The Dynamics of the Mississippi River Plume and Interactions with the Gulf of Mexico Offshore Circulation

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UNIVERSITY OF MIAMI

THE DYNAMICS OF THE MISSISSIPPI RIVER PLUME AND INTERACTIONS WITH THE GULF OF MEXICO OFFSHORE CIRCULATION

By
Rafael Vergara Schiller

A DISSERTATION

Submitted to the Faculty of the University of Miami in partial fulfillment of the requirements for the degree of Doctor of Philosophy

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THE DYNAMICS OF THE MISSISSIPPI RIVER PLUME AND INTERACTIONS
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River plumes often develop in complex environments, where variable coastal and bottom topography, ambient currents, winds and tides may play important roles in shaping the plume evolution. When all these factors are present, the plume dynamics may become intricate and unclear. The objective of this study is to understand the processes controlling the dynamics of a large river plume that is affected by strong boundary currents, variable winds and complex topography. The Mississippi River (MR) plume is the study case of this dissertation work, and focus is given to the interactions between the plume and the offshore circulation of the Gulf of Mexico (GoM).

A series of numerical experiments was designed to investigate the impact of different factors on the development of a large scale river plume in scenarios with variable degrees of complexity. First, a box-like model with an idealized estuary was designed to address the general development of a mid-latitude river plume and assess the variability of the plume with changes in the outflow conditions at the river mouth. The structure and development of the plume in the flat-bottom, receiving basin was highly dependent on the degree of freshwater mixing at the source. Larger freshwater mixing enhanced the estuarine gravitational circulation and modified the dynamical balance at the estuary
mouth. Those changes effectively modified the shape of the bulge and length/transport scales of the coastal current. Sloping-bottom conditions further modified the development of the plume.

Secondly, a Northern GoM model was designed and numerical experiments were conducted to investigate the specific dynamics of the MR plume, in the presence of both shelf and basin-wide circulation. In particular, buoyancy-driven (due to the MR and all other major Northern GoM rivers) and wind-driven currents were studied on the shelf, while the extension of the Loop Current and associated frontal eddies were considered as major factors in the shelf to offshore interactions; wind-driven, shelfbreak eddies were also considered. Process-oriented experiments demonstrate that westerly and southerly winds promoted the development of a surface Ekman layer that enhances the offshore advection of plume waters. The steep topography in the vicinity of the MR Delta was a favorable condition for that process. When the MR plume was subject to a full-blown scenario (realistically-forced experiment nested within a large-scale model), complex interactions between wind-driven and eddy-driven dynamics determined the fate of the plume waters. Offshore removal is a frequent plume pathway, and the offshore transport can be as large as the wind-driven shelf transport. The offshore pathways depend on the position of the eddies near the shelf edge, their life span and the formation of eddy pairs that generate coherent cross-shelf flows. Strong eddy-plume interactions were observed when the Loop Current (LC) system impinged against the shelfbreak, causing the formation of coherent, narrow low-salinity bands that extended toward the Gulf interior.
The offshore transport of MR water is a year-round process, but the interactions between the MR plume and the LC system have large inter-annual variability. Plume to LC interactions are determined by episodic northward intrusions of the LC system in the NGoM. The interactions are dictated by the proximity of the LC system to the MR Delta and by wind effects. On average, plume to LC interactions correspond to ~ 12 % of the year-round, total freshwater transport near the MR Delta, but this percentage can go up to 30 % in individual years. At the time of the plume to LC interactions, an average value of LC freshwater entrainment was estimated to be ~ 4,150 m³ s⁻¹. The findings presented here are a major contribution toward the understanding of the cross-marginal and basin-wide transport of MR waters by a large-scale current system, and the connectivity to remote regions, such as the South Florida region and the Florida Keys.
Dedication

This dissertation is dedicated to my family: to my parents, Vanderlei and Sonia, to my wife, Natasha, and to my son, Gabriel, for their unconditional love and support during my time in the Ph.D. program.
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# TABLE OF CONTENTS

<table>
<thead>
<tr>
<th>Chapter</th>
<th>Title</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>iv</td>
<td>List of Figures</td>
<td>viii</td>
</tr>
<tr>
<td>v</td>
<td>List of Tables</td>
<td>xxi</td>
</tr>
<tr>
<td>1</td>
<td>Introduction</td>
<td>1</td>
</tr>
<tr>
<td>1.1</td>
<td>Study Motivation</td>
<td>5</td>
</tr>
<tr>
<td>1.2</td>
<td>Background</td>
<td>7</td>
</tr>
<tr>
<td>1.3</td>
<td>Objectives and Strategy</td>
<td>10</td>
</tr>
<tr>
<td>2</td>
<td>The Impact of Discharge Conditions on Plume Development</td>
<td>12</td>
</tr>
<tr>
<td>2.1</td>
<td>Overview</td>
<td>12</td>
</tr>
<tr>
<td>2.2</td>
<td>Model description</td>
<td>14</td>
</tr>
<tr>
<td>2.2.1</td>
<td>Freshwater flux and river inflow parameterization</td>
<td>16</td>
</tr>
<tr>
<td>2.2.2</td>
<td>Vertical mixing schemes</td>
<td>18</td>
</tr>
<tr>
<td>2.2.3</td>
<td>Hybrid vertical coordinate grid generator</td>
<td>19</td>
</tr>
<tr>
<td>2.3</td>
<td>Box model domain set-up</td>
<td>20</td>
</tr>
<tr>
<td>2.4</td>
<td>River plume experiments</td>
<td>22</td>
</tr>
<tr>
<td>2.4.1</td>
<td>Control Experiment</td>
<td>22</td>
</tr>
<tr>
<td>2.4.2</td>
<td>Prescribed river inflow distributions inside the estuary</td>
<td>25</td>
</tr>
<tr>
<td>2.4.3</td>
<td>Enhanced vertical mixing inside the estuary</td>
<td>31</td>
</tr>
<tr>
<td>2.5</td>
<td>Discussion of results</td>
<td>33</td>
</tr>
<tr>
<td>2.5.1</td>
<td>Variability of outflow properties</td>
<td>33</td>
</tr>
<tr>
<td>2.5.2</td>
<td>Dynamical balance of the outflow</td>
<td>39</td>
</tr>
<tr>
<td>2.5.3</td>
<td>Topographic contrains on the development and transport of plume waters</td>
<td>43</td>
</tr>
</tbody>
</table>
2.5.4 Plume development in hybrid coordinate layers ........................................... 48
2.6 Summary and concluding remarks .................................................................. 51

Chapter 3 The Impact of Bottom Topography and Winds on the Dynamics of the
Mississippi River Plume ................................................................................. 56
3.1 Overview ........................................................................................................ 56
3.2 The NGoM-HYCOM model ......................................................................... 56
3.3 Process-oriented simulations ...................................................................... 59
3.4 Results ........................................................................................................... 60
   3.4.1 Impact of bottom topography on plume development (topography
experiments) ................................................................................................. 60
   3.4.2 Plume response to idealized wind forcing (wind experiments) ............ 63
3.5 Discussion and concluding remarks ............................................................. 70

Chapter 4 Interactions Between the Mississippi River Plume and Offshore Boundary
Currents ......................................................................................................... 73
4.1 Overview ........................................................................................................ 73
4.2 Realistically-forced simulation .................................................................... 73
4.3 Results ........................................................................................................... 76
   4.3.1 Shelf transport of plume waters around the MR Delta ....................... 79
   4.3.2 Offshore transport of MR waters ......................................................... 86
   4.3.3 Variability of MR plume pathways during 2004-2005 ....................... 96
4.4 Discussion .................................................................................................... 104
4.5 Concluding remarks .................................................................................... 109

Chapter 5 Loop Current Impact on the Transport of Mississippi River Waters .... 112
5.1 Overview........................................................................................................................................... 112
5.2 Long-term, realistically-forced simulation ...................................................................................... 112
5.3 Results................................................................................................................................................ 113
  5.3.1 Influence of the offshore NGoM circulation on the MR plume.............113
  5.3.2 LC dynamics and northward intrusions during 2004-2008............... 115
  5.3.3 Freshwater analysis and MR plume entrainment by the LC system ....... 120
  5.3.4 River discharge and wind conditions during the LC intrusion events....... 123
  5.3.5 Interactions between the LC and the Northern GoM shelf.................... 126
  5.3.6 Variability of offshore exportation of Northern GoM riverine waters.... 132
5.4 Discussion........................................................................................................................................... 135
5.5 Concluding remarks.......................................................................................................................... 142

Chapter 6  Summary and Conclusions.................................................................................................. 144

References............................................................................................................................................... 151
List of Figures

Figure 1.1 Gulf of Mexico region (bottom topography in meters). The location of the Mississippi River Delta is shown with the red circle. LaTeX: Louisiana-Texas shelf, MAFla: Mississippi – Alabama – Florida shelf .

Figure 2.1 Model configuration. Left: Idealized box-like basin. The thick short black line denotes where river discharge is imposed (estuary head). Land area is shaded. Center: Zoom of the estuarine region. Dashed lines show position of vertical sections where model results are evaluated: along the estuary and across the basin (1, extending to the east boundary), across the estuary mouth (2) and along-shore in the vicinity of the estuary (3). Right: Vertical section of bathymetry from the sloping bottom setup showing cartesian levels (thin lines, along section 1); additional hybrid configurations are exhibited in Figure 2.13.

Figure 2.2 Upper: Sea Surface Height contours in mm (left), Sea Surface Salinity contours (middle) and near surface velocity vectors in cm s⁻¹ (right) from the Control-flat experiment at day 60 (part of the model domain shown). Lower: Along-estuary/across-shore salinity vertical structure along section 1 (marked in Figure 2.1). The plume boundary (34.9) is represented by a white line. Salinity values less than 25 (inside the estuary) are not shown. Vertical black line denotes the position of the estuary mouth.
Figure 2.3 As Figure 2.2, but for the Control-slope experiment............................... 25

Figure 2.4 Sea Surface Salinity contours from experiments with variable distribution of river inflow inside the estuary, at day 60 (part of the model domain shown). The plume boundary (34.9) is represented by a white line. Salinity values less than 25 (inside the estuary) are not shown. The upstream (L_u), downstream (L_d) and offshore (L_o) plume intrusions for each case are displayed next to the plots. Downward penetration (pntr) and horizontal spreading (sprd) configurations that characterize each experiment are also presented.........................................................26

Figure 2.5 Near surface velocity vectors from experiments with variable distribution of river inflow inside the estuary, at day 60 (part of the model domain shown). Downward penetration (pntr) and horizontal spreading (sprd) configurations that characterize each experiment are presented. Vectors are plotted every other grid point for better visualization.................................................................27

Figure 2.6 Sea Surface Salinity contours (upper) and near surface velocity vectors (lower) from experiments with enhanced mixing (increased $K_{sw}$) inside the estuary, at day 60 (part of the model domain shown). The plume boundary (34.9) is represented by a white line. Salinity values less than 25 (inside the estuary) are not shown. The mixing information that characterizes each experiment is shown next to each plot (see section 2.4.3 for details). Vectors are plotted every other grid point for better visualization.................................................................32
Figure 2.7 Sea Surface Height (SSH, in mm), across-estuary vertical salinity structure (colors) and along-estuary velocity (u, cm s\(^{-1}\), solid for positive/offshore and dashed for negative/onshore contours) along section 2 (estuary mouth) from selected experiments, at day 60. The configurations that define each experiment and the outflow transport \(T_f\) for each case are shown.

Figure 2.8 Upper: Magnitude of each of the major momentum balance vectors 

\[ |\vec{c}\vec{o}\vec{r}| = \sqrt{(-f)(\hat{y})^2 + (f)(\hat{x})^2}, \quad |\vec{p}\vec{g}\vec{f}| = \sqrt{(-\rho_0^{-1}\partial p/\partial x)^2 + (-\rho_0^{-1}\partial p/\partial y)^2} \]

and

\[ |ac\vec{c}\vec{e}\vec{f}| = \sqrt{(D\vec{u}/Dt)^2 + (D\vec{v}/Dt)^2} \]

where \(D(u,v)/Dt = \partial(u,v)/\partial t + \vec{V} \cdot \nabla(u,v)\) from selected experiments. Values were extracted from section 2 (estuary mouth) at the location of the near surface outflow core, at day 60. Middle: Correspondent individual components of each geostrophic balance term normalized by the associated vector length (\(|\vec{c}\vec{o}\vec{r}|/|\vec{c}\vec{o}\vec{r}|, |\vec{c}\vec{o}\vec{r}|/|\vec{c}\vec{o}\vec{r}|, |\vec{p}\vec{g}\vec{f}|/|\vec{p}\vec{g}\vec{f}|\) and \(|\vec{p}\vec{g}\vec{f}|/|\vec{p}\vec{g}\vec{f}|\)).

Lower: Average positive (upward) and negative (downward) vertical velocity inside the estuary, at day 60.

