rotating experiments showed axially symmetric distribution of flow and salinity fields. Rotating cases under weak wind conditions showed low Ekman number, such that rotation effects translated into axial asymmetries of salinity and flow fields. The transverse circulation and the salinity field showed a distribution expected from the balance of the lateral density gradient force per unit mass and Coriolis accelerations. Rotating cases under strong wind conditions exhibited high Ekman numbers and tended to be axially symmetric, responding to a transverse balance between lateral pressure gradient and friction. Transverse flows showed two gyres located over the shoals in response to the lateral density gradient. These results compared favorably with a limited set of observations and are expected to motivate future measurements.
To my husband,

Héctor Perales,

with love.
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I am thankful to Hector, my husband, and to my parents, brothers and sisters for their love and spiritual support during these years.

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CHAPTER I
INTRODUCTION

Estuaries respond to a variety of forcing mechanisms over a wide range of time scales. Among all of them wind-induced circulation has been recognized as one of the most important mechanisms in the long-term transport, and in turn, in the dispersion of dissolved and suspended matter. Analytical and numerical models have been developed to study the wind-induced circulation under different scenarios of bathymetry, stratification, wind-stress and vertical mixing. Each of these agents has been found to affect the longitudinal and lateral distribution of the flow and density field. In most studies related to these topics, the effects of the earth's rotation have been ignored. This research addresses this void by focusing mainly in two aspects: 1) the influence of the earth's rotation on wind-induced flow in a homogeneous basin with laterally varying bathymetry, and 2) the interaction of wind-induced and buoyancy-driven flow (without and with rotation effects) in a basin with laterally varying bathymetry. Of particular interest was the description of the lateral distribution of the transverse flow, which has also been largely overlooked.

I.1 BACKGROUND

Analytical solutions for homogeneous systems

Csanady (1982) described the lateral variability of the vertically integrated transport driven by the wind over laterally varying bathymetry in absence of rotational effects. The solution (valid away from the closed ends) showed that the vertically integrated transport was downwind over the shoals and upwind in deeper water. In shallow water the accelerations produced by wind-stress are greater than those from the barotropic pressure gradient. In deep water the pressure gradient dominates, because of the

This dissertation follows the style of the Journal of Physical Oceanography.
depth-dependent nature of wind-induced accelerations, and a return flow develops. Hunter and Hearn (1987) solved the along-channel lateral and vertical variation of the wind-driven circulation in a non-rotating system and showed that the lateral distribution of the along-channel flow was sensitive to the bottom roughness and to the shape of the depth distribution. They used three different eddy viscosity parameterizations finding no-relevant differences in the results. Wong (1994), proposed an analytical solution for the lateral structure of the flow driven by local winds, remote forcing and buoyancy forcing for non-rotating systems, using a constant eddy viscosity in the cross-section. Results for local winds were consistent with those of Csanady (1982). Winant (2003) presented a three dimensional, linear, barotropic model to describe the wind-induced flow over laterally varying bathymetry on an f-plane. The influence of the earth’s rotation was characterized by the $h/d$ ratio, where $h$ is the maximum depth of the basin and $d$ is the Ekman depth. For large $h/d$ values, the along-channel flow showed axial asymmetries and transverse circulations played an important role. As $h/d$ approached one, the circulation pattern induced by the wind approached that described by Wong (1994) in a non-rotating system.

**Numerical solutions for homogeneous systems**

Hearn et al. (1987) found, through observations and numerical simulations, that the wind-induced circulation in a shallow bay was significantly modified by the bottom topography. They found that the water exchange between a shallow system and the adjacent water can be intensified by a deep channel dredged along the prevailing wind direction. They mentioned that even though the vertical spiraling expected from Ekman dynamics was not evident in shallow systems, the Coriolis acceleration affects the direction of the current vectors. Also, the water exchange between the systems could be strongly modified by the earth’s rotation under certain wind directions. Signell et al. (1990), found that the wind-driven circulation in shallow embayments, and in turn
the flushing time, was sensitive to both the shape of the cross-section and the effects of surface waves. Flushing time increased by increasing the slope of the cross-channel bathymetric profile, but decreased through inclusion of surface waves, which in turn increased the bottom drag and in consequence the strength of the circulation. Glorioso and Davis (1995) extended Signell et al.'s (1990) work by analyzing the influence of changes of bottom topography, eddy viscosity parameterization, and wave-current interaction, on the wind-driven circulation in shallow homogeneous systems. Using a full three dimensional numerical model, they found that for flat-bottom embayments the wind-driven flow was sensitive to the eddy viscosity formulation (i.e. to the turbulence closure). In contrast, laterally varying embayments were insensitive to the turbulence closure used. In basins with V-cross sections, horizontal variability rather than vertical variability dominates the circulation pattern. The vertical distribution of the flow is more uniform such that the circulation pattern, and in turn flushing time, is no longer sensitive to the eddy viscosity formulation.

*Analytical models for stratified systems*

Wong (1994) reported a simple analytical model of gravitational circulation in a channel with laterally varying bathymetry. Lateral variability of the along-channel gravitational flow dominated over vertical stratification in a flat bottom cross-section (Hansen and Rattray 1966). In the analytical model, the dynamical balance was considered between the tidally averaged horizontal pressure gradient and vertical shear stress divergence. Kasai et al. (2000) reported that the structure of the lateral variability depends on the competition between Coriolis and frictional forces. This competition can be described with the Ekman number $E = \frac{A_o}{f h^2}$, which is analogous to the ratio $d/h$. For $E \sim 1$, lateral variability prevails over vertical variability and for $E < 1$, vertical variability dominates over lateral structure. Valle-Levinson
et al. (2003a) extended Kasai et al.'s (2000) work by considering a more realistic lateral pressure gradient. They applied the analytical model to arbitrary bathymetries and analyzed the transverse circulation. The analytical solutions showed transverse asymmetries in the flow, relative to a mid-channel centerline. The flow became symmetric as the Ekman number increased. Observations in different rotating systems compared satisfactorily with the analytical solution. All of the above solutions omit wind-forcing effects.

**Numerical models for stratified systems**

Valle-Levinson and O'Donnell (1996) reported that density-driven flow in an estuary was influenced by the presence of a channel. Inflow developed in the channel and outflow developed over the shoals, in contrast with the 2-layer estuarine circulation described over a flat bottom estuary. They also found that the depth of the channel was important in determining the strength and salinity of the inflow. The strength of the flow increased as the channel became deeper. Xie and Eggleston (1999) studied the wind and density driven circulation pattern in the Croatan-Albemarle-Pamlico estuarine system by using a three-dimensional numerical model. The system has three inlets that communicate the estuary to the ocean. They found the circulation pattern and water exchange between the estuary and the continental shelf to be highly sensitive to the wind direction. The numerical results showed mainly two exchange patterns induced by W-SW and E-NE wind conditions. Under W and SW wind conditions (i.e. seaward) the exchange pattern showed two layer circulation, while for E and NE wind, the exchange pattern at the three inlets showed inward or outward flow through the entire water column.

**Novelties of this work**

Despite the extensive literature on the subject of wind-induced circulation on coastal plain estuaries, this is the first study in which the homogeneous case is treated with
a numerical model that emphasize lateral circulation. This study extends the work of Glorioso and Davies (1995) and Winant (2003) by including advective effects and turbulence closure schemes in the solution. It further describes the different wind-induced responses as a function of the ratio $h/d$ but exploring the distinct effects of wind stress and maximum water depth on that ratio. In terms of the stratified case, this is the first effort that explores the transverse structure of the flow that results from the competition between density-induced and wind-driven flows over laterally varying bathymetry. This is done with three-dimensional process-oriented numerical simulations that include and exclude Earth’s rotation effects.

I.2 THESIS OUTLINE

Two major issues related to the wind-induced circulation in laterally varying bathymetry have been addressed in this work:

The influence of the earth’s rotation in homogeneous and stratified systems with emphasis on the lateral distribution of the transverse circulation.

The lateral distribution of the flow and salinity fields in stratified systems under different wind-stress conditions.

These issues are addressed with a 3D numerical model. In Chapter 2, a brief description of the numerical model is presented. The influence of the earth’s rotation on the wind-induced circulation in homogeneous systems is examined in Chapter 3. The interaction between the wind-induced circulation and buoyancy-induced flow in non-rotating and rotating systems is presented in Chapter 4. The key findings of this investigation are summarized in Chapter 5.
CHAPTER II

METHODS

A brief description of the main features of the numerical model used and the model set-up are listed here.

II.1 MODEL DESCRIPTION

The numerical model used was the Regional Ocean Model System (ROMS). This model has been applied successfully to a wide range of regional and basin-scale studies (e.g. Haidvogel et al. 2000; She and Klinck 2000; MacCready and Geyer 2001; Marchesiello et al. 2001).

II.1.1 Equations of motion and boundary conditions

The numerical model is free-surface and finite difference based on the non-linear terrain-following coordinates of Song and Haidvogel (1994). It solves the incompressible hydrostatic primitive equations with potential temperature, salinity and a non-linear equation of state. This system (excepting the non-linear equation of state) in Cartesian coordinates can be written as (Hedström 1997):

\[
\frac{\partial u}{\partial t} + \vec{v} \cdot \nabla u - f v = - \frac{\partial \phi}{\partial x} + F_u + D_u, \\
\frac{\partial v}{\partial t} + \vec{v} \cdot \nabla v + f u = - \frac{\partial \phi}{\partial y} + F_v + D_v, \\
\frac{\partial T}{\partial t} + \vec{v} \cdot \nabla T = F_T + D_T, \\
\frac{\partial S}{\partial t} + \vec{v} \cdot \nabla S = F_S + D_S, \\
\frac{\partial \phi}{\partial z} = -\frac{\rho g}{\rho_0},
\]

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TABLE 1. The variables used in the numerical model.

<table>
<thead>
<tr>
<th>Variable</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>$u, v, w$</td>
<td>The $(x, y, z)$ components of the vector velocity $\vec{v}$</td>
</tr>
<tr>
<td>$\phi(x, y, z, t)$</td>
<td>Dynamic pressure $\phi = (P/\rho_0)$</td>
</tr>
<tr>
<td>$T(x, y, z, t)$</td>
<td>Potential temperature</td>
</tr>
<tr>
<td>$S(x, y, z, t)$</td>
<td>Salinity</td>
</tr>
<tr>
<td>$\rho_0 + \rho(x, y, z, t)$</td>
<td>Total in situ density</td>
</tr>
<tr>
<td>$f(x, y)$</td>
<td>Coriolis parameter</td>
</tr>
<tr>
<td>$g$</td>
<td>Acceleration of gravity</td>
</tr>
<tr>
<td>$D_u, D_v, D_T, D_S$</td>
<td>Diffusive terms</td>
</tr>
<tr>
<td>$F_u, F_v, F_T, F_S$</td>
<td>Forcing terms</td>
</tr>
<tr>
<td>$\zeta(x, y, t)$</td>
<td>The surface elevation</td>
</tr>
</tbody>
</table>

\[
\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0, \quad (6)
\]

where the variables are defined in Table 1. Equations (1) and (2) represent the horizontal momentum balance in the $x$ and $y$ directions, respectively. Equations (3) and (4) are the statements of conservation of heat and salt, respectively. Vertical momentum balance under the hydrostatic approximation (where density variations are neglected in the momentum equations except for their contribution to the buoyancy forcing) is given by (5). Equation (6) represents conservation of mass for an incompressible fluid.

ROMS allows temperature, salinity and momentum fluxes to be prescribed at the surface. Vertical boundary conditions in ROMS are prescribed as follows (Hedström 1997):

At the surface ($z = \zeta(x, y, t)$):

\[
A_z \frac{\partial u}{\partial z} = \tau^x(x, y, t)
\]

\[
A_z \frac{\partial v}{\partial z} = \tau^y(x, y, t)
\]
\[ K_T \frac{\partial T}{\partial z} = \frac{Q_T}{\rho_0 c_p} + \frac{dQ_T}{dT} \frac{T - T_{ref}}{\rho_0 c_p} \]

\[ K_S \frac{\partial S}{\partial z} = \frac{(E - P)S}{\rho_0} \]

\[ w = \frac{\partial \zeta}{\partial t} \]

At the bottom \((z = -h(x,y))\):

\[ A_z^x \frac{\partial u}{\partial z} = \tau_s^x(x,y,t) \]

\[ A_z^y \frac{\partial v}{\partial z} = \tau_s^y(x,y,t) \]

\[ K_T \frac{\partial T}{\partial z} = 0 \]

\[ K_S \frac{\partial S}{\partial z} = 0 \]

\[ \vec{v} \cdot \nabla h - w = 0 \]

where \(\tau_s\) represents the surface kinematic \((\frac{\tau_s}{\rho_0})\) wind-stress, \(Q_t\) is the surface heat flux, \((E - P)\) is evaporation minus precipitation and \(T_{ref}\) is the surface reference temperature. For lateral boundary conditions, ROMS is very flexible, having several options for closed and open boundaries (see Marchesiello et al. 2001).

Another important feature of ROMS is the terrain-following transformation (s-coordinate). This transformation allows a non-linear stretching of the vertical coordinates to better resolve the corresponding frictional layers (Haidvogel et al. 2000).