Figure 2.9 Snapshots of near surface velocity vectors from Riv2c experiments (both from flat and sloping bottom conditions) starting on day 5 to day 30, every 5 days (part of the model domain shown). Vectors are plotted every other grid point for better visualization.
Figure 2.10 Hovmöller diagrams of surface relative vorticity $\zeta = \left( \frac{\partial v}{\partial x} \right) - \left( \frac{\partial u}{\partial y} \right) \left( 10^{-5} \text{ s}^{-1} \right)$, left panels) and of vertical velocity $\left( 10^{-5} \text{ m s}^{-1} \right)$ at 15m below the surface (model layer 12, right panels) from the Riv2c experiments (upper: flat bottom, lower: sloping bottom) along section 3 (vicinity of the estuary, Figure 2.1).......................... 45

Figure 2.11 Locations of the offshore edge of the recirculating bulge from selected pairs of flat and sloping bottom experiments (Control, Riv2c and Mix4c). Positions are shown every 5 days for better visualization.......................................................... 47

Figure 2.12 Time series of integrated downstream coastal current transport $\left( \text{m}^3 \text{s}^{-1} \right)$, left) and displacement of the coastal current nose (km, right) away from the estuary mouth from selected pairs of flat and sloping bottom experiments (Control, Riv2c and Mix4c). Coastal current transports were calculated at an across-shore section 127.5km south of the estuary.......................................................... 48

Figure 2.13 Right: Across-shore salinity vertical structure along section 1 (see Figure 2.1), starting at the estuary mouth (where the slope starts) from the Riv2c-slope experiment with three different vertical layers setting, at day 60. Upper: cartesian-isopycnal. Middle: sigma only. Lower: sigma – isopycnal. Layer interfaces are shown as solid white lines. Left: Corresponding Sea Surface Salinity field, for each case. The plume boundary (34.9) is represented by a white line................................. 50
Figure 3.1 NGoM-HYCOM domain. Selected isobaths are shown and the shelfbreak (100 m isobath) is highlighted in blue. The location of the Mississippi River is shown (Mis, red dot). ................................................................. 58

Figure 3.2 Snapshots of surface salinity and surface velocity vectors after 15 and 30 days of buoyancy-forcing only in the presence of realistic bottom topography (left) and in the presence of a 20 m deep flat bottom (right). The isobaths of 10 m, 20 m, 50 m and 100 m (shelfbreak) are shown as solid gray lines. Vectors are shown every other 8 grid points for better visualization ................................................................. 61

Figure 3.3 Snapshots of surface salinity and surface velocity vectors after 36 and 72 hours of westerly (left) and southerly (right) winds. The shelfbreak (100m isobath) is represented as a solid gray line. Vectors are shown every other 8 grid points for better visualization. The solid black lines show locations of vertical across-sections in Figure 3.4 ........................................................................................................ 64

Figure 3.4 Vertical across-shore sections of salinity after 72 hours of westerly and southerly winds. The locations of the sections are shown in Figure 3.3. Dashed white lines show the locations of vertical profiles A, B, C and D where major momentum balance terms are calculated ................................................................. 67

Figure 3.5 Vertical profiles of momentum balance terms computed at points A, B, C and D (shown in Figure 3.4) after 72 hours of westerly (W) and southerly (S) winds.
Upper panels show across-shore (y dir.) terms and lower panels show along-shore (x dir.) terms.

Figure 3.6 Snapshots of surface salinity and surface velocity vectors after 72 hours of easterly (left) and northerly (right) winds. The shelfbreak (100m isobath) is represented as a solid gray line. Vectors are shown every other 8 grid points for better visualization.

Figure 4.1 (a): NGoM-HYCOM model. Selected isobaths are shown and the shelfbreak (100 m isobath) is highlighted in blue. The location of major rivers where freshwater discharges are imposed are shown as red dots (Atc: Atchafalaya, Mis: Mississippi, Mob: Mobile Bay). (b): Sea surface height (ssh, in cm) from the GoM-HYCOM model on August 7th, 2004. The location of the NGoM-HYCOM model is shown. The thick black line represents the Loop Current (LC) ssh 17-cm contour that is employed to track the maximum latitude of the LC in time (Section 4.3.2).

Figure 4.2 Sea surface salinity from selected days depicting different conditions of the MR plume (part of the model domain shown). (a): during a period of easterly, downwelling-favorable winds, (b): after a period of westerly, upwelling-favorable winds, (c): during a period of southerly winds. The gray lines represent the 100 and 1000 m isobaths. The blue vector shows the average wind direction during each event at a point in front of the MR Delta (white triangle). The black line shows the location of the vertical sections of salinity presented in (d) (day 87, easterly winds).
and (e) (day 100, westerly winds). Salinity values less than 26 are not shown. The vertical layers in the upper water column at the location of the section are illustrated in (e) as grey solid lines.

Figure 4.3 Location of the sections (m1, m2, z1, s1 and iso1000m) where freshwater transport analysis is performed. The area \( A \) represents the region where wind stresses are spatially averaged. Selected isobaths are shown (solid gray lines). The dots on section iso1000m show the initial (green dot) and final (red dot) points of the along-section distance in Figs. 4.7 and 4.8.

Figure 4.4 Vertical structure of the kinetic energy ratio \( Reke \) at sections m1, m2, z1 and s1 (marked in Fig. 4.3). The magenta line represents the 0.3 contour. Values above 1 are not shown.

Figure 4.5 (a): Time series of wind stress components computed as spatial averages from area \( A \). Selected wind periods for further analysis are delimited by vertical dashed gray lines and are identified as W1-6. (b), (c), (d) and (e): Time series of barotropic and baroclinic freshwater transport \( Q_{fw} \) across sections m1, m2, z1 and s1 (marked in Fig. 4.3). Positive/negative values represent eastward/westward transport across sections m1 and m2, northward/southward transport across section z1 and onshore/offshore transport across section s1.
Figure 4.6 Time series of the maximum latitude of the Loop Current (LC) calculated from the full GoM-HYCOM regional model. The maximum latitude is based on the northernmost point of the LC ssh17-cm contour exemplified in Fig. 4.1b. Three periods of LC impact on the MR dynamics that are further investigated are shown (LC1,2 and 3) and are delimited by vertical red lines. The horizontal gray line represents the latitude of the southern open boundary of the nested NGoM-HYCOM model.

Figure 4.7 Hovmöller diagram of ssh anomaly (a), across-section surface velocity (b) and total freshwater transport $Q_{fw}$ (c) across section iso1000m for year 2004. Positive/negative values for onshore/offshore velocities and transports. The solid black line in (a) represents the ± 0.1 contour. Wind periods of interest are shown as vertical magenta lines in (c). Periods of LC impact are delimited by horizontal dashed black lines. Selected anticyclonic (A*) and cyclonic (C*) eddy events are shown. The orientation of the diagrams is given by the green and red dots and is shown in Fig. 4.3.

Figure 4.8 Same as Fig. 4.7, but for year 2005.

Figure 4.9 Snapshots of sea surface salinity and surface velocity vectors from selected days during the wind periods W2 (a, left column) and W5 (b, right column). Part of the model domain shown. Vectors are shown every other 8 grid points for better visualization. The 100 and 1000 m isobaths are displayed as gray lines. A and C show the location of anticyclonic and cyclonic eddies, respectively. For each day,
the wind stress vector averaged over the area $A$ is presented on the upper-right corner of the plot. Salinity values less than 29 are not shown................................. 94

Figure 4.10 Same as Fig. 4.9, but for LC periods LC1 (a, left column) and LC2 (b, right column).................................................................................................................. 95

Figure 4.11 (a): Time series of total freshwater transport $Q_{fw}$ across sections m2 and iso1000m (marked in Fig. 4.3). Positive/negative values for eastward/westward transport across section m2. Only offshore (negative) portion of the transport is shown for the section iso1000m. Selected wind periods are delimited by vertical dashed gray lines. Periods of LC and eddy impacts are represented by horizontal blue lines. (b): Time series of Mississippi River discharge for the period of the simulation..................................................................................................................................... 97

Figure 4.12 Snapshots of sea surface salinity and surface velocity vectors during two eddy events, C3-A3 (a) and A2-C1 (b). Part of the model domain shown. Vectors are shown every other 8 grid points for better visualization. The gray lines represent the 100 and 1000 m isobaths. A* and C* show the location of anticyclonic and cyclonic eddies, respectively. For each day, the wind stress vector averaged over the area $A$ is presented on the upper-right corner of the plot. Salinity values less than 29 are not shown ...................................................... 98
Figure 4.13 (a): Oceansat-1 OCM chlorophyll $a$ images from selected days during events W2, W5, LC1 and LC2. The chlorophyll $a$ scale ranges from 0 to 3 mg m$^{-3}$. The 100 and 1000 m isobaths are shown as black lines. River water is depicted in red/brown tones near the mouth of rivers, where pigment concentrations are highest and orange/yellow tones in deep water, where pigment concentrations are reduced. (b): Snapshots of model sea surface salinity on the same days as the satellite images. The 100 and 1000 m isobaths are displayed as gray lines. Salinity values less than 29 are not shown.

Figure 5.1 Sea surface salinity and surface velocity vectors from selected days that exemplify different interactions between the MR plume and the NGoM offshore circulation (part of the model domain shown). Vectors are shown every 8 grid points for better visualization. The gray lines represent the 100 and 1000 m isobaths. A and C show the location of anticyclonic and cyclonic eddies, respectively. Salinity values less than 29 are not shown. (a): Eddy-induced entrainment after a period of eastward, wind-driven plume transport. (b): Offshore removal by a Loop Current Eddy (LCE). (c): Offshore removal induced by a cyclonic eddy field. (d): Offshore removal induced by an anticyclone in the presence of southwesterly winds.

Figure 5.2 Location of offshore sections o1 and o2 that are employed in the analysis of results. Selected isobaths are shown. The red triangle marks the location of wind stress time series in Figures 5.4-5.7.
Figure 5.3 (a): Time series of maximum Loop Current (LC) latitude calculated from the full GoM-HYCOM regional model. The maximum latitude is based on the northernmost point of the LC ssh 17 cm contour exemplified in Figure 4.1b. The horizontal gray line represents the latitude of the southern open boundary of the nested NGoM-HYCOM model. LC intrusion events (LC1 to LC6) are represented by shaded areas. (b): Time series of maximum sea surface height (ssh in m) calculated at the offshore section 02. (c) Same as (b), but for maximum temperature at the depth of 100 m.

Figure 5.4 (a): Time series of daily river discharge from the Mississippi River (MR), Atchafalaya River and all other rivers combined for years 2004 – 2008. The monthly climatology of the MR discharge is also shown for comparison. Periods of Loop Current (LC) intrusion events (LC1 to LC6) are shown as gray shaded areas. S-S-F-W stands for Spring-Summer-Fall-Winter seasons. (b) to (g): Time series of wind stress components at a point in front of the MR Delta (red triangle, Figure 5.2) during each of the LC intrusion events. In each plot, the gray shaded areas represent periods of direct interaction between the MR plume and the LC system.

Figure 5.5 Example of Case I of Loop Current (LC) intrusion event. (a): Time series of wind stress vectors at a point in front of the MR Delta (red triangle, Figure 5.2). (b): Time series of $Q_{nw}$ across 01 (Figure 5.2). The red and black dashed lines show the
days of the snapshots on (c) and (d). Shaded areas represent periods of time when the LC system interacted with the MR plume. (c) and (d): Snapshots of sea surface salinity and surface velocity vectors on selected days (part of the model domain shown). Vectors are plotted every other 8 grid points for better visualization. The gray lines represent the 100 and 1000 m isobaths. Salinity values less than 29 are not shown. LC and LCE show the locations of the Loop Current and Loop Current Eddy, respectively. (e) and (f): Snapshots of sea surface height (ssh in cm) from the GoM-HYCOM model on the same days as (c) and (d). The LC boundary is shown as a solid black line.

Figure 5.6 Same as Figure 5.5, but for an example of Case II of LC intrusion. Wind is smoothed with a 15 days, low-pass filter for better visualization.

Figure 5.7 Same as Figure 5.5, but for an example of Case III of LC intrusion.

Figure 5.8 (a): Time series of freshwater equivalent depth integrated along section 01 (f_{wed-int}). (b): Time series of Q_{fw} across section 01. Daily values are shown in blue, and 30 days, low-pass filtered values are shown in red. Periods of Loop Current (LC) intrusion events (LC1 to LC6) are shown as gray shaded areas. S-S-F-W stands for Spring-Summer-Fall-Winter seasons.

Figure 5.9 (a): Snapshots of sea surface salinity and surface velocity vectors on June 11th, 2007, during event LC6 (part of the model domain shown). Vectors are plotted
every other 8 grid points for better visualization. The gray lines represent the 100 and 1000 m isobaths. Salinity values less than 29 are not shown. (b): Snapshot of sea surface height (ssh in cm) from the GoM-HYCOM model on the same days as (a). The LC boundary is shown as a solid black line. LC, LCE and C show the locations of the Loop Current, Loop Current Eddy and a cyclonic eddy, respectively. (c): MODIS Aqua chlorophyll $a$ image on June 9$^{th}$, 2007
List of Tables

Table 2.1 Summary of the attributes from the study experiments. Downward penetration is given in percentage of the water column and in meters (parentheses). Lateral spreading is classified as none, short (4 grid points, half estuary length) or large (7 grid points, full estuary length). KPP background vertical mixing is characterized as standard (no modifications) or as Enhanced $K_{iw}$ (salinity diffusivity due to background internal wave mixing equals $10^{-4}$ m$^2$ s$^{-1}$) plus region where it is applied. See sections 2.4.2 and 2.4.3 for details. 22

Table 2.2 Outflow transport $T_f$, outflow upper layer 1 and inflow lower layer 2 vertical mean values of density $\rho$, along-estuary ($u$) and across-estuary ($v$) velocities and layer thickness $h$, for all flat bottom experiments at day 60. Layers 1 and 2 average values were calculated from a vertical profile located at the core of the surface outflow at the estuary mouth. See Table 2.1 for attributes from experiments and section 2.5.1 for details on the calculations. 36

Table 2.3 Non-dimensional numbers calculated for all flat bottom experiments at day 60. See Table 2.1 for attributes from experiments. $Ri$: Gradient Richardson number; $Fr$: Froude number; $Ro_i$: Inlet Rossby number; $K_i$: Inlet Kelvin number; The internal deformation radius $R_{di}$ (km), the squared stratification frequency $N^2$ (s$^{-2} * 10^2$) and the squared vertical velocity shear $S^2$ (s$^{-2} * 10^{-2}$) are also shown. Numbers were calculated using values presented in Table 2.2. See section 2.5.1 for details. 38
Table 3.1 Plume offshore area (beyond the shelfbreak) after 3 days of steady wind forcing from 8 different directions. Wind stress is constant at 0.075 Pa. Percentage represents an increase or decrease in the offshore area with respect to the initial offshore area at the beginning of the wind experiments.

Table 4.1 Total freshwater volume exported to the offshore region during each wind (W*), Loop Current (LC*) and eddy (A*,C*) event. Events that were superimposed in time are put together as one single exportation period (W5 and LC2, for example). The freshwater volume discharged by the Mississippi River during each period is also shown, together with the ratio between the volume exported and the volume discharged.

Table 5.1 Summary of Loop Current (LC) intrusion events. For each event, the start date, duration and interval until the next event are shown. A description and duration for each phase of the events is presented.

Table 5.2 Statistics of the offshore freshwater transport across section o1 for each LC intrusion event. Days of direct interaction between the MR plume and the LC system represent days when low-salinity bands are observed to extend from the MR plume beyond the shelfbreak, along the edge of the LC system.