II.1.2 Numerical approach

The model employs a centered second-order finite-difference approximation (Arakawa-C grid) in the horizontal discretization. It uses a staggered finite difference treatment
in the vertical coordinate (Haidvogel et al. 2000). A time-splitting technique is used to separate the baroclinic and barotropic components, allowing the use of internal and external time steps. For the time-step advancement a leapfrog/Adams-Moulton predictor-corrector scheme is used (Marchesiello et al. 2001). This scheme has been found to have excellent non-dispersive properties for the advection terms (Haidvogel et al. 2000). The vertical diffusion terms are solved by the semi-implicit Crank-Nicholson scheme.

Horizontal and vertical subgridscale problems can be solved using different closure schemes available in ROMS. Also, a quasi-horizontal smoothing of momentum and/or tracers can be done using a high-order advection scheme (third order and upstream biased), which implicitly contributes to the horizontal smoothing, as mentioned in Haidvogel et al. (2000).

A more complete and detailed description of the model can be found in Song and Haidvogel (1994), Hedström (1997), Haidvogel and Beckmann (1998), Haidvogel and Beckmann (1999), Haidvogel et al. (2000) and Marchesiello et al. (2001).

II.2 MODEL SET-UP

The numerical experiments focused on the wind-induced circulation in basins with laterally varying bathymetry under different scenarios of shoals-channel configurations, vertical mixing, stratification and rotation. The model was set up for two main scenarios: homogeneous and stratified systems.

All numerical experiments included all the terms of the primitive equations, (1) to (6), except the horizontal diffusive terms represented by $D_u$, $D_v$, $D_T$ and $D_S$. The advection scheme used was third order and upstream biased, so no explicit horizontal viscosity or diffusivity were needed, according to Haidvogel et al. (2000). Temperature was always kept constant. The KPP turbulence-closure scheme is used throughout, except where noted. Salinity and temperature surface fluxes were turned
off in all the experiments, while a constant surface wind-stress was prescribed in some of them. The bottom boundary conditions for the normal derivatives of salinity and temperature were set to zero. The quadratic drag law was used to parameterize bottom friction with a non-dimensional bottom drag coefficient of 0.0025. Free-slip condition was established for all closed boundaries and the no-gradient condition was used in all open boundaries for all variables.

II.2.1 Numerical domain

The numerical domain consisted of an elongated longitudinally uniform basin with laterally varying bathymetry. The domain was 100 km long and 10 km wide. The $x$-axis coincided with the southern lateral wall of the basin and pointed toward the head of the system. The $y$-axis laid along the open boundary at $x = 0$ (Fig. 1). Grid intervals were 2 km along the $x$-direction and 200 m in the $y$-direction. The number of vertical levels ranged from 10 to 30 depending on the maximum depth of the bathymetric profile employed. The vertical levels were spread out between the local bottom and the free surface, allowing more vertical resolution near the surface and near the bottom than in the interior.

FIG. 1. Sketch of the numerical domain.
II.2.2 Experiment design

Numerical experiments were classified into four main scenarios: homogeneous non-rotating cases, homogeneous rotating cases, stratified non-rotating cases and stratified rotating cases. These scenarios are outlined next.

Homogeneous cases

A total of 12 numerical experiments were carried out to illustrate the wind-induced circulation in homogeneous systems with and without rotation (Table 2). The non-rotating homogeneous cases focused mainly on bathymetric effects on the wind-induced flow. A total of six different bathymetric profiles were used for this purpose, with a triangular cross-section illustrating a baseline case that could be compared to analytical solutions (e.g. Wong 1994). Gaussian cross sections were chosen to approximate a more realistic bathymetry (shoals-channel configuration). The rotating homogeneous cases focused on the influence of the earth's rotation on the lateral distribution of the three velocity components and on particle trajectories. The influence of the earth's rotation was characterized as a function of the ratio of the maximum depth \( h \) to the Ekman depth \( d \) as described in Chapter III.

For all these experiments, the basin was open at the seaward boundary only. The fluid in the channel started from rest and a wind-stress increased linearly during the first six hours of simulation and was constant afterward. The wind stress acted toward the head of the system aligned with the x-axis and blew uniformly throughout the domain. The results of the simulations corresponding to homogeneous cases are presented in Chapter III.

Stratified cases

A total of 16 numerical experiments were carried out to study the effects of wind forcing on stratified systems with lateral depth variation (Table 3). Stratified non-rotating cases focused on the non-rotating interaction between the wind driven and
TABLE 2. Parameters used in the homogeneous numerical experiments. \( \tau \) stands for wind stress, \( f \) for Coriolis parameter, \( h_0 \) for minimum depth and \( h \) for maximum depth.

<table>
<thead>
<tr>
<th>Exp. No.</th>
<th>( \tau ) (Pa)</th>
<th>( f ) (s(^{-1}))</th>
<th>( h_0 ) (m)</th>
<th>( h ) (m)</th>
<th>( h/d )</th>
<th>Bathymetric profile</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.08</td>
<td>0</td>
<td>3</td>
<td>20</td>
<td>-</td>
<td>triangular</td>
</tr>
<tr>
<td>2</td>
<td>0.08</td>
<td>0</td>
<td>8</td>
<td>20</td>
<td>-</td>
<td>Gaussian*, ( c = 1200 )</td>
</tr>
<tr>
<td>3</td>
<td>0.08</td>
<td>0</td>
<td>8</td>
<td>20</td>
<td>-</td>
<td>Gaussian, ( c = 2000 )</td>
</tr>
<tr>
<td>4</td>
<td>0.08</td>
<td>0</td>
<td>10</td>
<td>20</td>
<td>-</td>
<td>Gaussian, ( c = 3000 )</td>
</tr>
<tr>
<td>5</td>
<td>0.08</td>
<td>0</td>
<td>6</td>
<td>20</td>
<td>-</td>
<td>Gaussian, ( a = 6, b_1 = 7, b_2 = 14 )</td>
</tr>
<tr>
<td>6</td>
<td>0.08</td>
<td>0</td>
<td>7</td>
<td>20</td>
<td>-</td>
<td>Gaussian, ( a = 7, b_1 = 2, b_2 = 14 )</td>
</tr>
<tr>
<td>7</td>
<td>0.08 ( 10^{-4} )</td>
<td>3</td>
<td>20</td>
<td>1.95</td>
<td>triangular</td>
<td></td>
</tr>
<tr>
<td>8</td>
<td>0.08 ( 10^{-4} )</td>
<td>8</td>
<td>20</td>
<td>2.00</td>
<td>Gaussian, ( c = 1200 )</td>
<td></td>
</tr>
<tr>
<td>9</td>
<td>0.08 ( 10^{-4} )</td>
<td>3</td>
<td>60</td>
<td>3.60</td>
<td>triangular</td>
<td></td>
</tr>
<tr>
<td>10</td>
<td>0.08 ( 10^{-4} )</td>
<td>3</td>
<td>8</td>
<td>1.16</td>
<td>triangular</td>
<td></td>
</tr>
<tr>
<td>11</td>
<td>0.005 ( 10^{-4} )</td>
<td>3</td>
<td>20</td>
<td>3.95</td>
<td>triangular</td>
<td></td>
</tr>
<tr>
<td>12</td>
<td>0.5 ( 10^{-4} )</td>
<td>3</td>
<td>20</td>
<td>1.19</td>
<td>triangular</td>
<td></td>
</tr>
</tbody>
</table>

*Gaussian profiles are defined in (7) and (8)

density-induced flow on the lateral distribution of the three velocity components and salinity field. The rotating cases explored the influence of the Coriolis acceleration on wind-induced circulation in non-homogeneous systems. For both, rotating and non-rotating experiments, sensitivity analysis of the gravitational adjustment to vertical mixing was carried out to assess the consistency of the numerical results with those reported by analytical solutions. For the sensitivity analysis of these cases, the KPP-closure scheme (which has been found to solve successfully the vertical mixing in shallow systems, i.e. Geyer and MacCready 2001) and three different constant eddy viscosity and eddy diffusivity values were used. The relative importance of wind-stress in the dynamics of buoyancy-driven flow was assessed with the non-dimensional Wedderburn number (e.g. Geyer 1997). The influence of the earth's rotation was characterized with the Ekman number as in Kasai et al. (2000). These influences are described in chapter IV.
TABLE 3. Parameters used in different stratified numerical experiments, $\tau$ stands for wind stress, $f$ for Coriolis parameter, $A_z$ is the eddy viscosity coefficient and $K_z$ is the eddy diffusivity coefficient.

<table>
<thead>
<tr>
<th>Exp. No.</th>
<th>$A_z$ (m$^2$s$^{-1}$)</th>
<th>$K_z$ (m$^2$s$^{-1}$)</th>
<th>$\tau$ (Pa)</th>
<th>$f$ (s$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>KPP</td>
<td>KPP</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>2</td>
<td>$1 \times 10^{-2}$</td>
<td>$1 \times 10^{-3}$</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>3</td>
<td>$1 \times 10^{-3}$</td>
<td>$1 \times 10^{-4}$</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>4</td>
<td>$5 \times 10^{-5}$</td>
<td>$6 \times 10^{-3}$</td>
<td>0.01</td>
<td>0</td>
</tr>
<tr>
<td>5</td>
<td>KPP</td>
<td>KPP</td>
<td>0.1</td>
<td>0</td>
</tr>
<tr>
<td>6</td>
<td>KPP</td>
<td>KPP</td>
<td>0.2</td>
<td>0</td>
</tr>
<tr>
<td>7</td>
<td>KPP</td>
<td>KPP</td>
<td>-0.01</td>
<td>0</td>
</tr>
<tr>
<td>8</td>
<td>KPP</td>
<td>KPP</td>
<td>0.01</td>
<td>0</td>
</tr>
<tr>
<td>9</td>
<td>KPP</td>
<td>KPP</td>
<td>0</td>
<td>$10^{-4}$</td>
</tr>
<tr>
<td>10</td>
<td>$1 \times 10^{-2}$</td>
<td>$1 \times 10^{-3}$</td>
<td>0</td>
<td>$10^{-4}$</td>
</tr>
<tr>
<td>11</td>
<td>$1 \times 10^{-3}$</td>
<td>$1 \times 10^{-4}$</td>
<td>0</td>
<td>$10^{-4}$</td>
</tr>
<tr>
<td>12</td>
<td>$5 \times 10^{-5}$</td>
<td>$6 \times 10^{-5}$</td>
<td>0</td>
<td>$10^{-4}$</td>
</tr>
<tr>
<td>13</td>
<td>KPP</td>
<td>KPP</td>
<td>0.01</td>
<td>$10^{-4}$</td>
</tr>
<tr>
<td>14</td>
<td>KPP</td>
<td>KPP</td>
<td>0.1</td>
<td>$10^{-4}$</td>
</tr>
<tr>
<td>15</td>
<td>KPP</td>
<td>KPP</td>
<td>0.2</td>
<td>$10^{-4}$</td>
</tr>
<tr>
<td>16</td>
<td>KPP</td>
<td>KPP</td>
<td>-0.01</td>
<td>$10^{-4}$</td>
</tr>
</tbody>
</table>

For all stratified experiments the basin was open at seaward and landward boundaries. All the experiments started from rest. The initial salinity field was vertically uniform with a longitudinal gradient that varied linearly in the domain. The salinity field was chosen such that the longitudinal density gradient was $1 \times 10^{-4}$ kg m$^{-4}$, which represent a typical value in coastal plain estuaries (Valle-Levinson and O'Donnell 1996). A constant river discharge (100 m$^3$s$^{-1}$) at the landward boundary was applied. During the first 30 h there was no longitudinal forcing other than that exerted by the baroclinic pressure gradient. Wind forcing was ramped-up linearly for 2 to 6 hours, depending on the wind stress value. The wind stress was aligned with the $x$-axis and blew uniformly throughout the domain. The results of the simulations corresponding to the stratified cases are presented in Chapter IV.
CHAPTER III

WIND-DRIVEN FLOW IN HOMOGENEOUS SYSTEMS

III.1 INTRODUCTION

The wind-induced exchange between an estuary and the adjacent continental shelf is recognized as one of the most important mechanisms in determining the long-term transport and distribution of dissolved and suspended matter that leave or enter an estuary. Both analytical (Csanady 1982; Hunter and Hearn 1987; Wong 1994; Friedrichs and Hamrick 1996; Winant 2003) and numerical models (Hearn and Hunter 1987; Signell et al. 1990; Glorioso and Davis 1995) on this topic have shown the importance of laterally varying bathymetry on the lateral distribution of the along-channel flow driven by local winds. For example, in flat bottom systems the velocity vertical profile is the same at each point of the cross section. The structure of the response is downwind at the surface and upwind at depth as required by continuity. In laterally varying bathymetry, the flow is strongly dependent on lateral position. The local wind forcing response is downwind at all depths in shallow regions and upwind mainly concentrated in the deep channel. Csanady (1982) showed that the zero vertically integrated transport driven by the wind, occurred at the average depth of the cross-section. In his calculations, he neglected Coriolis and bottom stress.