Table 5.3 Estimates of the total volume of freshwater exported to the offshore region across o1 for years 2004 – 2008 ($V_{fw-exported-year}$). The volume of plume freshwater
directly entrained by the LC system during the same years is also shown ($V_{fw\text{-plume-LC-year}}$). Volume unit is km$^3$ ................................................................. 134

Table 5.4 Freshwater volume discharged by the MR during each LC intrusion event ($V_{MR}$, in km$^3$). $V_{fw\text{-plume-LC}}$ is shown for comparison, as well as the ratio $V_{fw\text{-plume-LC}} / V_{MR}$................................................................. 138
Chapter 1

Introduction

River plumes are common features in the coastal regions of most continental shelves in the world. The discharge of freshwater from rivers and estuaries into the coastal ocean produces one of the most common types of density-driven oceanic flows. Since the freshwater discharge tends to be less saline and hence lighter than the ambient water, a plume typically forms as the buoyant water spreads away from the mouth of the river or estuary. The spreading of the buoyant plume is governed by the characteristic circulation pattern induced by the plume density gradients and volume inflow. The spreading may be influenced by a variety of physical processes on the shelf and offshore regions, at different temporal and length scales.

It is acknowledged that in the absence of external forcing (such as winds, tides and ambient currents) and if a river buoyant plume is large enough to be affected by the Coriolis force, riverine waters will turn anticyclonically when they reach the shelf, and move away from their land source as an along-shore, buoyancy-driven coastal current, in the direction of Kelvin wave propagation. The driving mechanisms of this circulation are the pressure gradients associated with the density and volume anomalies, and the subsequent geostrophic adjustment of the plume circulation on the shelf. Observations of large river plumes on open shelves such as along the U.S. east and west coasts include:
the Delaware (Münchow and Garvine, 1993a; b) and the Chesapeake (Boicourt, 1973; Mamorino and Trump, 2000) Bays, the Columbia (Hickey et al., 1998; Hickey et al., 2005; Horner-Devine, 2009), the Hudson (Chant et al., 2008) and the Niagara (Masse and Murthy, 1992; Horner-Devine et al., 2008) Rivers, and the low-salinity coastal band in the South Atlantic Bight (Blanton et al., 1994). Satellite and field studies have also shown evidence of a bulge-like region in the vicinity of the river inflow, where plume waters recirculate before feeding the coastal current (Masse and Murphy, 1992; Hickey et al., 1998; Chant et al., 2008; Horner-Devine et al., 2008; Horner-Devine, 2009). Similar behavior has also been observed in laboratory studies (Stern et al., 1982; Griffiths and Hopfinger, 1983; Whitehead and Chapman, 1986; Avicola and Huq, 2003a,b; Horner-Devine et al., 2006).

In addition to observations, numerical and analytical models have been used to understand and clarify the dynamics of coastal buoyant plumes. Many studies have been conducted in idealized scenarios. Rectangular basins were employed with simplified bottom topography, with buoyancy-forcing only or with additional simple external forcing, such as constant and unidirectional winds, along-shore ambient current and single component tides. Such idealized studies revealed many features of river plume dynamics generally hard to extract from observations, where complex circulation forcing mechanisms impact the plume behavior. Early studies recognized the importance of rotation, friction and non-linearity in the development of the frontal structure of large buoyant discharges (Kao et al., 1977; Beardsley and Hart, 1978; Kao, 1981; Minato, 1983; Ikeda, 1984; Garvine, 1987; O’ Donnell, 1990). Pioneering numerical modeling
studies demonstrated the impact of vertical mixing, bottom drag and sloping bottom on
the spin-up, maintenance and dissipation of river-forced plumes (Chao and Boicourt,
1986; Chao, 1988a,b) as well as the coastal current variability associated with barotropic
and baroclinic instabilities (Oey and Mellor, 1993). The variability of the bulge and the
coastal current from a river-forced plume was also demonstrated by Kourafalou et al.
(1996a), who elucidated the effects of buoyancy-induced stratification versus available
mixing in determining the expansion of the bulge and the coastal current meandering.
The impact of bottom slope on the plume offshore expansion was also investigated by
Kourafalou et al. (1996a) and Garvine (1999). The importance of river mouth conditions
for the bulge development and coastal current transport were also recognized by
Yankovsky and Chapman (1997), Garvine (1999), Yankovsky (2000) and Garvine

The influence of ambient forcing on the dynamics of river plumes has also been the
focus of different studies. Chao (1988b), Kourafalou et al. (1996a), Fong and Geyer
investigated the impact of wind stress on the plume development, offshore expansion and
mixing. The role of tidal mixing over the plume vertical structure was addressed by
Kourafalou et al. (1996b), Garvine (1999) and Soares et al. (2007a). The variability of the
bulge structure and coastal current transport due to the presence of along-shelf, ambient
currents was addressed by Fong and Geyer (2002) and Garcia Berdeal et al. (2002).
Numerical simulations by Soares et al. (2007b) expanded this topic, as they demonstrated
the impact of a large-scale current system (Malvinas-Brazil) on the offshore removal of La Plata River waters.

Interactions between large river plumes and large-scale, offshore currents have been observed for other rivers, such as the plumes from the Amazon, the Orinoco and the Mississippi Rivers (Walker et al., 1994; Ortner et al., 1995; Corredor et al., 2004; Ffiled, 2005; Hu et al., 2005). However, targeted studies to understand how the dynamics of large river plumes are influenced by offshore, energetic boundary currents, in addition to the effects of winds and topography, are uncommon. The development of a river plume in a complex environment where multiple shelf and offshore processes act simultaneously to change the plume dynamics is a challenging aspect of the study of buoyant plumes. The goal of this dissertation is to expand river plume studies and improve the understanding of how a large river plume interacts with offshore boundary currents, and how the process can be influenced by wind and topographic effects. An ocean general circulation model is employed for that purpose, and the Mississippi River plume is chosen as the study site. As discussed below, the Mississippi River plume is a unique study case of a large river plume that develops in a complex environment, where variable wind forcing, complex bottom and coastal topography, along with interactions with a strong boundary current and a rich eddy field, are integral parts of the plume dynamics.

This dissertation is organized as following: the study motivation, background on the study area, objectives and strategy are presented later on this chapter. A first time application of the numerical model to the study of buoyant plumes is presented in chapter
2. Process-oriented simulations of the Mississippi River plume that explore the effect of buoyancy-forcing only and buoyancy-forcing simultaneously to winds are presented in chapter 3. The influence of offshore circulation features and boundary currents are explored in realistically-forced simulations that are presented in chapter 4. A multi-year simulation that addresses the inter-annual variability of winds and offshore currents on the transport of Mississippi River waters is presented in chapter 5.

1.1 Study Motivation

The Mississippi River (Figure 1.1) is the major source of freshwater, sediments and nutrients for the Gulf of Mexico (GoM). Draining 41% of the continental United States, it is the largest river in North America, ranks as the 8th largest worldwide in terms of discharge (mean $1.35 \pm 0.2 \times 10^4$ m$^3$ s$^{-1}$, Hu et al., 2005) and transports about 210 million tons of sediment to the GoM annually (Milliman and Meade, 1983). The Mississippi River delivers highly productive waters that, together with other local rivers, fuel important fishery activities on the Louisiana-Texas shelf that yield approximately 28% of the total US catch (Rabalais et al., 1991). At the same time, the nutrient-rich, riverine waters are responsible for the recurrent summertime hypoxia (oxygen concentration < 2 mg l$^{-1}$) in bottom waters of the mid and inner-shelf west of the MR Delta (Rabalais et al., 1991, 2002, 2007).
Figure 1.1 Gulf of Mexico region (bottom topography in meters). The location of the Mississippi River Delta is shown with the red circle. Latex: Louisiana-Texas shelf, MAFla: Mississippi – Alabama – Florida shelf.

The Mississippi River also acts as a remote source of nutrients and pollutants for remote coastal systems, and connectivity to the South Florida ecosystem has been documented in the literature. One of the most representative events of direct linking occurred during the Mississippi River great flood in the summer of 1993 (Walker et al., 1994). In that time, the Mississippi River basin in the Midwestern United States experienced anomalously high rainfall, which produced abnormally high river discharges ($\sim 2 \times 10^4 \text{ m}^3 \text{ s}^{-1}$) from the Mississippi and Atchafalaya Rivers during July and August, traditionally months of low river discharge. Drifting buoys that were deployed to the west of the Mississippi River Delta in late July showed tracks that extended to the Straits of Florida (Walker et al., 1994). Moreover, during the period of September 10-13 1993, a band of anomalously low-salinity water embedded in the Florida Current and adjacent
coastal waters was observed extending from Key West to Miami (Ortner et al., 1995; Gilbert et al., 1996). Drifters, satellite data, and in situ observations within this band suggested that the Mississippi River was the source of this event, and that the Loop Current system was responsible for entraining riverine waters across the GoM. A more recent study by Hu et al. (2005) also reported the presence of a low-salinity band extending from the eastern Gulf of Mexico into the Straits of Florida and reaching the Gulf Stream off Georgia. This band was detected by MODIS imagery and sampled from ships in the Straits, and it was estimated to persist for a period of three months, from July to September of 2004. Again, the Loop Current – Florida Current system was the pathway for transporting plume waters across the basins.

With the across-basin transport of Mississippi River waters that carry nutrients and pollutants, comes the possibility of significant changes within the ecological balance maintained in the GoM and Florida waters. Understanding the processes controlling the transport and dispersion of the Mississippi River waters is important for ecosystem management and water quality purposes not only locally, in the Northern GoM region, but also regionally, in the GoM and South Florida region.

1.2 Background

The Mississippi River (MR) Delta divides the Northern GoM (referred as NGoM hereafter) continental shelf into two regions, the Louisiana-Texas Shelf towards the west and the Mississippi–Alabama–Florida shelf region to the east (Figure 1.1). On the west,
the continental shelf is wide, with the shelfbreak (roughly the 100 m isobath) located more than 100 km offshore. In contrast, the region east of the MR Delta is characterized by a narrow continental shelf and the presence of the DeSoto Canyon, where depths can go from 20 m close to shore down to 1000 m in about 100 km offshore. The Canyon deepest areas go beyond 2500 m of depth. The shelf off the MR Delta is narrow and steep, with the shelfbreak located just a few kilometers offshore. Shallow coastal areas (10-20 m deep) surround the northern part of the Delta.

River discharge into the NGoM shelf is distinctly seasonal, with highest flow occurring between March and May and lowest flow occurring between August and October. Approximately 70 % of the MR flow enters the NGoM through the bird-foot Delta, where riverine waters exit through several passes, the largest of which are the Southwest Pass, South Pass and Pass a Loutre. The remaining 30 % of the flow is discharged into the GoM by the Atchafalaya River further west (Walker et al., 1994). At the mouth of the passes (near-field region), the MR outflow is characterized by a bottom-attached salt front, and is subject to strong vertical mixing due to supercritical conditions (Wright and Coleman, 1971). Beyond this near-field region, mixing and spreading of the MR plume on the shelf are controlled by a variety of factors. The circulation in the NGoM inner continental shelf is primarily wind-driven (Cochrane and Kelly, 1986; Li et al., 1997; Nowlin et al., 2005), and the local wind is a major controlling factor over the surface circulation, structure and transport pathways of the MR plume (Walker, 1996; Walker et al., 2005a). Easterly winds (southeasterly to northeasterly, prevalent during autumn, winter and spring) drive westward surface currents along the south side of the
Delta, extend the MR plume towards the Louisiana-Texas shelf and promote accumulation of low-salinity waters, sediments and nutrients on the shelf between the Mississippi and Atchafalaya Deltas. This pattern is frequently disrupted by short-term wind reversals that occur due to the passage of colds fronts, which can reverse the direction of the coastal current, plume transport, and promote eastward and offshore dispersal of plume waters (Walker et al., 2005a). In the summer, the seasonal shift to more southerly winds also reverses the plume circulation and promotes an eastward transport of riverine waters towards the DeSoto Canyon (Walker et al., 1994; Morey et al., 2003a).

The offshore circulation in the GoM is dominated by the energetic Loop Current (LC) and mesoscale eddies. The LC is characterized by a cycle of northward penetration and westward bending until it reaches an unstable configuration, which leads to the shedding of an anticyclonic eddy (Loop Current Eddies, LCEs) (Maul, 1977; Vukovich et al., 1979; Hulburt and Thompson, 1980; Huh et al., 1981). This shedding process happens at intervals from 3 to 17 months (Sturges and Leben, 2000) and after LCEs are shed, they drift into the western GoM where they decay against the continental margin. Cyclonic frontal eddies (Loop Current Frontal Eddies, LCFEs) propagating clockwise along the LC front are also part of the system (Paluskiewicz et al., 1983; Vukovich, 1986; Fratantoni et al., 1998; Walker et al., 2003; Cherubin et al., 2006). The interaction of the LC system with shelf waters in the NGoM has been documented previously (Huh et al., 1981; Schroeder et al., 1987; Walker et al., 1994; Muller-Karger, 2000; Nowlin et al., 2000; Hamilton and Lee, 2005), and pronounced eddy energy and cross-shelf flows have been
observed in the vicinity of the MR outflow (Ohlmann et al., 2001). In this region, filaments of low-salinity waters can be entrained by the LC, LCEs or LCFEs when they impinge against the shelfbreak (Morey et al., 2003b; Walker et al., 2005a; Hamilton and Lee, 2005). This facilitates the likelihood for MR waters to be transported along the edge of the LC down to the Straits of Florida (Walker et al., 1994; Ortner et al., 1995; Gilbert et al., 1996; Hu et al., 2005). Direct offshore removal of MR waters induced by storms and strong wind events, followed by subsequent entrainment by slope eddies, has also been reported (Walker et al., 1996; Yuan et al., 2004; Stone et al., 2005; Walker et al., 2005b).

1.3 Objectives and Strategy

As mentioned before, the overarching goal of this dissertation is to expand river plume studies and improve the understanding of how a large river plume interacts with offshore boundary currents, and how the interactions are influenced by other environmental mechanisms, such as winds and bottom topography. More specifically, it is sought to understand the shelf and offshore processes that affect the development and transport of the MR plume, and the synergy of these processes on the offshore exportation of riverine waters. Specific objectives include:

1) Investigate the buoyancy-driven circulation induced by the MR plume, and how this circulation is affected by wind forcing.
2) Investigate how the MR plume interacts with the NGoM offshore circulation, and the impact of wind on that process.

3) Determine the most effective conditions for the offshore exportation of MR waters.

4) Quantify the fluxes of MR waters into the GoM interior and make a comparison with fluxes along the shelf.

5) Investigate the interactions between the Loop Current system and the MR plume, with emphasis on the transport of plume waters.

The strategy adopted in this dissertation is to employ an ocean general circulation model in a series of numerical simulations to address the objectives above. Firstly, the model is employed in a classic study of river plume dynamics in an idealized box domain. The aim is to validate the model for the study of river plumes and apply to the study of the impact of discharge conditions on the development of a buoyant plume. Secondly, a NGoM model is developed and employed in process-oriented simulations of the MR plume. These simulations address the plume buoyancy-driven circulation and the impact of wind. Finally, the NGoM model is applied in realistically-forced, long-term simulations to address the impact of offshore boundary currents on the plume dynamics, the combined role of winds in promoting the offshore removal of the plume and the interactions between the Loop Current system and the MR plume.
Chapter 2

The Impact of Discharge Conditions on Plume Development

2.1 Overview

The importance of the river mouth conditions to the variability of the bulge and coastal current transport has been reported by several studies. Yankovsky and Chapman (1997) developed a theory which relates properties of the estuarine discharge and cross-shore bottom slope to the bulge and coastal current structure. Garvine (1999) verified that the estuarine volume transport, scaled by the associated outflow geostrophic transport, controlled the greatest variance of the downshelf and across-shelf plume penetration. Fong and Geyer (2002) demonstrated that river mouth conditions affect the amount of freshwater transported by the coastal current relative to the bulge, which can accumulate low salinity waters, become unsteady and grow in time. They observed that when river outflows with larger Rossby number were simulated, more plume water recirculated within the bulge and that decreased the coastal current freshwater transport. In a series of laboratory experiments, Avicola and Huq (2003a,b) demonstrated how the “outflow angle” (angle between the outflow and the coastal wall) and the “impact angle” (angle at which the buoyant flow reattaches to the coast) affect the formation of the recirculating bulge. They suggested that the two angles are related (the outflow angle determines the
impact angle) and concluded that a coastal current formed at oblique impact angles, and the bulge recirculation increased as the impact angle approached 90 degrees.

Yankovsky (2000) and Garvine (2001) demonstrated that the implementation of the river boundary conditions may affect the near-field bulge circulation, more specifically the development of the plume upstream intrusion. Kourafalou et al. (1996a) showed that the upstream intrusion was due to a non-geostrophic balance between the along-shore acceleration and the along shore pressure gradient (due to low salinity waters near the river mouth and denser ambient waters up the coast). Yankovsky (2000) suggested that the upstream intrusion is enhanced by over simplified river boundary conditions that lack a baroclinic adjustment of the discharge (i.e., fixed uniform river inflow along the coastal wall). The blocking of the lower layer landward flow at the mouth promotes a strong cyclonic vorticity disturbance with corresponding upstream turning of the buoyant flow at the source, which enhances the upstream spreading of the plume. Yankovsky (2000) and Garvine (2001) concluded that the use of an inlet flow field that better mimics that observed at the mouth of estuaries (upper seaward buoyant flow on top of a lower landward undercurrent) reduces that impact.

In this chapter, the HYbrid Coordinate Ocean Model (HYCOM; Bleck, 2002; Halliwell, 2004; Chassignet et al., 2006) is used for the first time to investigate the dynamics of river plumes and the offshore propagation of buoyant waters. HYCOM is employed in an idealized estuary-coastal basin system to examine the development and evolution of a river plume under buoyancy forcing only and to investigate the plume
variability associated with changes in the conditions at the river mouth. These changes are shown to be the results of lateral and vertical spreading of the river inflow and variable mixing inside the estuary. These effects are expected to impact the estuarine circulation and the buoyant outflow, ultimately promoting changes in the recirculating bulge and the coastal current properties. Previous numerical modeling studies (Chao and Boicourt, 1986; Chao, 1988a; MacCready et al., 2009) have demonstrated the importance of the estuarine circulation to the development of the river plume. The focus here is to understand how the properties of the buoyant flow at the estuary mouth (which reflects the coupling between the estuarine and basin circulations) impact the development of the river plume in the receiving basin.

2.2 Model description

HYCOM is a primitive equation ocean general circulation model supported by code development and operational global/regional simulations associated with the HYCOM Consortium for Data Assimilative Modeling (see technical details in the model manual at www.hycom.org). HYCOM has been used in several large scale and marginal seas studies (Chassignet et al., 2003; Halliwell et al., 2003; Shaji et al., 2005; Kara et al., 2005a; Hogan and Hurlburt, 2006; Zamudio and Hogan, 2008), and it has been recently applied to the coastal ocean as well (Kourafalou et al., 2006; Olascoaga et al., 2006; Kourafalou et al., 2009; Halliwell et al., 2009). A comprehensive discussion of HYCOM's governing equations and numerical algorithms (including the hybrid coordinate grid generator) and the available vertical mixing schemes can be found in
HYCOM is a finite difference hydrostatic, Boussinesq primitive-equation model that solves 5 prognostic equations: one for each horizontal velocity component, a layer thickness tendency (mass continuity) equation, and two conservation equations for a pair of thermodynamic variables (salt, temperature or density). Here, salt and density are employed. Variables are stored on the Arakawa C grid. Thermodynamic variables and the horizontal velocity field are treated as “layer” variables that are vertically constant within layers but change discontinuously across layer interfaces. Other variables, such as pressure, are treated as “level” variables, defined on interfaces. These prognostic equations are complemented by a hydrostatic equation, an equation of state and an equation for the vertical mass flux through the layer interfaces, which controls their vertical movement and together with the hybrid coordinate grid generator defines the layer state (cartesian, sigma or isopycnal).

The prognostic equations are time-integrated using the split-explicit treatment of barotropic and baroclinic modes presented in Bleck and Smith (1990) with modifications by Morel et al. (2008). The baroclinic part of the solution is advanced in time with a leapfrog scheme, while the traditional forward-backward scheme for the barotropic solution presented in Bleck and Smith (1990) has been replaced by a new LSBM (Leapfrog for the Slow part of the Barotropic Mode) which leads to better stability properties (Morel et al., 2008). Horizontal mass fluxes are handled using the Flux Corrected Transport (FCT) scheme (Zalesak, 1979), while horizontal tracer advection is
computed using a centered difference scheme for tracer modified by an FCT procedure. Horizontal tracer diffusion and horizontal momentum viscosity follow Bleck et al. (1992). Wind-induced stress is assumed to be zero at the bottom of the mixed layer, and a quadratic form of bottom drag is employed to determine dissipation by the bottom. Different choices of vertical mixing schemes are available, which can calculate either vertical mixing throughout the water column or calculate mixed layer entrainment/detrainment separately from the weak, interior diapycnal mixing (Halliwell, 2004). Surface heat fluxes are parameterized according to Kara et al. (2000).

2.2.1 Freshwater flux and river inflow parameterization

In HYCOM, freshwater flux is parameterized as a virtual salt flux (Huang, 1993). While calculating vertical mixing at every baroclinic time step $dt_{bclin}$, the salinity $S$ in model layer 1 is updated to take into account changes due to freshening from river inflow or rain: $S = S + dS$. The salinity increment $dS$ is proportional to the virtual salt flux $S_f$:

$$dS = S_f \cdot dt_{bclin} \cdot g \cdot \frac{1}{dp} \quad \text{and} \quad S_f = \left[ - (P - E) - R \right] \cdot \frac{S \cdot \alpha_0}{1}; \quad g \text{ is gravity (9.806 m s}^{-2});$$

$dp$ is the layer thickness (in pressure units); $S_f$ is the virtual salt flux per unit of horizontal area divided by the reference specific volume $\alpha_0$ ($10^{-3}$ m$^3$ kg$^{-1}$); $P$, $E$ and $R$ represent the precipitation, evaporation and river input contributions, respectively (translated to m s$^{-1}$). If freshwater is to be added, $E < P + R, S_f < 0$ and the salinity decreases. If freshwater is to be removed, $E > P + R, S_f > 0$ and the salinity increases.

This implementation does not take into account the mass of freshwater that is introduced / removed, but only the resulting density changes. In this study, the
parameterization of the actual river mass inflow has been introduced by including an additional term in the barotropic pressure calculations. During the vertical mixing calculations at every baroclinic time step and at the grid points where the river inflow is imposed, the “barotropic pressure” term \((p'_b \eta)\) (see Bleck and Smith, 1990) is updated as 
\[
p'_b \eta = p'_b \eta + (dt_{b_{01}} \cdot g \cdot Q_{\text{river}}) \cdot \left( \alpha_0 \cdot A_{\text{grid}} \right),
\]
where \(Q_{\text{river}}\) is the freshwater flux (m\(^3\) s\(^{-1}\)) and \(A_{\text{grid}}\) is the horizontal grid area (m\(^2\)). This new formulation takes into account the pressure exerted by the mass of freshwater that is “virtually” introduced via the river inflow, and adds that contribution to the barotropic pressure of the water column. In theory, a river discharging freshwater into a more saline environment will generate a baroclinic anomaly (due to the density change) and a barotropic anomaly (due to the freshwater volume). Moreover, momentum is introduced into the system by the river inflow itself, at the head of the estuary. Although the freshwater volume anomaly is not taken into account in HYCOM, the barotropic pressure \((p'_b, \eta)\) exerted by this anomaly is calculated. Thus estuarine momentum naturally develops via the baroclinic and barotropic circulation induced by the freshwater inflow. The lack of river inflow momentum at the head of the estuary should not be detrimental to our model results. In the current domain configuration (section 2.3), a river volume inflow of 1000 m\(^3\) s\(^{-1}\) across the estuarine area \((3 \times 10^5 \text{ m}^2)\) would generate velocities of 0.003 m s\(^{-1}\), which are much smaller than the velocities that develop inside the estuary (reaching 0.2 m s\(^{-1}\), see section 2.4).

In the original HYCOM code, the freshwater flux due to river inflow was treated similarly to precipitation, injected in the first model layer. This could cause problems if
the model layers are too thin at river grid points, leading to a low salinity spike if the freshwater discharge is too high or if the model mixing is unable to mix the freshwater vertically or horizontally. Moreover, in nature, riverine waters occupy an upper layer of finite depth that changes under variable discharge and is also influenced by the available mixing conditions. An updated parameterization of river inflow in HYCOM was motivated by the present study and the user options described below (as well as the river mass inflow option discussed above) are available in the latest HYCOM code releases (A. Wallcraft, personal communication). The user is thus given the option to increase the downward penetration of the river inflow, effectively mixing the freshwater down to a specific depth. Another option is to increase the lateral spreading of the river inflow over specified cells. These are two ways of reducing the low salinity spike that may be created when all river discharge is concentrated in a few grid points. The physical meaning of the above modifications is to allow for additional vertical and horizontal mixing that would normally be available in realistic forcing experiments, where buoyancy is not the only external forcing.

2.2.2 Vertical mixing schemes

A detailed description of the vertical mixing schemes present in HYCOM is available in Halliwell (2004); an evaluation of their performances in low-resolution climatological simulations of the Atlantic Ocean is also given. In this study the three “continuous” differential models (which govern vertical mixing throughout the water column, not only at the mixing layer) are employed, namely the K-Profile Parameterization (KPP, Large et al., 1994), the NASA Goddard Institute for Space Studies level 2 turbulence closure
(GISS, Canuto et al., 2001, 2002) and the Mellor-Yamada level 2.5 turbulence closure (MY2.5, Mellor and Yamada, 1982).

### 2.2.3 Hybrid vertical coordinate grid generator

The foundation of the hybrid vertical coordinate system is the work by Bleck and Boudra (1981) and Bleck and Benjamin (1993). Each vertical layer in HYCOM is assigned a target density. At the end of each baroclinic time step, the model checks the calculated layer density against its target value and, if they differ, it tries to restore the former to the latter by allowing vertical movement of layer interfaces and vertical mass fluxes between them. However, if the vertical migration of grid points creates a crowding of coordinate surfaces, the model will produce (on a chosen number of upper \( n_{\text{hyb}} \) layers) a smooth transition from the isopycnal to the cartesian, fixed domain. This crowding is evaluated through a minimum thickness enforcement specified by the user, using a “cushion” function defined in Bleck (2002). Layers that transit to cartesian levels are then allowed to change their density freely and are no longer isopycnal. Likewise, the transition to sigma levels occurs only in the same \( n_{\text{hyb}} \) layers, where again a minimum thickness condition is evaluated. The choices of coordinate separation constrains that control the transition among the coordinate choices is left to the user. That allows different vertical coordinate possibilities, such as isopycnic-cartesian, isopycnic-sigma or fully hybrid domains. The model can also be run in purely isopycnal, cartesian or sigma mode. This flexibility was explored by Chassignet et al. (2003) in North Atlantic basin experiments, and will be also explored in the experiments presented herein.
2.3 Box model domain set-up

Following the idealized approach of the previously mentioned river plume studies, a box-like domain has been designed. It consists of a mid-latitude, f-plane (Coriolis parameter \( f = 10^{-4} \text{ s}^{-1} \)) rectangular flat bottom basin that is approximately 200 km long in the across-shore direction and 500 km long in the along-shore direction. The west boundary of the basin is a closed wall, whereas the north, east and south lateral boundaries are open water points. A 20 km long, 15 km wide simplified estuarine channel is located at the coast (Figure 2.1). The horizontal grid spacing is 2.5 km by 2.5 km throughout the domain. The basin water is initially homogeneous, with salinity of 35 and temperature of 28 °C, and it is in a state of rest. At the head of the estuarine channel, a freshwater discharge of 1,000 m\(^3\) s\(^{-1}\) is imposed (zero salinity, temperature 28 °C). The flow field is left to evolve for a period of 60 days. At the offshore lateral open boundaries, the baroclinic structure of the flow is relaxed to the basin initial state, whereas the barotropic structure is solved through the method of characteristics (Browning and Kreiss 1982; 1986). At the closed lateral boundary (the coast), a no-slip condition is applied to parallel velocities, and normal velocities are set to zero. At the bottom, momentum is dissipated by a quadratic bottom drag (drag coefficient \( C_d = 3 \times 10^{-3} \)), using a bottom velocity \( u_b \) that represents the average velocity in a slice of water situated just above the bottom. If the KPP vertical mixing scheme is used, this thickness is determined by a bottom boundary layer parameterization that is an adaptation of the algorithm used for the surface boundary layer (Halliwell et al., 2009). Otherwise, it is set to 1 m. Salt flux normal to the bottom and the coast is zero. At the surface, the river precipitation bogus is the only forcing mechanism.
Figure 2.1 Model configuration. Left: Idealized box-like basin. The thick short black line denotes where river discharge is imposed (estuary head). Land area is shaded. Center: Zoom of the estuarine region. Dashed lines show position of vertical sections where model results are evaluated: along the estuary and across the basin (1, extending to the east boundary), across the estuary mouth (2) and along-shore in the vicinity of the estuary (3). Right: Vertical section of bathymetry from the sloping bottom setup showing cartesian levels (thin lines, along section 1); additional hybrid configurations are exhibited in Figure 2.13.