Observations

From observational data focused on the lateral distribution of the along-channel flow in semi-enclosed shallow basins, Valle-Levinson et al. (2001a) reported one of the few examples of wind-induced circulation in homogeneous systems with laterally varying bathymetry. The set of observations consisted of hydrographic and horizontal velocity profiles obtained at the entrance to the Bay of Guaymas, in Mexico, in February 2000 (Fig. 2). Guaymas Bay is about 10.4 km long and 8.5 km wide and has an average
FIG. 2. Bay of Guaymas location. Gray contours indicate bathymetry at 1 m intervals. ADCP trajectories are represented with dots. CTD stations are marked with white diamonds.

depth of 2 m. Communication to the adjacent Gulf of California is through a channel 3 km long, 1.6 km wide. The transverse bathymetric profile of the communication channel featured a V-shape with minimum depth of 4 m and maximum depth of 14 m. The observational periods were influenced by moderate sea breeze which was reflected in the residual circulation. Sub-tidal flow was mainly driven by the wind such that the flow over the shoals followed the direction of the wind, and flow in the deep channel was upwind, in agreement with the analytical solution of Wong (1994). Valle-Levinson et al. (2001a) also pointed out the possible influence of the earth’s rotation on the lateral distribution of the along-channel flow, as the core of maximum

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inflow appeared on the right side of the channel (Fig. 3). The apparent influence of the Earth’s rotation in these observations motivated my numerical experiments in homogeneous systems like the one observed.

![Cross-section](image)

**FIG. 3.** Cross-section (looking into the Bay of Guaymas) of the principal-axis component of the mean flow during the two experiments of February 2000. Positive values (light-shaded contours) indicate net outflow (from Valle-Levinson et al. 2001a).

**Previous work and relevance of present work**

Most of the published work on local wind-induced circulation refers to non-rotating systems and mainly concentrates on the lateral distribution of the along-channel flow. Just recently, Winant (2003) published a semi-analytical solution of the linear problem of the wind-induced circulation on a homogeneous and rotating system with a V-shape cross-section. The work reported in this chapter was developed simultaneously to Winant’s solution.

In this chapter, I report advances in the understanding of the flow pattern driven
by local winds in homogeneous estuaries with laterally varying bathymetry. I focus on two main topics, the influence of bathymetric changes on the lateral distribution of the flow and on the influence of Earth's rotation on the circulation pattern induced by local winds with emphasis in the transverse flow. For the first topic I explore the role played by the shape of the bathymetry and the value of the mean-depth on the lateral structure of the wind-induced flow. I extend Csanady's (1982) work by analyzing the 3D distribution of the flow over arbitrary cross-sections, and by analyzing the distribution of the flow in channels shallower than the mean depth. The second topic analyzed in this chapter is similar to that addressed by Winant (2003) but I complement it with the use of a numerical model that considers all the non-linear terms, advective and frictional, which are usually important in shallow systems. These numerical simulations allow the eddy viscosity to vary in space and time. In the first section of this chapter, I analyze the influence of bathymetric changes on the lateral distribution of the flow. In the second section I explore the influence of the earth's rotation on the circulation pattern induced by local wind. At the end of this chapter, a comparison between the non-linear numerical results and the solution reported by Winant (2003) is made, including the novel contributions of my work.

III.2 NUMERICAL EXPERIMENTS

A total of 12 numerical experiments illustrate the wind-induced flow over laterally varying bathymetry in a homogeneous fluid (Table 2). The first six experiments examined bathymetric effects on the wind-induced flow in non-rotating systems, and the other six experiments looked at the influence of the earth's rotation on the wind-induced flow.

In all calculations, motion in the channel started from rest through a linear increment of wind stress during the first six hours of simulation and constant wind stress
afterward. The wind stress acted toward the head of the system aligned with the $x$-axis and blew uniformly throughout the domain.

All the non-rotating cases used the same wind stress (0.08 Pa) and six different bathymetric profiles. The baseline case was run over a triangular cross-section, to compare to analytical and semi-analytical solutions (e.g. Wong 1994, Winant 2003), with minimum depth of 3 m and maximum depth of 20 m. The other five bottom shapes for non-rotating cases were chosen to study the effects of arbitrary transverse depth variations. Emphasis was placed on the effects of channel-shoal configurations and on the bottom slope. The bottom configurations consisted of Gaussian functions that tend to emulate coastal plain estuary topographies (e.g. Li and Valle-Levinson 1999; Reyes-Hernandez 2001). The Gaussian functions adopted in the numerical experiments were:

\[
h(y) = 8 + 12 e^{-\left(\frac{y-D/2}{c}\right)^2} \tag{7}
\]

and

\[
h(y) = a + b_1 e^{-\left(\frac{y-D/5}{900}\right)^2} + b_2 e^{-\left(\frac{y-D/5}{1200}\right)^2} \tag{8}
\]

Equation (7) describes a channel of width $D$, 20 m deep in the middle and edges with minimum depth of 8 m. The values of $c$ are listed in table 2. Equation (8) describes a bottom profile with two channels of depths $(a + b_1)$ and $(a + b_2)$ and a middle shoal with minimum depth equal to $a$. The values chosen for $a$, $b_1$ and $b_2$ are listed in table 2. The depth profiles with two channels were chosen such that in one, the shallow channel was deeper than the mean depth and in the second, the shallow channel was shallower than the mean depth. The cross-sectional areas of the double-channel profiles were the same.

Earth’s rotation influence was characterized as in Winant (2003), as a function of $h/d$ (where $h$ is the maximum depth and $d$ is the Ekman layer depth ($d = \sqrt{2A_z/f}$),
with $A_z$ the eddy viscosity coefficient, and $f$ the Coriolis parameter. This is equivalent to the inverse of the Ekman number. Any given value of this ratio can be obtained through adjustment of the wind stress value (which modifies $d$) or of the maximum water depth ($h$). Winant (2003) only examined variations of $h$. In order to analyze Earth's rotation response in terms of the different influences of wind stress and maximum water depth, five numerical experiments were run over triangular cross-sections. In three of the examples with rotation, the maximum depth was the same and wind stress was prescribed to yield values of $h/d \sim 1, 2$ and 4. The other two experiments, with rotation and triangular cross-section, were forced with a given wind stress but changing the maximum depth. The wind stress was the same as in the case of $h/d \sim 2$ but the maximum depths were chosen such that $h/d$ took values of $\sim 1$ and 4. Finally, a third case with $h/d \sim 2$ was run over a Gaussian bathymetric profile given in (8) to compare the results to the non-rotating case over the same bathymetry.

### III.3 RESULTS

This section presents results obtained in non-rotating and rotating systems. Rotation influences will be characterized as in Winant (2003), as a function of $h/d$, i.e., the ratio of maximum depth to the Ekman layer depth. In contrast to Winant's approach, where he examines the exclusive effects of $h$ on $h/d$, I also look at the influence of the wind stress $\tau$ on $d$ and on $h/d$. So, for any $h/d$ the magnitude of the response may be different if $\tau$ decreases or if $h$ increases. The results are mainly analyzed at two lateral sections: midway between the basin ends and at the head of the system. In order to reach numerical convergence, the difference of the total energy in the entire numerical domain between two consecutive time steps had to be of order $10^{-6}$ J. All figures shown in this section portray views toward the head of the system.
III.3.1 Non-rotating homogeneous system

III.3.1.1 Wind-induced flow

*Baseline case*

The first experiment represents the baseline case in which the $u$-velocity component of the flow shows lateral variability related to the bathymetry (Fig. 4). Over shallow water (shallower than the average depth), the wind drives downwind circulation throughout the water column. Maximum values of the downwind flow (0.24 m s$^{-1}$) were located at the surface and decreased with depth as expected from bottom friction. In the channel (area deeper than the average depth), the flow was up-wind over much of the water column, with maximum magnitudes of 0.15 m s$^{-1}$ appearing...
below mid-water in the channel. The lateral structure of the flow associated with the laterally varying bathymetry showed strongest lateral shears around the zero isotach, which intersects the bottom at the mean depth. Over shallow water, the main momentum balance was purely frictional between wind stress and internal stress divergence. In the channel, the pressure gradient given by the sea level slope had to be added to the momentum balance. This was consistent with previous work (e.g. Csanady 1982; Signell et al. 1990; Wong 1994; Friedrichs and Hamrick 1996) and provided confidence on the model performance to this forcing. Most of the velocity field was dominated by the $u$-velocity component as the other two components, $v$ and $w$, were very small. The lateral flow $v$ was $O(10^{-4} \text{ m s}^{-1})$ and $w$ was $O(10^{-6} \text{ m s}^{-1})$. This was true for all $x$-locations except near the closed end.

Owing to the presence of the closed boundary, streamlines must close near the head, reflecting an area of flow return. In this area, the lateral velocity component had magnitudes of up to $0.08 \text{ m s}^{-1}$ and converged toward the center of the cross-section. The vertical component was downward with magnitudes of up to $2.6 \times 10^{-4} \text{ m s}^{-1}$. Away from the returning area (about 6 km wide), the transverse circulation decreased rapidly. At a distance 10 km away from the closed end, the maximum magnitude of $v$ was only $\sim 0.001 \text{ m s}^{-1}$. These distributions were also consistent with Winant’s (2003) solutions.

In summary, for this baseline case, up-estuary wind drove water toward the head of the system, where it sank and converged, and reversed direction to move toward the mouth. An important feature of this non-rotating response was that horizontal and vertical velocity-components were axially symmetric in the numerical domain.

**Bathymetric effects**

To study the effects of arbitrary transverse depth variations on the wind-driven circulation, similar numerical experiments to the baseline case were run over five different Gaussian bathymetric profiles given by (7) and (8). The channel-shoals combination...
in the Gaussian profile resulted in essentially the same overall wind-induced exchange flows. These exchange patterns only differed in a few features compared with the triangular cross-section. In Fig. 5, downwind flow over the flat areas showed well defined vertical gradients in contrast to the pattern in the channel, which showed well defined horizontal gradients. The velocity vertical profiles over the shoals were independent of lateral position. They showed maximum downwind flow at the surface and minimum flow at the bottom as expected from wind-forcing and bottom-friction, respectively. In the channel, the magnitude of the flow was strongly dependent of the local position. These rapid changes in slopes of the along-channel isotachs were associated with the edges of the channel. Relatively strong lateral shears should be expected around the edges of the channel induced by the opposing flows at the edge of the channel. Near the head of the system maximum lateral flow was located over
the largest bathymetric slope. Also at the head of the system, vertical flow was downward everywhere except over the flat areas near the lateral walls of the domain where relatively weak upward flows ($\sim 1.4 \times 10^{-5} \text{ m s}^{-1}$) were found. These were associated with the lateral divergence of the flow in the vicinity of the lateral walls of the domain. Maximum downward flows ($\sim 2 \times 10^{-4} \text{ m s}^{-1}$) were located at the edges of the channel.

FIG. 6. Same as Fig. 4 but for a Gaussian cross section given by (7) with $c = 2000$ ($\max |v| = 4.5 \times 10^{-2} \text{ m s}^{-1}$, $\max |w| = 1.8 \times 10^{-4} \text{ m s}^{-1}$).

Comparing the three Gaussian bathymetric cross sections (Figs. 5 to 7), it is clear that the lateral shear of the $u$-velocity component decreased as bathymetry became smoother. This was illustrated by the separation of the $u$-velocity isotachs. The different lateral shears associated with different bathymetries suggested that lateral
shears should play an important role in the dynamics of basins with rapid bathymetric changes in the $y$-direction (e.g. Fig. 5) and under strong wind conditions. Also, maximum upwind flow tended to concentrate near the bottom as bathymetric changes became smoother. In other words, as the slope of the bottom decreased the maximum upwind flow spread out over the bottom, such that the solution approached that for flat bottom systems (i.e. Wong 1994; Signell et al. 1900). This is the first time that a wind-induced response over Gaussian bathymetry is described in detail and constitutes one of the new contributions of this work.

Another new contribution emphasizes the role played by the shape of the bathymetry and the value of mean-depth on the lateral structure of the wind-induced

FIG. 7. Same as Fig. 4 but for a Gaussian cross section given by (7) with $c = 3000$ (max $|v| = 4.2 \times 10^{-2}$ m s$^{-1}$, max $|w| = 1.3 \times 10^{-4}$ m s$^{-1}$)
flow. For this purpose, two additional numerical experiments with double-channel profile (given by (8)) were analyzed. In the first double-channel profile, both channels were deeper than the mean depth, whereas in the second double-channel profile only one channel was deeper than the mean depth. Both cases were forced with an up-channel wind stress of 0.08 Pa and both cross-sectional areas were about the same.

![Channel midway](#)

![At 2 km from the head](#)

**FIG. 8.** Same as Fig. 4 but for a Gaussian cross section given by (8) with $a = 6$, $b_1 = 7$ and $b_2 = 14$ ($\max |u| = 5.7 \times 10^{-2} \text{ m s}^{-1}$, $\max |w| = 3.7 \times 10^{-4} \text{ m s}^{-1}$)

For the case where both channels were deeper than the cross-section mean-depth, the $x$-component of the flow was upwind in both channels and downwind elsewhere. Maximum upwind flow was found in the deepest channel (Fig. 8), as was expected from bottom friction acting more noticeably in the shallow channel. Both upwind cores were centered with respect to each channel. This was true for all $x$-locations.
except near the head, where the upwind flow cores moved toward the left (looking toward the head), because most of the incoming water that returned toward the mouth of the system followed the deepest channel. As in the one-channel Gaussian shapes, the other two velocity components (lateral and vertical) were almost zero except near the head (Fig. 6). In this area, the lateral flow showed convergence toward the center of both channels, where most of the downwind flow in the middle shoal moved toward the deepest channel. The channel showed downwelling cells associated with the convergence of the downwind flow, and upwelling cells from the divergence of the upwind flow.

For the case where only one of the two channels was deeper than the mean depth

FIG. 9. Same as Fig. 4 but for a Gaussian cross section given by (8) with \(a = 7\), \(b_1 = 2\) and \(b_2 = 14\) (\(\max |u| = 7.7 \times 10^{-2} \text{ m s}^{-1}, \max |w| = 4.3 \times 10^{-4} \text{ m s}^{-1}\))
(Fig. 9), both channels showed lateral structure of the along-channel flow, but only upwind flow developed in the channel that was deeper than the mean depth. Downwind flow in the middle flat shoal exhibited vertical shears in contrast with the shallowest channel that showed lateral shears. As in the double channel case described above, the upwind flow was centered with respect to the deepest channel in most of the numerical domain but near the closed end. In that area, the upwind core moved toward the left as expected from the asymmetric convergence of the lateral flow toward the deepest channel.

Double-channel bathymetric profiles showed that lateral variations of depth induced lateral shears in the along-channel velocity component, but upwind flow developed only in channels deeper than the cross-section mean depth. This was because the lateral partition of the $u$-velocity component into upwind flow and downwind flow occurred around the average depth of the bathymetric profile, in agreement with the vertically integrated transports of Csanady (1982). Also noteworthy was the rapid change of the slope of the $u$ isotachs associated to the bathymetric change. Over the middle shoal the isotachs became horizontal but in the channels they bent upward to reflect lateral variability of the flow. Away from the influence of the closed end, the velocity field was mainly dominated by the $u$-velocity component, and the downwind flow cores were centered in each channel. At the head, the upwind flow cores moved toward the left, as expected from the laterally varying bathymetry. Most of the downwind flow returned toward the mouth of the system following the deepest channel and the asymmetric convergence of the lateral flow pushed the upwind core toward the left.

In summary from the bathymetric effects, it was clear that the zero isotach of the $u$-velocity component intersected the bottom at around mid-depth, concentrating the upwind flow in the deeper part of the cross-section. Laterally varying bathymetry produced lateral variation of the $u$-velocity component even in channels shallower than the cross-section mid-depth. In cross-sectional bathymetric profiles consisting
of a channel flanked by flat shoals, the isotachs were horizontal over the flat areas and became almost vertical in the channel. For channels with less pronounced slopes, maximum upwind flow moved deeper, i.e., the flatter the bathymetry, the deeper the maximum upwind flow. In general, wind-driven circulation in laterally varying bathymetry showed lateral shears that might become important in the dynamic balance in shallow systems forced by strong wind and/or systems with strong lateral bathymetric changes. The strong link between bathymetric slopes and along-channel isotach slopes, suggested the importance of inclusion of variable eddy viscosity in the calculations. The eddy-viscosity over the shoals should be sensitive to the eddy viscosity formulation, in contrast to that expected in the channel. This is because in the channel the flow was almost vertically uniform and over the shoals the flow showed strong vertical shears.

III.3.1.2 Particle trajectories

Particle trajectories were described for the baseline non-rotating case. Each particle trajectory was calculated through temporal integration of the steady state velocity field. Four particles were released at the entrance to the system. The initial vertical positions of the particles were surface \(z_0 = 2\) and one sixth of the maximum depth \(z_0 = 3.5\) m, and the initial lateral position were 1 km and 9 km, i.e. 1 km away from each side of the channel. Figure 10 illustrates the trajectory of the four particles released at the entrance. The left panels show horizontal views \((x-y\) plane\) and the right panels show transverse views \((y-z\) plane\) of the particle trajectory released at four different positions. In general, particles moved downwind toward the head in a rectilinear path, arrived at the head of the system where they sank and then moved upwind toward the entrance in a rectilinear path. The trajectory of the particles was independent of the release position across the system as expected from the axial symmetry of the velocity field. Particles released at the same position but
deeper showed similar paths but took longer to move through the system, owing to the reduction of velocity by bottom friction. These straightforward paths will be compared to those resulting from the wind-induced homogeneous flow field affected by the earth's rotation. Such response is explored next.
FIG. 10. Particle trajectories released from different locations and different depths (indicated with the solid triangle) in the steady state for the homogeneous baseline case without rotation ($\tau = 0.08$ Pa.). Left panel $x$-$y$ plane, right panel $y$-$z$ plane. Solid line indicates the incoming path, dotted line signals the outgoing one.
III.3.2 Rotating homogeneous system

As mentioned before the rotating cases were characterized by the ratio \((h/d)\), where \(d\) was the Ekman layer depth and was evaluated with the cross-sectional average of the eddy viscosity for each numerical experiment, once numerical convergence had been reached. The Coriolis parameter was \(10^{-4} \text{s}^{-1}\) in all cases. Also, the cross-sectional average of the ratio \(|v/u|\) was adopted as a diagnostic to evaluate lateral asymmetries in the flow.

III.3.2.1 Wind induced flow

*Rotating baseline case*

In order to compare the results from non-rotating and rotating systems, the non-rotating baseline case was run including the effect of the earth's rotation. The bathymetric profile was triangular, with minimum depth of 3 m and maximum depth \(h\) of 20 m, and the wind stress was 0.08 Pa toward the head. Under these conditions, the thickness of the surface Ekman layer occupied about one half of \(h\) (i.e. \(h/d \sim 2\)).

Inclusion of the earth's rotation into the dynamics caused significant differences in all velocity components relative to the non-rotating case. Although the zero isotach still intersected the bathymetric profile around the mean depth, and downwind flow appeared over the shoals and upwind flow in the channel, the lateral distribution of the \(u\)-velocity component featured axial asymmetries (Fig. 8). Stronger downwind flow was found over the right shoal than over the left shoal (looking into the system and downwind). The asymmetries of the \(u\)-velocity component became more evident near the head of the system, around 10 km away from the closed end, where the downwind flow began to converge. The core of maximum upwind flow near the head of the system moved toward the left (again looking downwind) as expected from differential advection of the along-channel flow. Differential advection meant that more water was transported downwind over the right shoal than over the left shoal because of
rotation. This differential advection induced a lateral pressure gradient at the head of the system that pushed the upwind flow core toward the left. The asymmetries resulting from this experiment suggested that, under the condition of $h/d \sim 2$, the Coriolis term played an important role in the dynamics.

![Channel midway](image1)

![At 2 km from the head](image2)

**FIG. 11.** Velocity field for two transverse sections in the homogeneous rotating case with $h/d = 1.95$, $\tau = 0.08\,\text{Pa}$ and $h = 20\,\text{m}$. Shadows represent the velocity $x$-component. Transverse component, scaled with max $|u|$, is showed with vectors. For visualization purposes $w$ has been exaggerated 200 times respect to $v$, and in channel midway magnitudes have been magnified 10 times respect to vectors at 2 km from the head $(\max|v| = 0.11\,\text{m}\,\text{s}^{-1}, \max|w| = 3.7 \times 10^{-4}\,\text{m}\,\text{s}^{-1})$

More noticeable differences relative to the non-rotating baseline case appeared in the other two velocity components, i.e., in the lateral and vertical flow. Midway along the domain, the lateral flow showed two distinct layers with the upper layer moving toward the right of the wind (as expected from Ekman dynamics) and a compensatory lower layer flowing in the opposite direction. Maximum $v$-values on the
surface Ekman layer were located at the transition between downwind and upwind flows, while relatively weak \( v \)-values were found near the lateral walls of the basin and in the middle of the cross-section (Fig. 11). The latter response occurred because downwind flow over the shoals induced lateral circulation toward the right of the wind due to Earth’s rotation. In the channel, Coriolis deflection on the upwind flow acted in the opposite direction (toward the left of the wind). At the head of the system, the lateral distribution of the \( v \)-component showed convergence toward the center of the channel as in the non-rotating case, but was not axially symmetric. Owing to the axially asymmetric lateral distribution of the \( u \)-velocity component, the water over the left shoal (where the flow was weaker) started to reverse toward the mouth at a greater distance from the head than the water over the right shoal (where the flow was stronger).

Away from the area influenced by the closed end, the vertical velocity component showed two upwelling and two downwelling cells related to divergence and convergence of lateral flow. As mentioned before, the convergence and divergence of the \( u \)-velocity component was relatively small (except near the head), while the transverse component showed convergence and divergence near the lateral walls and on both sides of the upwind flow. Along the basin left wall (looking toward the head) lateral flow caused upwelling while on the right wall it caused downwelling. The other convergence and divergence regions related to the upwind flow induced relatively weak upwelling and downwelling cells (Fig. 11). As in the non-rotating case, the vertical flow near the head was almost everywhere toward the bottom except in the middle of the cross-section, but in the rotating case it was not axially symmetric (Fig. 11). Maximum downward flows \((3.7 \times 10^{-4} \text{ m s}^{-1})\) were located over the right bathymetric slope as expected from the asymmetries of the lateral distribution of the \( u \)-component. Relatively weak (up to \(10^{-5} \text{ m s}^{-1}\)) upward flow was found in the middle of the cross-section at the head, associated with the divergence of the upwind flow.
Another important feature of the solution caused by the inclusion of the earth’s rotation was the free surface elevation distribution in the domain (Fig. 12). In the non-rotating baseline case, the surface elevation distribution was almost only a function of along-channel direction. According with Glorioso and Davies (1995), the across-channel barotropic pressure gradient expected in wind-induced flow over flat bottom rotating systems is nearly a linear function of $y$-location. The combination of both Coriolis effects and laterally varying bathymetry produced a more complicated lateral barotropic pressure gradient distribution (Fig. 12). Surface elevation contours showed stronger lateral variability compared to the non-rotating case, due to the spatial variability in the longitudinal flow and vertical mixing dictated by the laterally varying bathymetry. The spatial distribution of the modeled surface elevation was consistent with that obtained by Glorioso and Davies (1995).

In general, the magnitude of the $u$-velocity component was about the same order of magnitude for both non-rotating and rotating baseline cases. But the magnitude of the lateral and vertical velocity component of the rotating cases was larger than the non-rotating case in areas away from the closed ends. The cross-sectional average of the $|v/u|$ ratio evaluated midway along the basin was 0.01 for the non-rotating baseline case and 0.15 for the rotating baseline case. Because the lateral velocity component in rotating cases was non-divergent, the ratio $|v/u|$ was adopted to quantify axial asymmetries of the $u$-velocity component. Then, as $|v/u|$ increased the lateral distribution of $u$ became more asymmetric.

**Different cross-section**

A similar numerical experiment to the rotating baseline case was carried out over a Gaussian cross-section to illustrate the effects of the earth’s rotation on basins with a channel-shoal configuration. The solution over a Gaussian cross-section (Fig. 13)
FIG. 12. Contours of the surface elevation for the non-rotating (left) and rotating (right) baseline cases.

was qualitatively similar to that described in the triangular cross-section, i.e. similar location of upwind and downwind flow and similar transverse circulation. But, as in the non-rotating case, the Gaussian profile allowed a better definition of some features of the flow. The upwelling-downwelling areas at the lateral walls and around the upwind flow could be better distinguished than over the triangular section. Also the gyre associated with the upwind flow extended throughout the shoals. At the head of the system the left shoal showed two-layer circulation that reduced to downwind flow at around 10 km from the closed end.

In summary from the rotating baseline case, the lateral distribution of the flow was asymmetric. Strongest downwind flow developed on the right shoal (looking into the estuary). Asymmetries became more evident at the head of the system. Lateral and vertical velocity components appeared to be important at areas away from the closed ends, compared to non-rotating baseline case. The ratio $|v/u|$ increased by including Coriolis effects into the dynamics. The transverse component of velocity described three gyres. The main gyre was clockwise (looking downwind) and occupied the entire basin cross-section. Looking downwind, upwelling developed at the left boundary and downwelling appeared at the right boundary. The surface flow moved from left to right (to the right of the wind in the northern hemisphere) and bottom flow moved
from right to left. The other two gyres were small and localized and were linked to the channel edges, at the transition between upwind and downwind flow. This linkage reflected a convergence of the lateral velocity component around the zero $u$ isotach. Associated with the lateral convergence were vertical motions that closed the gyres.

**Different $h/d$ by changing $h$**

In order to analyze rotation influences in terms of different maximum water depth, two different numerical experiments were run with the same wind stress as in the baseline rotating case (0.08 Pa toward the head) but the maximum depth was chosen such that $h/d$ took values around 1 and 4.

In the $h/d \sim 4$ case, the bathymetric profile was triangular, with minimum depth of 3 m, maximum depth of 60 m. Increasing the maximum depth of the channel ($h$),
caused an increase in the magnitude of the $u$-velocity component compared with the baseline rotating case, as expected from the reduction of bottom stress effects throughout the water column. On the other hand, the ratio of the magnitude of the horizontal velocity components $|v/u|$ increased with respect to the baseline rotating case. The $|v/u|$ value for the $h/d \sim 4$ case was $\sim 0.38$ while for the $h/d \sim 2$ case was $\sim 0.15$. Then, increasing $h/d$ through an increment of the maximum depth of the channel ($h$), causes the lateral distribution of the $u$-velocity component to become more asymmetric compared to the $h/d \sim 2$ case (Fig. 14). The small gyres formed around the upwind core were intensified.