Flat and sloping bottom topography configurations are employed. In the flat bottom set-up, 16 cartesian levels are used in the vertical and the bottom is 20m deep everywhere (layer spacing is 1.25m), which is a reasonable approximation to average inner-shelf depths. In the sloping bottom set-up, a gentle slope starts at the coast line. The estuary remains 20m deep and the shelf bottom depth goes down to 100m within 200km in the offshore direction. Now, a temperature profile is imposed such that the temperature is 28°C at the surface and decreases 0.5°C every 5m. The presence of a sloping bottom and ambient stratification allow the use of a fully hybrid vertical level set up in this domain. The standard configuration for sloping bottom experiments is 24 fixed cartesian levels in the upper 30m (layer spacing is 1.25m) and 6 isopycnal levels from 30m to 100m (Figure 2.1). This configuration enforces vertical resolution in the upper water column to be the
same as in the flat bottom experiments, and avoids the interaction of deep isopycnal levels with the surface plume dynamics while taking the advantage of the flexible vertical coordinate system in HYCOM. Sloping model experiments with all sigma layers (30 bottom following layers) and with a combination of sigma in the upper 50m and isopycnal below, were also employed.

2.4 River plume experiments

A series of experiments were employed for both the flat and sloping bottom basin configurations described in section 2.3. Twelve combinations of model parameters were explored, see attributes in Table 2.1.

Table 2.1 Summary of the attributes from the study experiments. Downward penetration is given in percentage of the water column and in meters (parentheses). Lateral spreading is classified as none, short (4 grid points, half estuary length) or large (7 grid points, full estuary length). KPP background vertical mixing is characterized as standard (no modifications) or as Enhanced $K_{iw}$ (salinity diffusivity due to background internal wave mixing equals $10^{-4}$ m$^2$ s$^{-1}$) plus region where it is applied. See sections 2.4.2 and 2.4.3 for details.

<table>
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<th>Experiment</th>
<th>Downward penetration (pntr)</th>
<th>Lateral spreading (sprd)</th>
<th>KPP background vertical mixing</th>
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</tr>
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</tr>
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<td>Enhanced $K_{iw}$ over full estuary length</td>
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2.4.1 Control experiment

A control experiment is configured to serve as “reference” against other experiments, as well as to evaluate the general plume dynamics. It employs the basic HYCOM river
parameterization (precipitation bogus, no downward penetration, no lateral spreading) with mass inflow parameterization included. Vertical mixing is governed by the KPP scheme.

The development of the control buoyant plume in the 20 m deep, flat bottom basin (Control-flat) follows the general description in the literature (Chao and Boicourt, 1986; Chao, 1988a; Kourafalou et al., 1996a; Garvine, 1999, among others). The buoyant plume reaches the shelf by day 5, while making an anticyclonic turn at the mouth of the estuary (not shown). A recirculating bulge develops and grows in time, followed by a coastal current propagating in the downstream direction. The river plume is well developed after 60 days (Figure 2.2), with a coastal current (~ 6-8 cm s\(^{-1}\)) that presents a developed meandering character due to barotropic/baroclinic instabilities (Oey and Mellor, 1993). An upstream penetration is also observed. Taking the plume boundary to be the 34.9 salinity isoline, the nose of the coastal current has reached 202.5 km south of the river mouth. The offshore extension of the bulge (72.5 km) is larger than the coastal current width (62.5 km), which represents a supercritical plume case (Chao, 1988a). A weaker undercurrent (not shown) runs in the upstream direction opposite to the downstream coastal current below 6 – 8 m, with velocities around 3 cm s\(^{-1}\). An across-shore vertical section of salinity contours (Figure 2.2) shows that the buoyant flow forms a 2-layer structure, with a surface buoyant layer on top of denser ambient water. This is the case of a surface-advect ed plume (Yankovsky and Chapman, 1997), which presents a recirculating bulge, a coastal current close to the coast and very little contact with the bottom. The maximum barotropic velocity at the bulge reaches 3 cm s\(^{-1}\), well below the maximum baroclinic velocity (12 cm s\(^{-1}\)).
Figure 2.2 Upper: Sea Surface Height contours in mm (left), Sea Surface Salinity contours (middle) and near surface velocity vectors in cm s$^{-1}$ (right) from the Control-flat experiment at day 60 (part of the model domain shown). Lower: Along-estuary/across-shore salinity vertical structure along section 1 (marked in Figure 2.1). The plume boundary (34.9) is represented by a white line. Salinity values less than 25 (inside the estuary) are not shown. Vertical black line denotes the position of the estuary mouth.

In sloping bottom conditions (Control-slope, Figure 2.3), the plume develops a recirculating bulge that is elongated in the upstream direction and shortened in the offshore direction in comparison to Control-flat. An enhanced upstream intrusion develops, as most of the buoyant outflow turns to the left upon exiting the estuary, before turning anticyclonically and merging to the coastal current (which exhibits less meandering). The enhancement of the upstream and shortening of the offshore intrusions have been reported in previous studies (Kourafalou et al., 1996a; Garvine, 1999), and the changes in the bulge structure suggest the effect of potential vorticity constrains imposed by the bottom slope (Chao, 1988a).
Figure 2.3 As Figure 2.2, but for the Control-slope experiment.

2.4.2 Prescribed river inflow distributions inside the estuary

We examine changes in the development and structure of the river plume when the river inflow distribution is prescribed inside the estuary. This is accomplished by employing the river parameterization options of enhanced downward penetration and horizontal spreading of the river inflow (described in section 2.2.1), which effectively change the vertical and horizontal mixing of the buoyant plume at the source. This redistribution of freshwater input is expected to change the properties of the buoyant outflow at the estuary mouth and impact the development of the river plume in the receiving basin. We enhance the downward penetration of the river inflow to 20% (4 m) and 40% (8 m) of the water column and impose a short lateral spreading (half the estuary length) and a large lateral spreading (the entire estuary length). Together with the Control case, nine different parameter combinations for the study experiments are employed (Table 2.1); “expt(s)” will abbreviate “experiment(s)” thereafter. All expts in this group
employ the KPP vertical mixing scheme. Snapshots of the plume Sea Surface Salinity (SSS) and the bulge near surface velocity vectors (both at day 60) are presented in Figures 2.4 and 2.5, respectively. The extensions of each plume upstream ($L_u$), downstream ($L_d$) and offshore ($L_o$) intrusions are also depicted in Figure 2.4.

Figure 2.4 Sea Surface Salinity contours from experiments with variable distribution of river inflow inside the estuary, at day 60 (part of the model domain shown). The plume boundary (34.9) is represented by a white line. Salinity values less than 25 (inside the estuary) are not shown. The upstream ($L_u$), downstream ($L_d$) and offshore ($L_o$) plume intrusions for each case are displayed next to the plots. Downward penetration (pntr) and horizontal spreading (sprd) configurations that characterize each experiment are also presented.
Figure 2.5 Near surface velocity vectors from experiments with variable distribution of river inflow inside the estuary, at day 60 (part of the model domain shown). Downward penetration (pntr) and horizontal spreading (sprd) configurations that characterize each experiment are presented. Vectors are plotted every other grid point for better visualization.
a) Variable downward penetration and no lateral spreading

Cases of variable downward penetration (0%, 20% and 40%) with no lateral spreading of the river inflow are shown in Figure 2.4 (upper panels). There is a considerable change in the shape and extension of the plume when the downward penetration of the river discharge is enhanced to 20% (expt Riv1b-flat); the anticyclonic bulge grows in size, with a larger offshore extension and a more circular shape. A stronger coastal current is present, with more meanders and a longer downstream extension. Interestingly, the upstream penetration is much reduced. This pattern is enhanced when this mixing is forced down to 40% of the water column (Riv1c-flat). In all cases, the buoyant outflow is denser with increasing downward penetration of the inflowing river discharge. The surface circulation from these experiments (Figure 2.5, upper panels) demonstrates a progressive strengthening of the buoyant outflow with increasing downward penetration of the river inflow by 20% (Riv1b-flat) and 40% (Riv1c-flat). A shift in the position and direction of the estuary outflow is also observed, which has clear impact on the shape and location of the offshore bulge. The plume outflow is concentrated in the northern wall of the estuary mouth and exits in a straight path (Control-flat), and as it intensifies it spreads across the estuary mouth (Riv1b-flat) and develops an enhanced anticyclonic turning (Riv1c-flat).

The effect of the bottom slope is demonstrated on the Riv1c case, depicting a marked impact on both the estuary outflow and the bulge development (Riv1c-slope, Figures 2.4 and 2.5). In this case the buoyant outflow does not present an anticyclonic veering, as it exits the estuary in a straight path (similar to Control–slope). The bulge presents a
marked upstream displacement and upstream flow intrusion, which are accompanied by a shorter offshore extent. As the bottom slope “squeezes” the buoyant flow area against the coast, the plume is elongated in the along-shore direction and develops a longer coastal current region.

**b) Variable downward penetration and short lateral spreading**

The general pattern in the surface salinity field discussed above (reduction of the upstream intrusion, larger offshore bulge and larger downstream penetration) is also observed when the downward penetration of the river inflow is increased in the presence of short lateral spreading (Figure 2.4, middle panels). However, some distinctions from the same cases with no lateral spreading are observed. The upstream intrusion enhances from the expt Control-flat to Riv2a-flat and the bulge is slightly less circular in expt Riv2b-flat than in Riv1b-flat. No major changes are observed in the salinity field from expts Riv1c-flat and Riv2c-flat, except the outflow from Riv2c-flat is less buoyant and the bulge is larger than in experiment Riv1c-flat. The general pattern of buoyant outflow intensification and development of an anticyclonic veering is also observed when downward penetration increases from 0 to 40% (Figure 2.5, middle panels). Finally, the trend imposed by the bottom slope in expt Riv1c-slope is also observed in Riv2c-slope, where the bulge is displaced in the upstream direction, the outflow does not develop an anticyclonic veering and the coastal current region is elongated.
c) Variable downward penetration and large lateral spreading

The plume surface salinity field changed considerably when the vertical penetration of the river inflow was varied while employing large spreading (Figures 2.4 and 2.5, lower panels), as now both vertical and horizontal salinity gradients were impacted. The same pattern of enhancement of the upstream intrusion is observed at 0% downward penetration (Riv3a-flat), which vanishes at 20% downward penetration (Riv3b-flat) as the bulge becomes less circular and less distinct from the coastal current (in comparison to Riv2b-flat). The largest changes in plume shape were observed at 40% downward penetration (Riv3c-flat), when the offshore bulge did not develop and the plume turned abruptly to the right and moved downstream forming a coastal current that started unidirectional and then developed a meandering pattern, starting with a feature resembling a secondary bulge due to the large amount of low salinity water that has leaked along the coast. The near surface velocity field (Figure 2.5, lower panels) confirms that the outflow from expt Riv3c-flat developed an abrupt right turn at the estuary mouth and all plume waters were deflected southward, increasing the downstream coastal current penetration. Conversely, the development of the plume in sloping bottom conditions (Riv3c-slope) is considerably different as an offshore bulge develops in front of the estuary, the buoyant outflow exits in a straight path and a slight upstream intrusion is observed. The presence of a slope appears to overwhelm the impact of lateral and/or vertical mixing inside the estuary, as suggested by the similarities of the Riv2c-slope and Riv3c-slope expts, in contrast to their clearly different flat bottom counterparts.
2.4.3 Enhanced vertical mixing inside the estuary

The impact of changes in the vertical mixing at the freshwater source on the development of the river plume is also investigated. Instead of mixing the river freshwater by redistributing it, the background mixing within the KPP vertical mixing scheme was enhanced. Specifically, the vertical salinity diffusivity due to background internal wave mixing $K_{iw}$ was increased to $10^{-4}$ m$^2$ s$^{-1}$ (10 times its standard background value) inside the estuary. Three expts were performed, where $K_{iw}$ was enhanced in three distinct regions: at the estuary head, from the head to half the estuary length and from the head to the estuary mouth (full estuary length). The choice to change $K_{iw}$ and not other aspects of the KPP vertical mixing scheme was based on the fact that the background internal wave mixing is the main contributor to vertical mixing in the study experiments (see discussion below). In order to access the sensitivity of the plume structure to the choice of vertical mixing scheme, twin expts of the Control-flat case that employ the MY2.5 and GISS vertical mixing schemes were performed.

The enhancement of $K_{iw}$ inside the estuary effectively impacts the outflow properties and the development of the buoyant plume. SSS and near surface velocity vectors from this group of experiments (Figure 2.6) demonstrate that as the estuarine region was mixed through enhanced $K_{iw}$, the outflow became less fresh, progressively developed an anticyclonic turning and decreased the plume upstream penetration. The offshore bulge was clearly impacted by those changes, as it is shown to shift downstream and finally vanish with the outflow being deflected in the downstream direction. The surface fields
from the half-estuary (Mix4b-flat) and full-estuary (Mix4c-flat) cases resemble those impacted by 40% downward penetration at short (Riv2c-flat) and large (Riv3c-flat) horizontal spreading, respectively. In the presence of the bottom slope (Mix4c-slope), the plume evolves to the same structure as in the expt Riv3c-slope.

![Figure 2.6](image)

**Figure 2.6** Sea Surface Salinity contours (upper) and near surface velocity vectors (lower) from experiments with enhanced mixing (increased $K_{iw}$) inside the estuary, at day 60 (part of the model domain shown). The plume boundary (34.9) is represented by a white line. Salinity values less than 25 (inside the estuary) are not shown. The mixing information that characterizes each experiment is shown next to each plot (see section 2.4.3 for details). Vectors are plotted every other grid point for better visualization.

The mixing expts revealed that the plume structure was not sensitive to the choice of vertical mixing schemes. Employing the MY2.5 or the GISS vertical mixing schemes produced river plumes that had the same vertical salinity structure as the Control experiment (KPP), as well as the SSS and near surface currents (not shown). This finding suggests that the vertical eddy mixing coefficients computed by each scheme are
approximately the same and that vertical mixing is being controlled by the same process. In this study, two common components of the ocean interior mixing should be important: the unresolved background internal wave mixing and shear instability. In HYCOM and for all vertical mixing schemes employed here, the background internal wave mixing is parameterized through constant coefficients: scalar diffusivity is equal to $10^{-5} \text{ m}^2 \text{ s}^{-1}$, whereas background viscosity is $10^{-4} \text{ m}^2 \text{ s}^{-1}$. The parameterization of shear-driven mixing depends on the choice of vertical mixing scheme, but it is commonly related to a critical Richardson number below which the parameterization is activated. The Richardson number at model interfaces from the experiments with the three different vertical mixing schemes was calculated and compared to the Richardson number threshold from each scheme (not shown). All values of the calculated Richardson number within the buoyant plume were above the critical value below which mixing occurs, which suggests that in the study experiments the parameterization of shear-induced mixing is not triggered for any choice of scheme, and that the buoyant plume vertical mixing is controlled by the background internal wave parameterization.

2.5 Discussion of results

2.5.1 Variability of outflow properties

Enhanced vertical and horizontal mixing of the river inflow inside the estuary impacted the river mouth conditions and the structure of the buoyant outflow. Figure 2.7 shows across-estuary sections (along section 2 at the estuary mouth, Figure 2.1) of Sea
Surface Height (SSH), salinity and along-estuary (u) velocity for selected experiments, at day 60. As expected, progressively increasing the downward penetration of the river inflow (Riv2a,b,c–flat) or enlarging the estuary area with enhanced $K_{iw}$ (Mixa,b,c–flat) generated plume outflows that were deeper and denser. A coupled upper outflow / lower inflow structure is observed, which represents the classic gravitational circulation verified in previous numerical studies (Chao, 1988a). This gravitational circulation was enhanced as the mixing of river inflow inside the estuary increased, and the upper outflow and bottom inflow increased in magnitude. This effectively enhanced the outflow transport $T_f = \int_{A_f} u dA_f$ ($A_f$ is the upper outflow area where $u$ is positive, see Figure 2.7 for selected expts and Table 2.2 for all flat bottom expts). $T_f$ increased 146% from Riv2a to Riv2b, 55% from Riv2b to Riv2c, 70% from Mix4a to Mix4b and 18% from Mix4b to Mix 4c. All other “Riv” flat bottom cases present increases in $T_f$ that are of the same order as in expts Riv2a,b,c-flat. Across-estuary (v) velocity distributions (not shown) demonstrate the development of a veering pattern of the plume outflow, which intensifies as freshwater mixing inside the estuary is enhanced. The above changes were accompanied by a progressive steepening of the salinity isolines, which was also followed by a slight steepening of the across-estuary SSH. The plume vertical structure suggests that the outflow is in geostrophic balance, which is strengthened with enhanced mixing of riverine waters inside the estuary. The presence of a bottom slope in the basin also impacted the vertical structure of the outflow. As the outflow shifted to the north side of the estuary mouth, the low-salinity area was enlarged and the isopycnals and SSH became flatter on the south side of the mouth (Riv2c–slope and Mix4c–slope). The plume outflow structure of the Control–slope case is the same as in the presence of a flat
bottom. Interestingly, the slope experiments presented smaller $T_f$ in comparison to their flat bottom counterparts (Figure 2.7), which suggests a two way interaction between the dynamics of the buoyant flow in the estuary and in the receiving basin (in Figure 2.7, a comparison between Control–slope and Riv2a-flat is valid because the plume does not really change from the Riv2a-flat to the Control-flat case).

Figure 2.7 Sea Surface Height (SSH, in mm), across-estuary vertical salinity structure (colors) and along-estuary velocity ($u$, cm s$^{-1}$, solid for positive/offshore and dashed for negative/onshore contours) along section 2 (estuary mouth) from selected experiments, at day 60. The configurations that define each experiment and the outflow transport $T_f$ for each case are shown.
Table 2.2 Outflow transport $T_f$, outflow upper layer 1 and inflow lower layer 2 vertical mean values of density $\rho$, along-estuary ($u$) and across-estuary ($v$) velocities and layer thickness $h$, for all flat bottom experiments at day 60. Layers 1 and 2 average values were calculated from a vertical profile located at the core of the surface outflow at the estuary mouth. See Table 2.1 for attributes from experiments and section 2.5.1 for details on the calculations.

<table>
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<th>Expts.</th>
<th>$T_f$ $*10^3$ m$^3$s$^{-1}$</th>
<th>$\rho_1$ sigma</th>
<th>$u_1$ m s$^{-1}$</th>
<th>$v_1$ m s$^{-1}$</th>
<th>$h_1$ m</th>
<th>$\rho_2$ sigma</th>
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<th>$v_2$ m s$^{-1}$</th>
<th>$h_2$ m</th>
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</tbody>
</table>

In order to summarize the main geophysical properties of the different outflows, we calculated common non-dimensional numbers from all flat bottom experiments. We concentrated on the flat bottom expts, as they facilitate comparisons to 2-layer analytical models. Such calculations involved an approximation of the plume outflow to a two-layer formulation defined by $h_r$, the depth where the along-estuary velocity ($u$) becomes negative. We defined an outflow upper layer with thickness $h_1$, density $\rho_1$ and velocities $u_1$ and $v_1$, and an inflow lower layer with thickness $h_2$, density $\rho_2$ and velocities $u_2$ and $v_2$. For each experiment, vertical profiles of $u$, $v$ and $\rho$ were extracted at the location of the core of the surface outflow at the estuary mouth; $\rho_1$, $\rho_2$, $u_1$, $u_2$, $v_1$ and $v_2$ were calculated as vertical mean values from the model grid points that are within each layer $h_1$ and $h_2$ (above and below $h_r$, respectively) and are presented in Table 2.2. These calculations did not involve averages in the across-estuary direction because in some cases the outflow velocity field is clearly concentrated on one side of the estuary channel and an average in the y direction would underestimate the outflow velocity. We calculated the gradient
Richardson number $Ri = N^2/S^2$, the Froude number $Fr = |\vec{V}_i|/c_i$, the inlet Rossby number $Ro_i = |\vec{V}_i|/(fW)$ and the inlet Kelvin number $K_i = W/R_{di}$.

$N^2 = \left( -\frac{g}{\rho_o} \right) \left( \rho_1 - \rho_2 \right) / \Delta z$ is the squared stratification frequency, $S^2 = \left( u_1 - u_2 / \Delta z \right)^2 + \left( v_1 - v_2 / \Delta z \right)^2$ is the squared velocity vertical shear, $\nu = \sqrt{g' \left( h_1 * h_2 \right) / \left( h_1 + h_2 \right)}$ is the phase speed of long internal gravity waves and $R_{di} = c_i / f$ is the internal radius of deformation; $g' = g \left( \rho_2 - \rho_1 \right) / \rho_o$ is the reduced gravity, $g$ is the gravitational acceleration ($9.806 \text{ m s}^{-2}$), $\rho_o$ is the initial ambient density ($1022.40 \text{ kg m}^{-3}$), $\Delta z$ is the distance from the surface ($z = 0$, axis positive upwards) down to the interface of the layers (equal to $h_1$); $|\vec{V}_i|$ is the length of the upper layer velocity vector, $W$ is the estuary width (15 km) and $f$ is the Coriolis parameter ($10^{-4} \text{ s}^{-1}$).

The non-dimensional numbers presented in Table 2.3 summarize and corroborate with the properties of the distinct outflows. $K_i$ was larger than 1 in all experiments, which indicates that the study experiments are all large scale discharges and that the dynamics at the estuary mouth are affected by the earth’s rotation. The fact that $Ro_i$ is smaller than 1 in all experiments suggests that advection plays a secondary role in the dynamics governing the immediate vicinity of the estuary outflow. The outflows should be approximately in geostrophic balance, which is in agreement with the salinity vertical structures presented in Figure 2.7 and will be explored further in the next section.

Considerably large $Ri$ ($>20$) was found in experiments that did not have enhanced vertical mixing of the river inflow (expts Control-flat, Riv2a,3a-flat), and progressively decreased to values below 3 (expts Riv2c,3c-flat and Mix4c-flat) as we enhanced vertical
mixing (via downward penetration or larger $K_{in}$). As expected, this was followed by an opposite behavior of $Fr$. Vertical stratification is important to the flow dynamics (all cases presented $Fr$ less than one) and $Fr$ increases with increasing downward penetration, meaning that the importance of stratification (here measured by $c_i$) decreases relative to the horizontal advection of the flow. All cases fall in the category of large scale buoyant flows (Garvine, 1995) which are characterized by large $K_i$, small $Ro_i$ and small $Fr$.

Table 2.3 Non-dimensional numbers calculated for all flat bottom experiments at day 60. See Table 2.1 for attributes from experiments. $Ri$: Gradient Richardson number; $Fr$: Froude number; $Ro_i$: Inlet Rossby number; $K_i$: Inlet Kelvin number; The internal deformation radius $R_{di}$ (km), the squared stratification frequency $N^2 (s^{-2} * 10^{-2})$ and the squared vertical velocity shear $S^2 (s^{-2} * 10^{-2})$ are also shown. Numbers were calculated using values presented in Table 2.2. See section 2.5.1 for details.

<table>
<thead>
<tr>
<th>Experiment</th>
<th>$Ri$</th>
<th>$Fr$</th>
<th>$Ro_i$</th>
<th>$K_i$</th>
<th>$R_{di}$</th>
<th>$N^2$</th>
<th>$S^2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Control</td>
<td>22.10</td>
<td>0.20</td>
<td>0.04</td>
<td>4.79</td>
<td>3.13</td>
<td>0.36</td>
<td>0.01</td>
</tr>
<tr>
<td>Riv1b</td>
<td>6.82</td>
<td>0.35</td>
<td>0.06</td>
<td>5.62</td>
<td>2.67</td>
<td>0.26</td>
<td>0.04</td>
</tr>
<tr>
<td>Riv1c</td>
<td>4.40</td>
<td>0.44</td>
<td>0.09</td>
<td>5.10</td>
<td>2.94</td>
<td>0.24</td>
<td>0.05</td>
</tr>
<tr>
<td>Riv2a</td>
<td>24.24</td>
<td>0.19</td>
<td>0.04</td>
<td>4.43</td>
<td>3.38</td>
<td>0.43</td>
<td>0.02</td>
</tr>
<tr>
<td>Riv2b</td>
<td>3.30</td>
<td>0.50</td>
<td>0.08</td>
<td>5.89</td>
<td>2.54</td>
<td>0.34</td>
<td>0.10</td>
</tr>
<tr>
<td>Riv2c</td>
<td>2.07</td>
<td>0.71</td>
<td>0.13</td>
<td>5.50</td>
<td>2.73</td>
<td>0.21</td>
<td>0.10</td>
</tr>
<tr>
<td>Riv3a</td>
<td>25.29</td>
<td>0.18</td>
<td>0.04</td>
<td>4.33</td>
<td>3.46</td>
<td>0.45</td>
<td>0.02</td>
</tr>
<tr>
<td>Riv3b</td>
<td>3.63</td>
<td>0.45</td>
<td>0.09</td>
<td>4.87</td>
<td>3.08</td>
<td>0.50</td>
<td>0.14</td>
</tr>
<tr>
<td>Riv3c</td>
<td>2.80</td>
<td>0.45</td>
<td>0.08</td>
<td>5.95</td>
<td>2.52</td>
<td>0.15</td>
<td>0.05</td>
</tr>
<tr>
<td>Mix4a</td>
<td>16.17</td>
<td>0.23</td>
<td>0.05</td>
<td>4.49</td>
<td>3.34</td>
<td>0.32</td>
<td>0.02</td>
</tr>
<tr>
<td>Mix4b</td>
<td>4.17</td>
<td>0.50</td>
<td>0.09</td>
<td>5.74</td>
<td>2.61</td>
<td>0.25</td>
<td>0.06</td>
</tr>
<tr>
<td>Mix4c</td>
<td>2.75</td>
<td>0.50</td>
<td>0.07</td>
<td>6.66</td>
<td>2.25</td>
<td>0.19</td>
<td>0.07</td>
</tr>
</tbody>
</table>

Certain relationships between the non-dimensional numbers, outflow properties and plume length scales are observed. As expected, stronger outflow transports ($T_f$) were associated with smaller $Ri$ values (enhanced turbulent mixing and intensified estuarine gravitational circulation) which also led to stronger coastal current signals and longer downstream intrusions. Moreover, the degree of upstream intrusion was related to the buoyancy of the outflow. In the experiments that had only horizontal redistribution of the river inflow (expts Control-flat, Riv2a,3a-flat), longer upstream penetrations were
observed with larger buoyancy (smaller vertical turbulent mixing, larger $Ri$). All other experiments that had less buoyant outflows presented no upstream intrusion. Chapman and Lentz (1994) reported the same relationship, where the rate of upstream movement of a plume was highly dependent on the outflow density anomaly. Finally, results suggest a positive relationship between the turning of the outflow and $K_i$. The anticyclonic turning of the outflow became stronger when $K_i$ progressively increased, which happened with increasing vertical mixing throughout the estuary (expts Riv3a,b,c-flat and Mix4a,b,c-flat).