In the following experiment, reducing the maximum depth of the channel to 8 m caused the lateral distribution of the $u$-velocity component to become symmetric as in

FIG. 14. Same as Fig. 11 but $h/d \sim 3.54$, $\tau = 0.08$ Pa and $h = 60$ m ($\max |u| = 0.13 \text{ m s}^{-1}$, $\max |w| = 1.3 \times 10^{-3} \text{ m s}^{-1}$).

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the non-rotating case (Fig. 15). The lateral velocity component still showed a 2-layer structure. The magnitude of the \( u \)-component was reduced slightly compared with the baseline rotating case, which was forced with the same wind stress, as expected from increased bottom stress effects. The \( v \)-component was reduced by at least one order of magnitude compared with the baseline rotating case such that the \( |v/u| \) ratio for this rotating case was 0.08. As the system became shallower, the vertical shear stress became more important in the dynamics such that the main balance became that described in the non-rotating condition. This indicated that in the along-channel balance the Coriolis term became less important as the ratio \( h/d \) approached 1, which was consistent with the semi-analytical solution of Winant (2003). The noticeable reduction of the magnitude of the other two velocity components (\( v \) and \( w \)) was also consistent with decreased rotation influences.

### Different \( h/d \) changing \( d \)

Two different numerical experiments were run with the same bathymetry as in the rotating baseline case but the wind stress was chosen such that \( h/d \) took values of 1 and 4. The cross-section was triangular with minimum depth of 3 m and maximum depth of 20 m. By forcing the system with a weak wind stress of \( \tau = 0.005 \) Pa the ratio \( h/d \) took values of 4, i.e., the Ekman surface layer was limited to one quarter of the water column. The magnitude of all velocity components decreased (compared to the baseline rotating case) as expected from the reduction of the wind-stress. In contrast with the other \( h/d \sim 4 \) case described above, the asymmetries of the \( u \)-velocity component increased only slightly compared to the baseline rotating case (Fig. 16). The \( |v/u| \) ratio midway along the channel was only 0.2, while for the previous \( h/d \sim 4 \) case \( |v/u| \) was 0.38.

In the other experiment that changed \( d \), in order to approach the value of \( h/d \) of 1, the system was forced with a strong wind of 0.5 Pa. The Ekman surface layer
occupied the entire water column at all y-locations. The magnitude of all velocity components increased as expected from the increment of the wind stress, but the horizontal velocity ratio $|v/u|$ decreased to $\sim 0.09$. Lateral distribution of the $u$-velocity component became symmetric as in the previous $h/d \sim 1$ case (Fig. 17). Under these strong wind conditions, the lateral change of the $u$-velocity component became more important than in the previous cases. This was demonstrated by the $u$-velocity isolach separation.

Although the lateral gradient of the $u$-velocity component suggested that lateral mixing might be important in the $u$-momentum balance, lateral mixing was not included explicitly into the numerical model. First of all, because the selection of an appropriate value and/or function of the horizontal eddy viscosity is a difficult task.
Secondly because, as was mentioned in chapter 2, weak lateral mixing was included implicitly through the advection scheme used (Haidvogel et al. 2000).

In summary, variations of \( h/d \) through \( h \) and maintaining the same wind stress, cause maximum values of the \( v \)-velocity component to be found in the deepest case. The maximum values of the \( v \)-velocity component corresponded to the case where \( h \) was the largest, showing the relative importance of the earth’s rotating effects on the momentum balance. Also, the horizontal velocity components ratio \( |v/u| \) increased as \( h \) increased. On the other hand, variations of \( h/d \) through applications of different wind stress cause flows proportional to the forcing. The magnitudes of the \( v \)-velocity component also increased as the wind-stress increased, but the ratio \( |v/u| \) decreased, because vertical mixing became more important as the kinetic energy of the system.

FIG. 16. Same as Fig. 11 but \( h/d \sim 3.95, \tau = 0.05 \text{ Pa} \) and \( h = 20 \text{ m} \) (max \( |v| = 4.3 \times 10^{-3} \text{ m s}^{-1} \), max \( |u| = 1.1 \times 10^{-4} \text{ m s}^{-1} \)).
increased. Then, augmenting $h/d$ by increasing $h$ or by reducing the wind-stress, the Coriolis terms appear to be important to the momentum balance. Because the distribution of the $u$-velocity component is mainly laterally varying, lateral mixing might play an important role under strong-wind conditions such as the $h/d \sim 1$ case where wind-stress was 0.5 Pa.

The horizontal velocity components ratio was a useful parameter to estimate asymmetries. Comparing the two $h/d \sim 4$ cases, it was clear that both were more asymmetric than the rotating baseline case, and that the asymmetries were more evident in the case where $h/d$ was augmented by increasing $h$ than by decreasing $d$.

In general, inclusion of Coriolis accelerations and non-linear terms in the dynamics extended previous efforts on the topic of transverse structure of wind-induced circulation in laterally varying bathymetry basins in three main aspects: a) by showing
axial asymmetries of the flow are associated with the relative importance of Coriolis acceleration and friction, b) by studying the transverse circulation, and c) by noting the importance of considering spatial variations of internal friction (vertical shears appeared over the shoals and lateral shears developed in the channel).

III.3.2.2 Particle trajectories (comparison with non-rotating case)

To analyze the influence of Earth's rotation on particle trajectories, the non-rotating case was compared to two rotating cases: \( h/d \sim 2 \) and \( h/d \sim 4 \). The first rotating case \( (h/d \sim 2) \) had the same depth as the non-rotating baseline case. The second rotating case \( (h/d \sim 4) \) was most affected by the earth's rotation \( (h = 60 \, \text{m}) \). In both cases, four particles were released in the same initial locations as in the non-rotating case. The particles were released at the entrance to the system. The initial vertical positions were 2 m depth and one sixth of the maximum depth, while the initial lateral positions were 1 km away from each lateral side of the basin. Trajectories are presented in Figs. 18 and 19. Left panels illustrate horizontal views \((x-y)\) and right panels illustrate cross-section views \((y-z\) looking downwind) of the four particle trajectories.

In general, inclusion of the Coriolis acceleration in the numerical solution, causes particle trajectories to become much more complicated than those without rotation. The four particle trajectories illustrate the importance of the initial location of the particle on the path and time that the particle will take to leave the system. The trajectories of the particles released near the upwelling wall (i.e. left shoal, looking into the system) were longer compared to particles released near the downwelling wall. Particles released near the downwelling wall sank and moved toward the deep channel where they were incorporated to the upwind flow (flow toward the entrance). Particles released near the upwelling wall, moved toward the head of the system in the surface layer, where upwind flow is minimum. The particles sank when they arrived
at the head or the downwelling (right) side of the system and then began to move toward the entrance.

Not only the lateral initial position played an important role on the trajectory of the particle but also the initial vertical location. Particles released at the surface remained longer in the system than particles released deeper. A particle released at the surface near the downwelling wall took longer to sink than a particle released deeper. The latter sank and moved toward the entrance by the upwind flow of the middle of the cross-section.

In summary, the trajectories of the non-rotating case contrasted with the rotating cases in the sense that initial position across the channel and depth of release yielded dramatically different trajectories. Most of the particle trajectories became more complicated as the ratio $h/d$ increased (Figs. 18 and 19). Therefore, it was clear that rotation had a tremendous influence on the path followed by a particle and the time it remained in the system.

III.3.3 Numerical results vs. analytical solutions

Comparison to analytical solutions

The numerical results described above showed similarities and differences with the linear semi-analytical solution presented by Winant (2003). In both semi-analytical (Winant 2003) and numerical solutions, the transverse component of velocity described a clockwise gyre (looking downwind) that showed upwelling on the left wall and downwelling on the right wall. This gyre occupied the entire cross-section and resulted from rotation effects on wind-induced flow constrained by basin walls. Also in both numerical and analytical solutions, the lateral distribution of the rotating $u$-velocity component approached the non-rotating solution when the ratio $h/d$ approached 1 and became axially asymmetric as $h/d$ increased.

The inclusion of all terms in the momentum equations with the eddy viscosity
varying in time and space, however, yielded a few more features in the numerical solution, relative to Winant's solution. The numerical solution resolved two small-scale gyres in the $y$-$z$ plane linked to the lateral distribution of the along-channel velocity component, which in turn was dictated by bathymetry. This was true for all the rotating experiments. In addition, examination of two different cases for each $h/d$ value, through variation of $h$ or $d$, showed that the lateral distribution of the $u$-velocity component became more asymmetric (as characterized by increased $|v/u|$ ratios) as the maximum depth of the channel $h$ increased. This was in contrast to decreasing $d$ through weaker wind stress. Earth's rotation effects, manifested by enlarged asymmetries or increased $|v/u|$ ratios, were noticeable for the cases with $h/d > 2$. In cases in which $h/d = 1$ over relatively deep channels (e.g. $\sim 20$ m, Fig. 17), the wind stress was such that it drove $u$-velocities with large lateral shears. The largest lateral shears were localized along the partition between upwind and downwind flows.
FIG. 18. Particle trajectories released from different locations (indicated with the solid triangle) in the steady state for the homogeneous case with rotation, $h/d \sim 4$ and $h_0 = 60$ m. Left panel $x$-$y$ plane, right panel $y$-$z$ plane. Solid line indicates the incoming path, dotted line signals the outgoing one.
FIG. 19. Particle trajectories released from different locations (indicated with the solid triangle) in the steady state for the homogeneous baseline case with rotation, $h/d \sim 2$ and $h_0 = 20$ m. Left panel $x$-$y$ plane, right panel $y$-$z$ plane. Solid line indicates the incoming path, dotted line signals the outgoing one.
CHAPTER IV

WIND-DRIVEN FLOW IN STRATIFIED SYSTEMS

IV. 1  INTRODUCTION

Observational studies in different coastal plain estuaries have shown appreciable transverse variability of the salinity distribution and of the residual circulation induced by wind and buoyancy forcing (e.g. Wong 1994). Bathymetry has also been shown to play an important role in the transverse variability of both wind-driven and buoyancy-driven flow (Valle-Levinson et al. 2001a; Valle-Levinson et al. 2001b; Valle-Levinson et al. 2003a). Observational studies related to the competition between wind-driven and buoyancy-driven flow are scarce. Valle-Levinson et al. (2003b) present one of the few observational examples of wind-induced circulation under weak vertical mixing (weak tidal forcing) and where transverse flows reflect the relative importance of the earth's rotation compared to friction effects. Their data consisted of hydrographic and water velocity profiles measured along the entrance to the Bay of Concepcion, in Chile. The surveys took place during neap tides and under moderate wind forcing (0.07 Pa). Through a scaling analysis they showed that although the wind-induced flow overwhelmed the gravitational circulation, the wind-stress had a direct influence only in the upper part of the water column. As mentioned in Valle-Levinson et al. (2003b), the wind-forced surface layer moved downwind and the transverse flow responded to the earth's rotation. This means that the transverse flow component of the surface layer was toward the left of the wind as expected in the southern hemisphere. The location of the core of maximum outflow and maximum inflow at the entrance of Bay of Concepcion also reflected the influence of the earth's rotation.

Analytical and numerical models related to stratified flows over bathymetry have mostly concentrated on the lateral structure of density-driven longitudinal flow. Also,
analytical models have considered wind and buoyancy forces separately. In this investigation, the interaction between these two forces was explored. This study focused mainly in two aspects, 1) the basic pattern of the local wind-induced currents in a stratified system with laterally varying bathymetry, and 2) the role of wind-induced mixing on the buoyancy-induced circulation pattern. Emphasis was placed on the lateral distribution of all velocity components resulting from the interaction of wind-induced and buoyancy-driven flow. To address these questions, a series of numerical experiments was carried out over a stratified estuary with transverse variations in bathymetry.

IV.2 NUMERICAL EXPERIMENTS

A total of 16 numerical experiments were carried out to study the effects of wind forcing on stratified systems with lateral depth variation (Table 3). The analysis consisted of two parts: one focused on the non-rotating interaction between wind-driven and density-induced flow (eight experiments), and a second focused on the earth's rotation effects in stratified systems influenced by wind-stress (other eight experiments).

In order to compare the results with analytical solutions and to assess the model performance, the first four experiments (1 to 4) of the group without rotation examined the sensitivity of gravitational circulation to vertical mixing. For that purpose different parameterizations of eddy viscosity and diffusivity were used. The following four experiments (5 to 8) of the non-rotating group, looked at the interaction between wind-induced and density-driven flow. The system was forced with the same longitudinal density gradient but different wind stresses. The relative importance of the wind-stress in the dynamics of buoyancy-driven flow was assessed with the non-dimensional Wedderburn number \( W_e = l \tau / (gh^2 \Delta \rho) \) (e.g. Geyer 1997), where \( l \) is
the total length of the estuary, $L$ is its mean depth and $\Delta \rho$ is the along channel density variation. Wedderburn number compares accelerations produced by wind stress and baroclinic pressure gradient. The magnitude of the wind was chosen such that the values of the Wedderburn number were $< 1$, $\sim 1$, and $> 1$. For the second set of experiments (those with rotation), four (experiments 9 to 12) looked at the influence of the Ekman number on the lateral distribution of the longitudinal-flow and salinity field (as explored by Kasai et al. 2000). Similar to the non-rotating cases, the longitudinal density gradient was the same for these experiments, but different eddy viscosity parameterizations were used. The rest of the experiments (13 to 16) explored the influence of Coriolis effects on wind-induced circulation in non-homogeneous systems. The wind stresses applied were the same as in the non-rotating stratified cases.

As mentioned before, all experiments started from rest with an initial longitudinal density gradient of $10^{-4}$ kg m$^{-4}$, typical value for coastal plain estuaries (e.g. Valles-Levinson and O'Donnell 1996), and were forced with a constant wind stress after 30 h of gravitational adjustment. Also a constant river discharge (100 m$^3$ s$^{-1}$) at the landward boundary was applied. Experiments 1 to 4 and 9 to 12 were also performed with the purpose of comparing previous findings to the performance of the model. The rest of the numerical experiments constitute new contributions to the understanding of the competition between wind-driven and buoyancy-driven flows over bathymetry.

IV. 3 RESULTS

This section presents results obtained for a gravitational adjustment over laterally varying bathymetry and for the interaction of wind-driven and buoyancy-induced flow in non-rotating and rotating systems. Gravitational circulation was evaluated by using a specific turbulence closure scheme. Sensitivity of the gravitational circulation to vertical mixing typically induced by tides and/or wind was explored by using three different constant eddy viscosity values. Rotation influence was characterized as in
Kasai et al. (2000), as a function of Ekman number \( \left( E = \frac{A_z}{fH_0} \right) \), or equivalently \( d/h \).

The results were analyzed at a cross section midway along the basin and presented in figures that portray views landward.

IV.3.1 Non-rotating system. Gravitational adjustment

IV.3.1.1 Gravitational adjustment

**Baseline case: Sensitivity analysis to vertical mixing values**

The gravitational adjustment over laterally varying bathymetry was explored using the KPP vertical turbulent closure scheme (Large et al. 1994). This vertical closure scheme allows vertical mixing to vary in space and in time (i.e. \( A_z = f(x, y, z, t) \)). Forcing the system only with a longitudinal density gradient of order of \( 10^{-4} \text{ kg m}^{-4} \) the longitudinal circulation and salinity field showed well-defined vertical gradients (Fig. 20), similar to those expected over flat bottom (Pritchard 1956; Wong 1994). This behavior could be explained by weak vertical mixing that did little to disrupt the vertical density gradient. The cross-sectional average of the vertical eddy viscosity was \( \sim 8 \times 10^{-5} \text{ m}^2\text{s}^{-1} \) and the average Richardson number was \( \sim 4.8 \). It is clear from these results that even in laterally varying bathymetry, the gravitational circulation showed vertical gradients in velocity and salinity fields. These results contrasted with the analytical solution reported by Wong (1994) that shows transverse gradients rather than vertical. The reason for such discrepancy is explained next.

In order to analyze the influence of variations in turbulence on gravitational circulation (specific viscosity background assumed to be of wind and/or tidal origin), three different numerical experiments were carried out by using the same conditions as in the previous case but constant eddy viscosity \( (A_z) \) and eddy diffusivity \( (K_S, K_T) \). Constant eddy viscosity coefficients were \( 10^{-2} \text{ m}^2\text{s}^{-1} \), \( 10^{-3} \text{ m}^2\text{s}^{-1} \) and \( 5 \times 10^{-5} \text{ m}^2\text{s}^{-1} \) (similar to those used by Kasai et al. 2000; and to those reported by Wong (1994).
FIG. 20. Velocity and salinity fields for midway transverse section for the non-rotating gravitational adjustment problem using KPP closure scheme. In the upper panel, shadows represent the velocity $x$-component. Transverse component, scaled with max $|u|$, is showed with vectors. For visualization purposes $w$ has been exaggerated 200 times respect to $v$ (max $|v| = 3.0 \times 10^{-2}$ m s$^{-2}$, max $|w| = 1.5 \times 10^{-4}$ m s$^{-2}$)

as typical values for Delaware Bay), while $K_S$ and $K_T$ values used were an order of magnitude smaller than the corresponding $A_z$. The last value is similar to the cross-sectional mean obtained in the KPP numerical experiment.

**High vertical mixing ($A_z = 10^{-2}$ m$^2$s$^{-1}$)**

Under relatively high vertical mixing conditions represented by this eddy viscosity, the longitudinal flow and salinity fields showed lateral variability related to bathymetry. This variability is predicted by linear analytical models (e.g. Wong 1994) and also has been observed in non-homogeneous shallow systems (Wong 1994). In general, outflow of fresher water occurred over the shoals and inflow of saltier water appeared in
the channel (Fig. 21). Both longitudinal flow and salinity fields showed strong lateral gradients, as seen in the vertical distribution of isotachs and isohalines. Maximum inflow values were found mid-water in the channel. This lateral partition of the flow is basically caused by the depth-dependent nature of the baroclinic pressure gradient (Wong 1994). The lateral distribution of the along-channel flow and salinity field induced a lateral baroclinic pressure gradient that, in turn, induced axial convergence at the surface and a compensated divergence underneath. The vertical velocity component showed upward and downward flow related to the divergence the lateral flow and the presence of the basin lateral walls. Lateral flows were, in general, one order of magnitude smaller than the longitudinal velocity component. The vertical velocity component was at least four orders of magnitude smaller than the horizontal flow. A special feature of these results is the axial symmetry of all velocity components and
of the salinity field.

*Intermediate vertical mixing* \( (A_z = 10^{-3} \text{m}^2\text{s}^{-1}) \)

Keeping the same conditions as in the previous case but decreasing the vertical eddy viscosity \( (A_z) \) caused isohalines and isotachs to show a combination of lateral and vertical gradients in the cross-section (Fig. 22). In general, the lateral distribution of the longitudinal velocity component and salinity showed similar cross-section distributions as in the previous case. Outflow of fresher water appeared over the shoals and inflow of saltier water in the channel. The salinity field showed weaker transverse salinity gradients in the upper part of the water column than in the previous case \((\sim \text{first } 8 \text{ m})\). Below 8 m depth, the salinity field was vertically stratified. Longitudinal flow over the shoals showed gradients with weakest flow near the bottom, as expected from bottom friction. In the channel, maximum inflow values were concentrated in the deep part. Over the top 8 m, lateral flows showed two-layer circulation with convergence at the surface and divergence underneath. Below 8 m (where the salinity field was vertically stratified), the lateral flow was near zero. It is noteworthy that, although the lateral salinity gradient was weaker than in the previous case with \( A_z = 10^{-2} \text{m}^2\text{s}^{-1} \), the lateral flow was the same order of magnitude owing to the reduction of vertical mixing. In general, the magnitude of horizontal flows increased relative to the previous experiment with a constant eddy viscosity coefficient, as expected from the reduction of vertical mixing. In other words, less kinetic energy was converted into potential energy. Lateral and vertical flow components remained, respectively, one and four orders of magnitude smaller than the along-channel component. Lateral distribution of all velocity components and the salinity field still showed axial symmetry as in the previous case.

*Weak vertical mixing* \( (A_z = 5 \times 10^{-5} \text{m}^2\text{s}^{-1}) \)

Further reduction of vertical eddy viscosity \( (A_z = 5 \times 10^{-5} \text{m}^2\text{s}^{-1}) \) caused both salinity
and $u$-velocity component to show marked vertical gradients (Fig. 23) as in the numerical experiment where the KPP closure was used. The lateral distribution of the along-channel flow and salinity field exhibited the characteristic behavior of estuarine circulation over a flat bottom. Outflow of fresher water appeared at the surface and inflow of saltier water developed near the bottom. In general, the magnitude of the along-channel velocity component increased, as expected from the reduction of vertical mixing. The across-channel and vertical velocity components decreased almost to zero, because of the reduction of the across-channel baroclinic pressure gradient.

In summary, the lateral distribution of the buoyancy-driven flow was markedly sensitive to vertical mixing. This was varied using three different constant eddy viscosities ($A_z$) and the KPP-closure scheme. Circulation and salinity fields in the buoyancy-driven flow range from strong lateral variability for high $A_z$ values (strong mixing),
to strong vertical structure as $A_z$ decreased (weak mixing). These suggested that the circulation pattern developed in an estuary can ranged from lateral structure to vertical structure depending on the kinetic energy budget, as has been reported from observational data at the Chesapeake Bay entrance (e.g. Reyes-Hernandez 2001). These numerical results agreed with Kasai et al. (2000) analytical results (who characterized the lateral distribution of the flow as a function of Ekman number), in the sense that as vertical mixing increased the flow pattern is dominated by transverse gradients. By use of the KPP-closure scheme, the gravitational circulation over laterally varying bathymetry showed vertical gradients of longitudinal flow and salinity fields, similar to estuarine circulation over flat bottom. In other words, in the absence of an external kinetic energy source (other than that produced by bottom friction) the gravitational circulation was dominated by vertical gradients.

From the sensitivity analysis of the gravitational adjustment to the vertical eddy viscosity values, it was clear that the circulation in non-rotating systems became mainly longitudinal as vertical mixing decreased, owing to the weakening of the lateral baroclinic pressure gradient. A suggestion of how the transverse circulation might affect the mass exchange was illustrated by particle trajectories in section IV.3.2.3.

In summary, these results indicate that not only bathymetry but also the strength of vertical mixing plays an important role on the lateral variation of the flow and salinity fields in the gravitational adjustment. Increased vertical mixing caused gravitational circulation to be dominated by transverse gradients related to inflow in the deep channel and outflow over the shoals. These experiments were carried out only to investigate the consistency of the numerical approach with previous solutions and showed the satisfactory performance of the model.
FIG. 23. Same as Fig. 20 but using a constant $A_z = 5 \times 10^{-5} \text{m}^3\text{s}^{-1}$ and $K_S, K_T = 5 \times 10^{-6} \text{m}^2\text{s}^{-1}$ ($\max |v| = 2.0 \times 10^{-2} \text{m}\text{s}^{-1}$, $\max |w| = 3.6 \times 10^{-4} \text{m}\text{s}^{-1}$).

IV.3.1.2 Wind effect on gravitational adjustment

Four different numerical experiments were carried out to analyze the influence of wind-driven flow over a stratified system. In all of the experiments, constant wind stress and constant input of fresh water at the head of the system were applied over the basin and the KPP-closure scheme was used. The wind stress was aligned with the x-axis. In the first three experiments the wind blew toward the head of the system (i.e. counteracting the gravitational circulation) and for the last experiment the wind blew toward the mouth. Results were analyzed for four different wind-conditions ($\tau = \pm 0.01 \text{Pa}, 0.1 \text{Pa}, 0.2 \text{Pa}$), such that the Wedderburn number ($W_e$) took values lower than 1, near 1 and larger than 1. River discharge was, as in the gravitational adjustment cases, 100 $\text{m}^3\text{s}^{-1}$. Results were analyzed at $x = 50 \text{ km}$ (mid-domain).
and after two days that the maximum wind-stress had been reached, at quasi-steady conditions. These conditions were determined when the local temporal variation of $u$ was at least one order of magnitude smaller than the pressure gradient.

![Channel midway](image)

**FIG. 24.** Velocity and salinity fields for midway transverse section in non-rotating system with landward wind stress ($\tau$) of 0.01 Pa. Representation characteristics are the same as in Fig. 20, ($\max |v| = 3.2 \times 10^{-2} \text{ m s}^{-1}$, $\max |w| = 3.5 \times 10^{-4} \text{ m s}^{-1}$).

**Weak winds**

Under weak wind conditions ($W_e \sim 0.01$) buoyancy-driven flow dominated the quasi-steady state circulation. The lateral distribution of salinity and along-channel velocity component showed vertical gradients that reflected weak mixing (Fig. 24). The cross-sectional mean eddy viscosity coefficient at $x = 50 \text{ km}$ was $1.4 \times 10^{-4} \text{ m}^2\text{s}^{-1}$. The surface layer (from surface to 4 m depth) showed weak lateral salinity gradients with
fresher water over the shoals and saltier water in the center of the cross-section. Below the surface layer, the salinity field was strongly stratified except in the deepest part of the channel, where it was mixed. Wind forcing caused the maximum outflow to appear below the surface instead of at the surface. The maximum outflow was reduced from $\sim 0.30 \text{ m s}^{-1}$ (gravitational circulation without wind) to $\sim 0.11 \text{ m s}^{-1}$, in consequence, the zero $u$ isotach sank to keep the mass balance. Weak lateral flow developed from lateral density gradients and was at least one order of magnitude smaller than the longitudinal velocity component. This lateral flow showed axial convergence at the surface and compensating divergence in the layer below. The vertical velocity component was at least 4 orders of magnitude smaller than the horizontal flow. Therefore, the influence of weak winds on gravitational circulation is restricted only to a thin surface layer.