2.5.2 Dynamical balance of the outflow

As mentioned before, all experiments reproduced large-scale plume outflows because the inlet Kelvin number ($K_i$) was always larger than 1 (Table 2.3). The low inlet Rossby number ($Ro_i$) conditions and the vertical structure of the outflows (Figure 2.7) suggest that the dynamical balance at the estuary mouth is (at a first order) geostrophic and that the outflow is in thermal wind balance. The steepening of the salinity isolines with increasing vertical mixing of freshwater (flat bottom experiments) also suggests an intensification of the geostrophic balance. Major terms of the momentum balance were computed at the location of the core of the near surface outflow (estuary mouth) and the predominant balance was geostrophic as shown by the length of each of the vectors $|c\mathbf{\alpha}r| = \sqrt{\left(- f \frac{\partial}{\partial y} + (f \frac{\partial}{\partial x}) \right)^2 + \left(- \frac{\rho_0}{\rho} \frac{\partial p}{\partial x} \right)^2}$ in Figure 2.8. The term $|\mathbf{a\alpha}c\mathbf{\alpha}cel| = \sqrt{(Du/Dr)^2 + (Dv/Dr)^2}$ is also presented in Figure 2.8, exhibiting minimal
contribution to the momentum balance. In agreement with the across-estuary section from Figure 2.7, the magnitude of the geostrophic balance terms increased as the vertical mixing of freshwater was enhanced (expts Riv2a,b,c-flat). In the case of expts Mix4a,b,c-flat, the difference between the pressure gradient and the Coriolis forces increased (as shown in Mix4c-flat), which is followed by the increased contribution of the acceleration term. Although the buoyant outflow evolved into a different configuration in the presence of a bottom slope (Figures 2.5 and 2.6), the geostrophic character of the surface outflow did not change (Riv2c-slope and Mix4c-slope).
Figure 2.8 Upper: Magnitude of each of the major momentum balance vectors ($|\mathbf{c}\mathbf{\cdot}\mathbf{r}| = \sqrt{(-f\mathbf{\cdot}\mathbf{y})^2 + (f\mathbf{\cdot}\mathbf{u})^2}$, $|p\mathbf{g}\mathbf{f}| = \sqrt{(-\rho_o^{-1}\partial\mathbf{p}/\partial\mathbf{x})^2 + (-\rho_o^{-1}\partial\mathbf{p}/\partial\mathbf{y})^2}$) and $|\mathbf{a}\mathbf{c}\mathbf{c}\mathbf{e}\mathbf{l}| = \sqrt{\left(D\mathbf{u}/D\mathbf{t}\right)^2 + \left(D\mathbf{v}/D\mathbf{t}\right)^2}$ where $D(u,v)/Dt = \partial(u,v)/\partial t + \mathbf{V} \cdot \nabla(u,v)$) from selected experiments. Values were extracted from section 2 (estuary mouth) at the location of the near surface outflow core, at day 60. Middle: Correspondent individual components of each geostrophic balance term normalized by the associated vector length ($|\mathbf{c}\mathbf{or}_x|/|\mathbf{c}\mathbf{\cdot}\mathbf{r}|$, $|\mathbf{c}\mathbf{or}_y|/|\mathbf{c}\mathbf{\cdot}\mathbf{r}|$, $|p\mathbf{g}\mathbf{f}_x|/|p\mathbf{g}\mathbf{f}|$ and $|p\mathbf{g}\mathbf{f}_y|/|p\mathbf{g}\mathbf{f}|$). Lower: Average positive (upward) and negative (downward) vertical velocity inside the estuary, at day 60.
Because this is a zonal estuary, the primary geostrophic balance of the outflow is in the across-estuary (y) direction (\( f \mathbf{u} = -\rho_0^{-1} \frac{\partial p}{\partial y} \)). The development of the anticyclonic turning as vertical mixing of estuarine waters is increased (Figures 2.5 and 2.6) suggests that the outflow also develops a geostrophic balance in the along-estuary (x) direction (\(- f \mathbf{v} = -\rho_0^{-1} \frac{\partial p}{\partial x}\)). The geostrophic balance components in each direction (x and y) normalized by their respective vector lengths (\(|c_{or_x}|/|c_{or_y}|\), \(|c_{or_x}|/|c_{or_y}|\), \(|p_{gf_x}|/|p_{gf_y}|\) and \(|p_{gf_x}|/|p_{gf_y}|\), Figure 2.8) show that the relative importance of the along-estuary components (x direction) indeed increased with enhanced mixing in all cases. This was followed by a slight decrease in the relative importance of the across-estuary components (y direction). The development of the along-estuary balance was suppressed in the presence of a sloping bottom (Riv2c-slope and Mix4c-slope), which corroborates with the lack of an anticyclonic veering by the outflow. Finally, the velocity contours and \(T_f\) values from Figure 2.7 suggest that the intensification of the estuarine gravitational circulation with enhanced freshwater mixing must be accompanied by an increase in the upward entrainment from the bottom inflow to the surface outflow. Average upward velocities inside the estuary (Figure 2.8) confirmed this pattern and show that they can be six times larger in the presence of enhanced freshwater mixing (expts Riv2a,c-flat). Very interestingly, the bottom slope worked against this intensification and promoted vertical velocities that were slightly smaller (Riv2c-flat and Riv2c-slope), as well as smaller \(T_f\) values (Figure 2.7). The average negative vertical velocity became more negative with increased vertical mixing, meaning that downward motions were also enhanced. The concomitant increase of both upward and downward motions suggests that the estuarine secondary circulation was intensified.


2.5.3 Topographic constrains on the development and transport of plume waters

The most pronounced impact of the bottom slope over the structure of the buoyant plume was the restriction of the offshore development and changes in the shape and position of the recirculating bulge (Figures 2.5 and 2.6). The bottom slope greatly impacted the buoyant outflow conditions at the river mouth, even though the estuary was still at flat bottom and only the “shelf” had varying topography. In particular, the outflow did not develop an anticyclonic veering when it encompassed the sloping bottom (expts Riv1c, 2c, 3c-slope and Mix4c-slope).

The question suggested by these findings is: How does the bottom slope change the properties of the geostrophic buoyant outflow and how does that affect the development of the anticyclonic bulge? This question is tackled by investigating the evolution of the buoyant plume in expts Riv2c–flat and Riv2c–slope. A series of snapshots of near surface velocity vectors (Figure 2.9) reveals that the surface plume “feels” the sloping bottom in its early stage of development. The buoyant outflow gradually shifted from a configuration with an anticyclonic veering to a configuration where it exited the estuary in a straight path and was concentrated on the northern side of the mouth. This pattern was accompanied by the development of a bulge in front of the estuary. This development was very different in the presence of a flat bottom, where the outflow maintained an anticyclonic veering configuration and exited the estuary as jet that was free to expand offshore, which ultimately led to the development of a large bulge south of the river mouth. These changes in the buoyant outflow circulation also reflected
modifications in the surface relative vorticity $\zeta = \left(\frac{\partial v}{\partial x}\right) - \left(\frac{\partial u}{\partial y}\right)$ (Figure 2.10) in the vicinity of the river mouth (along section 3, Figure 2.1). In the presence of a flat bottom, the $\zeta$ field remained constant throughout time and reflected the buoyant outflow jet with opposite $\zeta$ signals on each side of the flow. When the bottom slope was employed, $\zeta$ progressively changed in time and reflected the development of the recirculating bulge with a negative signal (anticyclonic) that is surrounded by positive signals to the north (cyclonic turning of the upstream intrusion) and to the south (cyclonic turning of the bulge intrusion at the coast), followed by a negative signal associated with a coastal current meander (see Figure 2.5).

Figure 2.9 Snapshots of near surface velocity vectors from Riv2c experiments (both from flat and sloping bottom conditions) starting on day 5 to day 30, every 5 days (part of the model domain shown). Vectors are plotted every other grid point for better visualization.
Figure 2.10 Hovmöller diagrams of surface relative vorticity $\zeta = \left( \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y} \right)$ ($10^{-5}$ s$^{-1}$, left panels) and of vertical velocity ($10^{-5}$ m s$^{-1}$) at 15m below the surface (model layer 12, right panels) from the Riv2c experiments (upper: flat bottom, lower: sloping bottom) along section 3 (vicinity of the estuary, Figure 2.1).

Numerical experiments performed by Chao (1998a) and Kourafalou et al. (1996a) investigated the impact of a sloping bottom on the structure of the river plume in a domain configuration similar to the one in this study (rectangular basin, idealized flat bottom estuary and gentle slope starting at the coast). Their results demonstrate that the bottom slope induced gain in anticyclonic vorticity for the plume due to enhanced upwelling and surface divergence. The higher surface elevation near the river mouth increased the along-shore (sloping northward) pressure gradient which enhanced the north extension of the plume, while the increased shoreward flow south of the mouth produced a stronger southward coastal current. Figure 2.9 suggests that this upstream shift is triggered as soon as the buoyant plume reaches the basin, between 5 and 10 days. Hovmöller diagrams of vertical velocity in the vicinity of the estuary mouth (also along section 3) demonstrate the differences in the vertical velocity pattern between flat and
slope conditions (Figure 2.10). Until approximately 10 days, the vertical velocity field from the flat and slope conditions were similar, although enhanced upwelling was observed in the presence of the slope. Both fields start to diverge after 10 days which is a result from the development of different circulation patterns. In flat bottom conditions, the vertical velocity field was characterized by a downwelling fringe around the bulge and an upwelling signal inside it (the section only captures the northern part of the bulge). In the presence of a slope, the upwelling signal inside the bulge (red region) was considerably stronger, which was followed by a strong downwelling signal (blue region) where the bulge circulation turns cyclonically to feed the coastal current, the latter presenting an upwelling signal (yellow region) associated with the meandering of the coastal current. This comparison suggests that indeed the enhanced upwelling is related to the northward displacement of the bulge, which could shift the buoyant outflow to a different configuration. This pattern is also observed in the expts Riv1c,3c-slope and Mix4c–slope and to a less degree in the Control–slope case (not shown).

The position of the offshore edge of the plume in time is compared between selected slope experiments and their flat bottom counterparts (Figure 2.11). Apart from the Control cases, the recirculating bulges in flat and sloping bottom conditions started to diverge from each other around day 15 and the bottom slope imposed significant upstream displacements of the bulge after 60 days. The above changes in the recirculating bulge also impacted coastal current properties such as the displacement of the coastal current nose and the integrated transport (m$^3$ s$^{-1}$), which is defined as the downstream, along-shore (v) transport at y = - 127.5km integrated in time (Figure 2.12). The bottom
slope promoted a longer coastal current region, and the most pronounced differences were observed between Riv2c–Slope and Riv2c–flat, the former presenting a coastal current that is 50km longer and an integrated transport that is 75,000 m$^3$ s$^{-1}$ larger than the latter (after 60 days). On the other hand, the experiment Mix4c–slope presented a coastal current that is slightly shorter and with a smaller transport than the one in Mix4c–flat, a fact that could be attributed to the large upstream displacement of the plume by the bottom slope.

Figure 2.11 Locations of the offshore edge of the recirculating bulge from selected pairs of flat and sloping bottom experiments (Control, Riv2c and Mix4c). Positions are shown every 5 days for better visualization.
Figure 2.12 Time series of integrated downstream coastal current transport (m$^3$ s$^{-1}$, left) and displacement of the coastal current nose (km, right) away from the estuary mouth from selected pairs of flat and sloping bottom experiments (Control, Riv2c and Mix4c). Coastal current transports were calculated at an across-shore section 127.5km south of the estuary.

2.5.4 Plume development in hybrid coordinate layers

In applying the HYCOM model on the idealized basin that includes “coastal” and “offshore” settings, it is important to consider if the choice of vertical coordinates can impact the plume dynamics, namely the along-shore and across-shore evolution of the plume and its vertical structure. Therefore, we explored the hybrid layer capability of the model and reproduced all slope experiments with two additional vertical coordinate configurations. In the first case, we substituted the standard cartesian-isopycnal configuration with purely sigma coordinates. 30 sigma levels were imposed with thicknesses ranging from 0.66m in the estuary and near the coastline to 3.33 m in deep water. In the second case, a sigma-isopycnal configuration was imposed. 24 sigma levels were prescribed in the first 48 m of depth (thicknesses ranging from 0.83m in the estuary...
and near the coastline to 2 m in deep water) which laid on top of 6 deep isopycnal levels. In both configurations, the sigma levels were set to remain fixed, ie. they could not transform to isopycnal layers.

Salinity horizontal (near surface) distributions and vertical cross-shore sections for a selected experiment (Riv2c-slope) employing its standard vertical coordinate configuration (cartesian–isopycnal) and using the two new cases (sigma-only and sigma-isopycnal) are presented in Figure 2.13. In all cases we found that the plume vertical and horizontal structures are not impacted by the hybrid vertical coordinate choices. In the cases with 2 types of layers (cartesian-isopycnal and sigma-isopycnal), the upper ocean region where the fixed levels were imposed was always deeper than the plume region (buoyant plume and bottom undercurrent), which was a necessary measure to ensure proper vertical resolution of the plume structure. The choice of the depth that defines the region of permanent fixed levels was critical. Results from experiments where isopycnal layers could reach 10m below the surface or less (not shown) had isopycnals interacting with the bottom of the buoyant plume, which was detrimental for the vertical structure of the plume. The flexibility that vertical coordinates have to transform to fixed (cartesian / sigma) or isopycnal layers is a powerful tool to provide the best vertical resolution for different ocean processes. However, for the purposes of this study (which involves a freshwater source, hence a process that continually changes density at the surface), it was equally important to maintain the surface layers as fixed levels at all times (i.e., permanent cartesian / sigma levels that cannot transform to isopycnal layers) in order to ensure adequate resolution.
Figure 2.13 Right: Across-shore salinity vertical structure along section 1 (see Figure 2.1), starting at the estuary mouth (where the slope starts) from the Riv2c-slope experiment with three different vertical layers setting, at day 60. Upper: cartesian-isopycnal. Middle: sigma only. Lower: sigma – isopycnal. Layer interfaces are shown as solid white lines. Left: Corresponding Sea Surface Salinity field, for each case. The plume boundary (34.9) is represented by a white line.

Hybrid coordinate issues have been previously addressed by Winther and Evensen (2006), who tested three different vertical level configurations (involving cartesian and sigma coordinates) on numerical simulations of the shelf-shelf break circulation and water masses formation in the North Sea and Skagerrak region. They concluded that model results from each configuration did not differ considerably from each other; they employed comparison with in situ and satellite data to evaluate model errors associated with the model setup and properties of the vertical mixing scheme. When employing a
hybrid structure and a nested approach with HYCOM to study coastal processes beyond the purely buoyancy-driven problem addressed in this study, vertical resolution is an important issue, as pointed out by Halliwell et al. (2009). Large scale models in HYCOM (global and basin-wide) used to extract boundary conditions employ a vertical coordinate strategy in the stratified open ocean that limits the thickness of the near-surface fixed coordinate domain and maximizes the ocean region represented by isopycnal coordinates. This strategy usually provides poor vertical resolution over the middle / outer continental shelf so that the bottom boundary layer cannot be resolved in the outer, larger scale domain and is detrimental for nested coastal models. It is, therefore, advisable to expand the near surface fixed coordinate domain in the outer model fields by adding additional layers before nesting to the coastal domain.