**Moderate winds**

Under moderate wind conditions ($0.1 \text{ Pa, } W_e \sim 1.29$), vertical mixing increased such that the momentum and salinity fields showed weak vertical gradients (Fig. 25). The average eddy viscosity over the cross-section at $x = 50 \text{ km}$ was $\sim 1.8 \times 10^{-3} \text{ m}^2 \text{s}^{-1}$. In this case, wind-driven circulation overwhelmed buoyancy-induced circulation. The lateral distribution of the longitudinal flow and salinity fields featured a transverse structure with downwind inflow of saltier water over the shoals and upwind outflow of fresher water in the channel. The downwind flow averaged $0.12 \text{ m s}^{-1}$, while the upwind flow averaged $0.1 \text{ m s}^{-1}$. This pattern was opposite to the buoyancy-induced flow (e.g. Wong 1994) and, to the best of my knowledge, it has not been described in natural environments. This result will hopefully motivate observations aimed at describing it.

The lateral velocity component was smaller but the same order of magnitude as the longitudinal velocity component, owing to the increase of lateral salinity gradients. The transverse circulation showed two gyres related to axial divergence at the
surface and convergence at the layer below and downward flow at the lateral basin walls. This pattern of transverse circulation was also opposite to the buoyancy-driven flow under no wind forcing (e.g. Nunes and Simpson 1985). It has not been described in natural systems either and will hopefully motivate future observations. This experiment showed that moderate winds ($W_e > 1$) are expected to completely reverse the buoyancy-induced exchange flow.

![Diagram](channel_midway)

**FIG. 25.** Same as Fig. 24 but for $r = 0.1 \text{ Pa}$. ($\max |v| = 0.1 \text{ m s}^{-1}$, $\max |w| = 8.2 \times 10^{-4} \text{ m s}^{-1}$).

*Strong winds*

Keeping the same conditions as in the previous case, but increasing the wind-stress to 0.2 Pa, $W_e$ took values of $\sim 2.59$ and the cross-sectional average of the eddy viscosity coefficient, at $x = 50 \text{ km}$, took values of $\sim 2.7 \times 10^{-3} \text{ m}^2\text{s}^{-1}$. The transverse structure

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of both salinity and along-channel component were maintained but the magnitude of the flow increased (Fig. 26), i.e., cross-sectional average of downwind flow was 0.2 m s\(^{-1}\) and cross-sectional average of upwind flow was 0.15 m s\(^{-1}\). The pattern of seaward advection of low salinity in the channel and landward advection of salty water over the shoals was similar to the case of moderate winds. The magnitude and the relative importance of the lateral and vertical velocity component increased, in terms of the longitudinal velocity component. The transverse component described two axially symmetric gyres as in the previous case. This transverse circulation is slightly stronger than under moderate winds and might modify the flushing time of the system, as discussed in the next section.

FIG. 26. Same as Fig. 24 but for \(\tau = 0.2\) Pa. (\(\max|v| = 0.18\) m s\(^{-1}\), \(\max|w| = 2.2 \times 10^{-3}\) m s\(^{-1}\)).
Down-estuary weak winds

Under down-estuary wind conditions the buoyancy-induced and wind-driven flows were additive. Applying a weak wind-stress ($W_e \sim 0.01$), the buoyancy-induced circulation dominated the dynamics. The main effect of the down-estuary wind stress was to increase the magnitude of the maximum outflow and inflow (Fig. 27). In general, the velocity and salinity lateral distributions were similar to the case of weak up-estuary wind stress, but the water exchange was slightly enhanced. Maximum outflow (located at the surface) and inflow were 0.40 m s$^{-1}$ and 0.35 m s$^{-1}$, respectively. In the weak up-estuary wind case maximum outflow (located underneath the surface) was $\sim 0.11$ m s$^{-1}$ and maximum inflow was $\sim 0.20$ m s$^{-1}$.

In summary, the circulation pattern and salinity distribution induced by wind over a stratified system showed that wind forcing dominated over buoyancy influences when the Wedderburn number ($W_e$), is near or exceeds 1. For weak up-estuary wind ($W_e \ll 1$), the gravitational circulation remained almost unaltered. Only the upper part of the water column became vertically homogeneous and the maximum outflow migrated to the subsurface. Under strong winds ($W_e \geq 1$) the wind-induced pattern of downwind flow over shoals and up-wind flow in the channels masked the gravitational circulation and the water column remained nearly vertically homogeneous. Transverse circulation showed two symmetric gyres with divergence at the surface. These gyres were produced by the lateral baroclinic pressure gradient, which in turn resulted from differential advection of salinity by the lateral distribution of the longitudinal velocity component. This is the first time that the exchange flow pattern that results from the competition between wind forcing and buoyancy forcing has been described. Future observations should be designed to document these patterns.
FIG. 27. Same as Fig. 24 but for seaward wind stress of 0.01 Pa (max $|v| = 0.03$ m s$^{-1}$, max $|w| = 7.1 \times 10^{-4}$ m s$^{-1}$).

IV.3.1.3 Particle trajectories

Particles trajectories were described for two different cases of gravitational adjustment that have constant eddy viscosity ($10^{-3}$ m$^2$ s$^{-1}$ and $5 \times 10^{-5}$ m$^2$ s$^{-1}$). Each particle trajectory was evaluated through temporal integration of the velocity field. Four particles were released midway along the system with initial vertical positions of $z_0 = 2$ m and $z_0 = 5$ m, and initial lateral positions of 2 km and 8 km. Figures 28 and 29 illustrate the trajectory of the four particles (for each $A_z$ value), where the left panels show horizontal views and the right panels show transverse views.

For the case of moderate mixing ($A_z = 10^{-3}$ m$^2$ s$^{-1}$), the trajectory showed dependency on the released vertical location (Fig. 28). The particles released at 5 m depth moved seaward much faster than particles released at 2 m depth, but in general
particles moved toward the ocean-open boundary in a helical path. This is, particles converged toward the channel, sank and then diverged toward the shoal with a net seaward displacement. For the case of weak vertical mixing ($A_z = 5 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$), particles remained approximately at the depth of release (Fig. 29). The trajectory showed dependence only on the depth of release. Particles released at the surface left the system following a nearly straight path. Particles released at 5 m depth showed weak convergence toward the channel with a net seaward displacement. Particle trajectories were much more complicated in the case of strong vertical mixing owing to the lateral baroclinic pressure gradient that induced lateral convergence and divergence. These divergences/convergences prolonged the time that particles remained in the system. It is essential to note that under no rotation, particle trajectories were axially symmetric.
FIG. 29. Particle trajectories for the gravitational adjustment problem using constant eddy viscosity ($A_z = 5 \times 10^{-5}$ m$^2$ s$^{-1}$) in a non-rotating system, released from different locations (indicated with the solid diamonds). Left panel $x$-$y$ plane, right panel $y$-$z$ plane.

IV.3.2 Rotating system

IV.3.2.1 Gravitational adjustment

The effects of the earth’s rotation were studied first for the gravitational adjustment over laterally varying bathymetry (using the KPP vertical mixing scheme) to assess the model performance under these forcing conditions. The initial conditions were similar to those in the non-rotating case (longitudinal density gradient $10^{-4}$ kg m$^{-4}$). Under the effects of Earth’s rotation the gravitational adjustment showed isohalines and longitudinal-component isotachs tilting upward to the right (looking landward) similarly to the response over flat bottom systems (Pritchard, 1956). Maximum outflow of fresher water was concentrated over the left shoal and maximum inflow of saltier water was concentrated at mid-water in the channel (Fig. 30). In this case...
FIG. 30. Velocity and salinity fields for midway transverse section for the rotating gravitational adjustment problem using KPP closure scheme to evaluate the eddy viscosity and eddy diffusivity. In the upper panel, shadows represent the velocity x-component. Transverse component, scaled with \( \max |u| \), is showed with vectors. For visualization purposes \( w \) has been exaggerated 200 times respect to \( v(\max |v| = 4.0 \times 10^{-2} \text{m s}^{-2}, \max |w| = 1.8 \times 10^{-4} \text{m s}^{-2}) \)

the Ekman number \( E \) was 0.002 evaluated at the cross-section in the middle of the domain. The results agreed with Kasai et al. (2000) analytical solutions in the sense that inflow did not reach the surface due to the relatively low Ekman number. These numerical results were more realistic than the analytical solution because in the latter the flow was axially symmetric. In general, the outflow of fresher water had a lateral velocity component toward the left (looking into the system) while the inflow of saltier water had a lateral velocity component toward the right. This reflected the influence of the earth's rotation on the longitudinal pressure gradient and under weak friction, i.e., a lateral balance dominated by geostrophy. The asymmetries shown by the lateral distribution of the longitudinal flow and salinity fields, as well as the
lateral flow structure, were as those predicted in Valle-Levinson et al. (2003a). These numerical results showed that even in laterally varying bathymetry the gravitational circulation, when influenced by weak vertical mixing, featured the typical two-layer estuarine circulation described in flat bottom rotating systems.

Sensitivity analysis of the gravitational circulation to vertical mixing was performed including the earth’s rotation. This was done in order to further test the model performance and to compare the numerical results to analytical solutions. Same constant eddy viscosity values ($A_z = 10^{-2} \text{m}^2 \text{s}^{-1}, 10^{-3} \text{m}^2 \text{s}^{-1}$ and $5 \times 10^{-5} \text{m}^2 \text{s}^{-1}$) and eddy diffusivity ($K_S, K_T = 10^{-3} \text{m}^2 \text{s}^{-1}, 10^{-4} \text{m}^2 \text{s}^{-1}$ and $5 \times 10^{-6} \text{m}^2 \text{s}^{-1}$), as in the non-rotation cases, were used. The influence of the earth’s rotation was characterized.
with the Ekman number \((E)\) as in Kasai et al. (2000).

Using a high vertical eddy viscosity coefficient \((A_z = 10^{-2} \text{m}^2 \text{s}^{-1})\), the Ekman number took values of 0.25. Under these vertical mixing conditions, salinity and velocity fields featured similar lateral distribution to those described in the non-rotating system (Fig. 21). For a relative large Ekman number, the main dynamic balance was between pressure gradient and friction. Gravitational circulation was dominated by a transverse structure with outflow of fresher water over the shoals and inflow of saltier water elsewhere (Fig. 31). The lateral momentum balance was also influenced by strong vertical mixing, such that it was ageostrophic. Transverse flows showed two axially symmetric gyres, describing convergence toward the channel at the surface and divergence below that, as expected from the baroclinic pressure gradient developed (i.e. fresher water was concentrated at the center of the cross-section). Downward flow in the channel and upward flow at the shoals, owing to the lateral basin walls, closed the gyres. This transverse circulation pattern agreed with the analytic model reported by Nunes and Simpson (1985) and Swift et al. (1996), where the lateral momentum balance excluded the earth’s rotation and the pressure gradient was only balanced by internal friction. Even though the prescribed \(E_k\) was 0.25, rotation effects were negligible in this solution.

Decreasing the constant vertical eddy viscosity to \(10^{-3} \text{m}^2 \text{s}^{-1}\), the Ekman number became 0.025. For this case, gravitational circulation still dominated by a transverse structure but salinity and longitudinal velocity component showed axial asymmetries. Stronger outflow of fresher water was concentrated over the left shoal, and inflow of saltier water (which reached the surface) concentrated in the channel (Fig. 32). The inflow core moved toward the right. Similar to the high \(E\) case, transverse circulation showed two gyres, but the convergence and divergence areas moved toward the right following the inflow core. The gyre that developed over the right shoal showed slightly larger lateral velocity magnitudes than the gyre over the left shoal. This was because
FIG. 32. Same as Fig. 30 but using a constant $A_z = 10^{-3} \text{m}^2\text{s}^{-1}$ and $K_S, K_T = 10^{-4} \text{m}^2\text{s}^{-1}$ ($\max|v| = 3.0 \times 10^{-2} \text{m s}^{-1}$, $\max|w| = 1.3 \times 10^{-4} \text{m s}^{-1}$).

the Coriolis force and the pressure gradient force (in the lateral momentum balance) acted in the same direction to reinforce the near-surface flow to the right. These forces acted in opposite direction over the left shoal, i.e., Coriolis to the right and pressure gradient to the left. Although the lateral momentum balance was dominated by the pressure gradient force and friction, Coriolis effects allowed axial asymmetries in the lateral circulation too.

In the low $E$ case ($A_z = 5 \times 10^{-5} \text{m}^2\text{s}^{-1}$ and $E \sim 0.001$), velocity and salinity fields were similar to those obtained in the KPP rotating case (Fig. 33), that is, longitudinal velocity component and salinity fields were similar to those that have been reported in flat-bottom stratified rotating systems.
In summary, the circulation pattern of gravitational adjustment in a rotating system was found to be strongly dependent to vertical mixing or/and to the Ekman number. The circulation pattern was dominated by lateral structure as the Ekman number increased. Also, the circulation pattern became two-layered as the Ekman number decreased. These numerical results agreed with Kasai et al. (2000) analytical solution in the sense that the inflow of saltier water reaches the surface only with relatively high Ekman numbers. The inflow-core of saltier water remains detached from the surface for low Ekman numbers. An important characteristic that appeared in the numerical results, and differs from Kasai et al. (2000), is the axial asymmetry of the flow that developed in response to the Coriolis acceleration. Kasai et al.
(2000) solution is axially symmetric, independently of the Ekman number. The numerical solutions presented here were symmetric only for high Ekman numbers, i.e., when Coriolis force was completely overwhelmed by frictional forces. The numerical results showed that as Ekman number decreased, not only the flow and salinity pattern showed vertical gradients, but also the two fields showed axial asymmetries. These asymmetries had important implications on particle trajectories and, in turn, in residence time as described at the end of this chapter. The transverse circulation for high Ekman number cases showed two axially symmetric gyres that developed over the shoals, similarly to those described by Nunes and Simpson (1985) and Swift et al. (1996), where rotation was neglected. As the Ekman number decreased, the transverse circulation approached a geostrophic balance showing two-layer lateral circulation instead of symmetric gyres. These results are similar to those reported by Valle-Levinson et al. (2003a).

**IV.3.2.2 Wind effect in gravitational adjustment**

Four different numerical experiments were carried out to analyze the influence of the earth’s rotation on the interaction of wind-induced and buoyancy-driven flow. The initial condition of the longitudinal density gradient, wind stress and river discharge were similar to those in the non-rotating cases (sec. IV.3.2). Applying a wind stress of ±0.01, 0.01, 0.1 and 0.2 Pa, \( W_e \) took values lower than 1, near 1 and larger than 1, respectively.

*Weak winds*

For the lowest \( W_e \) case (\( \tau = 0.01 \) Pa), the gravitational adjustment was very slightly modified by the wind-induced flow. Similarly to the non-rotating case (with similar \( W_e \)), the effect of the wind-stress was more noticeable in the upper part of the water column where the magnitude of the outflow was reduced and the maximum outflow moved below the surface. Comparing the non-rotating and rotating cases (Figs. 24
and 34) it is clear that the inclusion of the earth’s rotation induced axial asymmetries of the velocity and salinity fields. For this experiment, the Ekman number was 0.003 and reflected the importance of Coriolis effects over vertical mixing. The lateral circulation was similar to that in the rotating gravitational adjustment problem described in the previous section (KPP-case, Fig. 30), but the thinner surface layer affected by the wind-stress showed lateral circulation toward the right of the wind, as expected from the earth’s rotation influence (Fig. 34).

![Channel midway](image)

**FIG. 34.** Velocity and salinity fields for midway transverse section in rotating system with landward wind stress ($\tau$) of 0.01 Pa. Features are the same as in Fig. 20, ($\max |u| = 5.8 \times 10^{-2} \text{ m s}^{-1}$, $\max |w| = 1.6 \times 10^{-4} \text{ m s}^{-1}$).

*Moderate winds*

For the case where $W_e$ was near 1 ($\tau = 0.1 \text{ Pa}$), wind-driven flow masked the...
buoyancy-induced flow (as in the non-rotating case with similar $W_e$ value) and the Ekman number took the value of 0.035. Buoyancy-induced flow was reversed by the wind, such that the lateral distribution of the longitudinal velocity component showed inflow of saltier water over the shoals and outflow of fresher water concentrated in the channel (Fig. 35). The resulting lateral baroclinic pressure gradient was opposite to that described in the gravitational adjustment under constant vertical mixing with similar Ekman number, $E = 0.02$ (Fig. 32). This is a lateral structure of flow and salinity fields that has not been described and that is expected to motivate future observations.

The transverse flow associated with this experiment featured two gyres (Fig. 35) that rotated in opposite direction with respect to those in Fig. 32, that is, lateral circulation showed divergence at the surface layer and convergence below that, downward flow near the lateral basin walls and upward flow in the channel. This pattern was likely a result of the lateral baroclinic pressure gradients established by wind forcing. Although the formation of the lateral gyres indicates an ageostrophic lateral momentum balance, axial asymmetries of the lateral flow evidence the contribution of the Coriolis terms. Over the right shoal, the lateral pressure gradient force and Coriolis force were additive, while over the left shoal these forces were opposite. Then, due to the earth's rotation effect, the right gyre intensified while the left gyre weakened.

**Strong winds**

For the experiment with relatively strong winds ($\tau = 0.2$ Pa) the $W_e$ increased to about 2.6 and the Ekman number took a value of 0.05. Flow and salinity fields featured similar lateral distribution to those of the previous case where $E \sim 0.035$ (Fig. 36). A noteworthy difference was that, in general, the velocity and salinity fields became more axially symmetric. These results showed that as the Ekman number increased by increasing the wind stress, the wind-driven circulation masked the buoyancy-induced circulation, and at the same time rotation influences. This
solution resembled the non-rotating solution featured in Fig. 26 in both along-estuary flow and salinity fields. The lateral flow, however, still showed axial asymmetries with a better developed right gyre (looking downwind) than left gyre.

*Down-estuary weak winds*

Changing the direction of wind forcing to down-estuary, caused buoyancy-induced and wind-driven flows to be additive, similarly to the non-rotating results. Application of a weak wind-stress ($\tau = 0.01 \text{ Pa}$) yielded $W_c$ lower than 1 and Ekman number of 0.0058. These values of the two non-dimensional numbers indicated that the buoyancy-induced circulation dominated the dynamics and that the Coriolis terms played an important role compared to vertical mixing (Fig. 37). Wind-stress effects
manifested only at the surface layer, in both longitudinal and lateral flows. The longitudinal velocity component at the surface layer was enhanced by the wind-stress (compared to the rotating up-estuary wind case). The lateral velocity component was toward the left (looking into the system), i.e., toward the right of the wind as expected from the earth’s rotation. Maximum values of the lateral velocity component were located at the surface.

In summary, inclusion of Earth’s rotation allowed flow and salinity field to show axial asymmetries. The solution approached the non-rotating cases as the Ekman number increased. Similarly to the non-rotating cases, wind-driven circulation dominated over buoyancy-driven flow when $W_e$ is near or bigger than 1.
FIG. 37. Same as Fig. 34 but for seaward wind stress of 0.01 Pa. ($\max |u| = 6.3 \times 10^{-2} \text{ m s}^{-1}$, $\max |w| = 3.0 \times 10^{-4} \text{ m s}^{-1}$).

IV.3.2.3 Particle trajectories

Similarly to the non-rotating cases, particle trajectories were evaluated for two different rotating cases of gravitational adjustment with constant eddy viscosity ($10^{-3} \text{ m}^2 \text{s}^{-1}$ and $5 \times 10^{-5} \text{ m}^2 \text{s}^{-1}$). The released initial positions were the same as those in the non-rotating cases, ($x_0 = 50 \text{ km}$, $y_0 = 2$ and $8 \text{ km}$; and $z_0 = 2$ and $5 \text{ m}$). Figures 38 and 39 illustrate the trajectory of the four particles (for each $A_z$ value), where the left panels show horizontal views and the right panels show transverse views.

For moderate vertical mixing ($A_z = 10^{-3} \text{ m}^2 \text{s}^{-1}$), all particles showed lateral convergence and divergence in their way toward the ocean (Fig. 38). In contrast with the non-rotating case, the trajectory of the particles was a function of the lateral
initial position and the particle trajectories were not axially symmetric. Particles released deeper were closer to the ocean after the same period of time owing to the weaker transverse circulation that affected them.

For weak vertical mixing \((A_z = 5 \times 10^{-5} \, \text{m}^2 \, \text{s}^{-1})\), the particle trajectories reflected the influence of the earth's rotation on the velocity field. Due to axial asymmetries of the flow, particle trajectories were highly dependent on the lateral and the vertical positions (Fig. 39). Particles released over the right shoal (looking into the system) moved toward the left shoal. When the particle released at 5 m depth reached the channel, it sank and moved with the inflow core. Particles released over the left shoal, left the system in a nearly straight path, owing to the strong longitudinal flow
developed at that location.

For strong vertical mixing \((A_z = 10^{-2} \text{ m}^2 \text{ s}^{-1})\), the particle trajectories in non-rotating and rotating systems showed similar paths (not shown here) as expected from the weak relative contribution of the Coriolis accelerations compared with frictional effects.

![Particle trajectories](image)

**FIG. 39.** Particle trajectories for the gravitational adjustment problem using constant eddy viscosity \((A_z = 5 \times 10^{-5} \text{ m}^2 \text{ s}^{-1})\) in a rotating system, released from different locations (indicated with the solid diamonds). Left panel \(x-y\) plane, right panel \(y-z\) plane.

In the non-rotating cases, the particles left the system faster in the weak vertical mixing case than in the case with strong vertical mixing (if the particles were at the surface). In the rotating system this was not the case. Particles released over the right shoal remained longer in the system in the weak vertical mixing case compared with strong vertical mixing.
In summary, the transverse flow, owing to the earth's rotation or lateral baroclinic pressure gradient, must be examined carefully in order to better understand the water exchange in semi-enclosed basins. Most of the particle trajectories became more complicated as the lateral baroclinic pressure gradient increased, passing from a lineal to a helical path. Also, comparing the non-rotating and rotating cases under weak vertical mixing conditions, it was clear that rotation had a tremendous influence on the paths followed by the particles and the time they remained in the system. Under weak mixing condition and rotation, the path of the particles showed a strong dependence on the initial release location, due to the axial asymmetries of the flow.
A three-dimensional primitive equation model has been used over idealized bathymetry with lateral depth variation to examine the lateral distribution of the wind-driven flow in four general scenarios: homogeneous fluid in a non-rotating frame, homogeneous fluid in a rotating-frame, stratified fluid in a non-rotating frame, and stratified fluid in a rotating-frame. In all homogeneous cases a turbulence closure model was used to examine the non-linear interaction between current and vertical mixing. In the stratified cases both a) fixed diffusivity and viscosity, and b) turbulence closure scheme model were employed. Fixed diffusivity/viscosity values were used in order to examine the sensitivity of the gravitational circulation to vertical mixing. The turbulence closure scheme was used to study the non-linear interaction of vertical mixing and currents induced by the wind and baroclinic pressure gradients.

In the homogeneous cases, both non-rotating and rotating, the wind-stress induced a circulation pattern consisting of downwind flow over the shoals and upwind flow in the channel along the entire domain. This pattern was consistent with other studies. For rotating cases, the details of the wind-induced pattern depended on the ratio of Ekman layer depth \((d)\) to the maximum depth \((h)\). In cases where \(h/d \geq 2\) the longitudinal flow was axially asymmetric in such a way that maximum downwind flow was located over the right shoal (looking downwind) in the northern hemisphere. The rotating solution approached to that described in a non-rotating case as \(h/d\) remained equal to or below one. Earth’s rotation effects were most evident in the transverse circulation. In non-rotating cases, the transverse circulation was very weak and became relevant only at the head of the system, where the water sank and reversed direction to move toward the mouth. In rotating systems, the relative importance of the transverse circulation to the longitudinal circulation increased as
$h/d$ increased. In general, the transverse circulation in rotating systems described three gyres. The main gyre was clockwise (looking downwind) and occupied the entire basin cross-section. Looking downwind, upwelling developed at the left boundary and downwelling appeared at the right boundary. The surface flow moved from left to right (to the right of the wind in the northern hemisphere) and bottom flow moved from right to left. The other two gyres were small and localized and were linked to the channel edges, at the transition between upwind and downwind flow. The new finding of these homogeneous cases were as follows. The inclusion of all terms in the momentum balance (as in the numerical model) reflect a transverse circulation with three gyres, as opposed to only one as in the linear solution of Winant (2003). Characterization of rotation effects in terms of the ratio $h/d$ will be depend, for a given value of $h/d$, on whether $h$ change or $d$ change.

In stratified systems, the gravitational circulation is unaffected by bathymetry when the longitudinal pressure gradient is only balanced by the stress divergence (friction) derived from turbulence closure. The lateral distribution of the longitudinal velocity component showed transverse structure only under relative high vertical mixing conditions (prescribed constant). These findings were consistent with previous studies. The new findings of the stratified cases are as follow. For both rotating and non-rotating systems under strong winds ($W_e \geq 1$), the wind-induced pattern of downwind flow over shoals and up-wind flow in the channels masked any buoyancy or rotation effects as the water column remained nearly vertically homogeneous and symmetric about the middle of the channel. For weak up-estuary wind ($W_e \ll 1$), the gravitational circulation (without or with rotation) was weakly altered. Only the upper part of the water column was modified by the wind-stress, showing an increased surface mixed layer and reduction of the flow. Down-estuary winds enhanced the density-induced exchange flows. The influence of the earth's rotation on flow and salinity patterns was characterized in terms of the Ekman number $E$. Low Ekman
number $(E \ll 1)$ cases showed strong asymmetries of the salinity and longitudinal flow and high Ekman number $(E \gtrsim 1)$ cases resembled those described in non-rotating systems (axially symmetric) as friction overwhelmed Coriolis effects. The transverse circulation and the salinity field showed a distribution expected from the balance of the lateral pressure gradient, friction and the earth’s rotation. For low Ekman numbers, the transverse dynamics approached geostrophy and the transverse circulation was linked to the along-channel flow through rotation. For high Ekman numbers, the transverse dynamics were dominated by density gradient and friction and the circulation showed two gyres located over the shoals. These results are expected to motivate future measurements that verify the numerical results in terms of axial asymmetries produced by rotation and the associated transverse circulation and in terms of the inverse estuarine circulation induced by up-estuary winds.


Reyes-Hernández, A. C., 2001: Tidal and subtidal lateral structures of density and velocity in the Chesapeake Bay entrance. Ph.D. Dissertation, Old Dominion University, Norfolk, VA, 131 pp.


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