2.6 Summary and concluding remarks

Previous numerical modeling and laboratory experiments have shown that the development of the recirculating bulge and properties of the coastal current are sensitive to different conditions at the source of freshwater, such as the momentum, buoyancy and overall outflow transport (Yankovsky and Chapman, 1997; Garvine, 1999; Fong and Geyer, 2002), the angle of the buoyant outflow with the coast line (Garvine, 2001; Avicola and Huq, 2003a, b) and the actual river boundary conditions in numerical models (Yankovsky, 2000; Garvine, 2001). In this chapter, numerical experiments were designed to investigate how the variability in the structure of a river plume is connected to changes in the vertical mixing of riverine waters inside an estuary-like source. Although the
estuarine dynamics were not elaborated in detail and in spite of the model configuration, it was demonstrated that the dynamics of the flow prior to reaching the receiving basin play an important role on the properties of the buoyant plume in a flat bottom basin. The results show that increased vertical and horizontal mixing of freshwater inside the estuary enhanced the estuarine gravitational circulation and led to stronger and less buoyant outflows that developed a consistent anticyclonic veering at the river mouth. This shift in the outflow properties clearly impacted the near (bulge) and far (coastal current) fields of the plume, since it led to the development of river plumes that varied between having smaller bulges with a coherent upstream intrusion, or presenting large and circular bulges with no upstream intrusion or not even developing a coherent bulge, as all outflow was deflected in the downstream direction.

The impact of the earth’s rotation on estuarine/bay dynamics has been modeled in different large scale systems (Valle-Levinson et al., 1996, 2007; Kourafalou, 2001; Soares et al., 2007a,b). Chao and Boicourt (1986) and Chao (1988a) demonstrated that under the effect of the earth’s rotation, the upward entrainment caused by the estuarine gravitational circulation will induce an anticyclonic shear on the surface estuarine circulation, a cyclonic shear on the bottom inflow and the development of an S-shaped secondary circulation. Moreover, the plume outflow should be in approximate geostrophic balance (for low Ekman and Rossby numbers) with the development of a Margules density front at the estuary mouth. This pattern is observed in the simulations and resembles the vertical structure of a large scale estuary such as the Delaware Bay (Münchow and Garvine, 1993b; Sanders and Garvine, 2001). The results expand on
previous findings by showing that the strength and direction of this geostrophic outflow are dependent on the degree of freshwater vertical mixing inside the estuary. As this mixing was enlarged, there was an increase in the average upward vertical velocity inside the estuary. This shift, in accordance to Chao (1988a), was concomitant with an intensification of the geostrophic outflow and an enhancement of the anticyclonic veering at the estuary mouth. Changes in the outflow angle (the angle the outflow makes with the coastline) have been related to the geometry and orientation of the estuary / bay previously (Garvine, 2001; Avicola and Huq, 2003a,b). This angle is also dependent on the estuarine dynamics.

The development of the river plume in the presence of a sloping bottom is in agreement with previous studies (Chao, 1988a; Kourafalou et al., 1996a; Garvine, 1999) as well as with observations on topographic effects on plume development (Valle-Levinson et al., 2007). The results highlight that the impact of changing the estuarine mixing conditions was greatly minimized when the buoyant plumes developed in the presence of a gentle sloping bottom. Although the plumes were not in contact with the bottom (surface-advected plumes), their development was affected by the bottom slope. This was especially evident in the case of plumes that were very distinct in the flat bottom domain experiments, but developed very similar features in the presence of a sloping bottom. The largest impacts were in the shift of the recirculating bulge in the upstream direction and in the subsequent change in the configuration of the outflow. Very interestingly, these impacts appear to transmit changes into the estuary, since the buoyant outflow $T_f$ decreased together with a slight decay in the average estuarine vertical
velocity in comparison to the same experiments in the flat bottom basin (Figures 2.7 and 2.8).

Although the approach is idealized (box-like domain with simple estuary, flat or gently sloping bottom and no external forcing), the results demonstrated that a two-way interaction may exist between the buoyant plume and the estuarine circulation and that both should be considered as part of a single system (MacCready et al., 2009). It is important to emphasize that this study is in the context of large-scale estuaries ($K_i > 1$) where the effects of rotation are part of the estuarine dynamics. These concepts may not be applicable to narrow estuaries ($K_i < 1$), where constrictions (lateral jetties, sills) that act as hydraulic controls in the estuarine channel may have a profound impact on the plume outflow (Hetland, 2005; MacDonald and Geyer, 2004; MacDonald et al., 2007; McCabe et al., 2008). The assumption of no temperature difference between river inflow and shelf is generally valid, as river plume dynamics are controlled by the salinity gradients. However, strong coastal temperature gradients imposed either by a cold discharge or by increased cooling of the shallow portion of the shelf due to cold air outbreaks can have implications for density-driven coastal currents. An example is the West Adriatic Coastal Current which is largely driven by river runoff, with the exception of the winter season, when it becomes barotropic (wind-driven), due to compensation of the temperature and salinity gradients in the density field (Zavatarelli et al., 2002).

The choice of vertical coordinate (fully cartesian, fully sigma, cartesian-isopycnal, sigma-isopycnal) did not induce major changes in the vertical structure of the plume,
although having a number of near surface layers that are permanently cartesian/sigma
levels was necessary to ensure the vertical resolution of the plume. Isopycnal layers
cannot change their assigned density, and therefore they can be detrimental to the vertical
resolution of the buoyant plume in case it interacts with them. Based on the results, it is
recommended that isopycnal layers should remain well below the bottom of the buoyant
plume. More generally, the choice for this “minimum isopycnal depth” should be
carefully determined by the user and is undoubtedly dependent on the process under
study, the forcing mechanisms and the general framework of the simulations (process-
oriented or realistic). The choice of hybrid layers is a beneficial model feature for
realistic simulations with variable bathymetry, where the inner-shelf is connected to
deeper shelf areas and to the open ocean.
Chapter 3

The Impact of Bottom Topography and Winds on the Dynamics of the Mississippi River Plume

3.1 Overview

The first step to address the dynamics of the Mississippi River (MR) plume and the impact of a multiple forcing environment of the plume dynamics is to conduct process oriented simulations that examine the model response to individual forcing mechanisms. This approach was adopted with success by other authors such as Kourafalou et al. (1996a) and Soares et al., (2007a), and it is employed in this study. In this chapter, a high resolution regional model of the Northern Gulf of Mexico (NGoM) region is developed and numerical simulations are performed to isolate the impact of realistic bottom topography and wind forcing over the dynamics and structure of the MR plume. The emphasis is on the shelf conditions that enhance offshore removal beyond the shelfbreak, where plume waters can interact with the mesoscale eddy circulation.

3.2 The NGoM-HYCOM model

The NGoM-HYCOM model is an application of HYCOM as a high resolution domain (Mercator mesh of 1/50 °, about 1.8 km) that covers part of the Northern Gulf of Mexico.
Mexico. It was developed for this study and includes the majority of the Northern Gulf coastline, extending from the Big Bend region in the Florida Panhandle to part of the northern Texas coastline (Figure 3.1). Bottom topography is derived from the 2-min NAVO/NRL DBDB2 global data set with true coastline, minimum depth of 2 m and maximum depth of about 3000 m in the interior of the DeSoto Canyon. 30 hybrid vertical levels are employed. In the upper 40 m of the water column, we impose 15 permanent fixed levels (i.e., levels that cannot revert to isopycnal layers) which transit from z-levels that are spaced 0.25 m apart close to the coastline, to terrain-following (sigma) levels over the inner continental shelf and back to z-levels in mid-shelf waters. The transition of the sigma levels back to z-levels occurs when the thickness of each vertical level reaches a predetermined maximum value that ranges from approximately 1 to 5 m (increasing downwards). This feature gives the model layers the flexibility to have a variable distribution throughout the domain. Model layers quickly transform from sigma to z-levels in regions of steep topography (east of the MR Delta) and remain as sigma levels for a longer extent over regions where the continental shelf is wider (west of the MR Delta). Below 40 m of depth, vertical levels are initialized as isopycnal layers (thicknesses increasing from approximately 20 to 500 m towards the bottom) and may transform to z-levels, if necessary. This particular distribution of vertical levels is chosen such that adequate vertical resolution in the upper water column always exists to resolve the vertical structure of the MR plume. The results from the previous chapter suggest that when simulating river plumes, permanent fixed levels (either sigma or z) should be imposed in the upper water column. The reason is to prevent isopycnal layers from interacting with the bottom of the buoyant plume, which could be detrimental for the
vertical structure of the plume. Further illustration of this structure of vertical coordinates is shown in Figure 2.13, Chapter 2. The current choice to have half of the vertical levels as isopycnal layers is based on the fact that this study also involves the simulation of mesoscale eddies and deep ocean currents like the ones generated by Loop Current features, and isopycnal layers provide optimal resolution for such oceanic processes.

![Figure 3.1 NGoM-HYCOM domain. Selected isobaths are shown and the shelfbreak (100 m isobath) is highlighted in blue. The location of the Mississippi River is shown (Mis, red dot).](image)

Lateral open boundaries are on the southern boundary and part of the western boundary of the domain. At the coastal wall, the normal velocity is zero and a no-slip condition is used for the tangential velocity. There is no flow normal to the topography. At the bottom, momentum is dissipated by a quadratic bottom drag (drag coefficient $C_d = 3 \times 10^{-3}$), using a bottom velocity $u_b$ that represents the average velocity in a slice of water situated just above the bottom. The thickness of this slice is determined by the bottom boundary layer parameterization from the KPP scheme (Halliwell et al., 2009). Salt and heat fluxes normal to the bottom and to the coast are set to zero.
3.3 Process-oriented simulations

The following process-oriented experiments were designed: firstly, the buoyancy-driven circulation of the MR plume in the presence of realistic topography is accessed by conducting an experiment where the only forcing mechanism is the discharge from the MR Delta. Basin waters are initially in a state of rest, and an ambient stratification is imposed based on the Generalized Digital Environmental Model 3 (GDEM3) climatology for the Gulf of Mexico region (Teague et al., 1990). Temperature and salinity vary only in the vertical (leveled density layers) and profiles represent an annual average of the monthly GDEM3 climatology from a point at the center of the GoM basin. An average MR discharge of 13,500 m$^3\cdot$s$^{-1}$ is imposed and the model plume is left to evolve for a period of 30 days. At the offshore lateral open boundaries, the baroclinic structure of the flow is relaxed to the basin initial state, whereas the barotropic structure is solved through the method of characteristics (Browning and Kreiss 1982, 1986). The development of the buoyant plume for 30 days allows studying buoyancy processes away from the freshwater source and evaluating the influence of topographic controls over the plume development. In addition, it enables the investigation of the impact of wind-driven circulation over the developed, large buoyancy signal. In order to evaluate how the MR plume would evolve when it is free of bottom topography constraints, a twin experiment is performed where the realistic bottom topography is replaced by a 20 m flat bottom. All initial conditions remain the same, except that 30 cartesian levels are prescribed in the vertical (spaced every 0.66 m). These two experiments will be referred as “topography experiments” hereafter.
The final day of simulation from the buoyancy-driven experiment with realistic topography will serve as initial condition for the experiments where the impact of the wind-driven circulation is evaluated ("wind experiments" hereafter). A set of experiments are initially designed where the MR plume is forced by 4 basic wind patterns: westerly (upwelling-favorable), easterly (downwelling-favorable), northerly and southerly winds. For each wind pattern, a steady 0.075 Pa wind stress forcing is imposed for a 3 days duration period.

3.4 Results

3.4.1 Impact of bottom topography on plume development (topography experiments)

The development of the buoyancy-driven MR plume in the presence of a 20 m deep flat bottom is presented in Figure 3.2 (right panels). The development of the broad MR plume is characteristic of large-scale river plumes that are affected by the earth’s rotation and are free to evolve in the absence of bottom slope constraints (Chao, 1988a Kourafalou et al., 1996a Garvine, 1999; Schiller and Kourafalou, 2010). The expansion of the plume is characterized by an anticyclonic turning of the buoyant flow at the point of discharge, with the development of a coastal current signal that propagates in the direction of Kelvin wave propagation (westward in the NGoM region, "downstream" direction). By day 5 (not shown), an anticyclonic circulation (bulge) develops around the head of the Delta and by day 15 a young coastal current transports plume waters towards
the Louisiana-Texas shelf at approximately 0.2 m s\(^{-1}\). By day 30, large meanders have developed around the Delta, which enhanced the offshore expansion of the plume. The coastal current hugs the coast and presents a weak meandering character, suggestive of barotropic/baroclinic instabilities (Oey and Mellor, 1993; Kourafalou et al., 1996a. A vertical across-shore section at the coastal current region (not shown) indicates that the plume is in geostrophic balance, with a salinity front that reaches 15 m of depth but does not intersect the bottom (surface-advected plume, as defined by Yankovsky and Chapman, 1997).

Figure 3.2 Snapshots of surface salinity and surface velocity vectors after 15 and 30 days of buoyancy-forcing only in the presence of realistic bottom topography (left) and in the presence of a 20 m deep flat bottom (right). The isobaths of 10 m, 20 m, 50 m and 100 m (shelfbreak) are shown as solid gray lines. Vectors are shown every other 8 grid points for better visualization.
The development of the MR plume in the presence of the realistic bottom topography deviates from the pattern above (Figure 3.2, left panels), and the changes are in agreement with previous findings on the impact of a bottom slope over the development of a river plume (Kourafalou et al., 1996a; Garvine, 1999; Schiller and Kourafalou, 2010). The across-shore sloping bottom tends to elongate the plume in the along-shore direction and reduce the offshore expansion of the plume due to potential vorticity constrains. Although in the first 5 days the development of the plume is similar to the case in flat bottom conditions (not shown), the development changes as soon as the plume feels the influence of the bottom topography. By day 15, the plume has elongated in the along-shore direction and a weak upstream intrusion transports a smaller fraction of plume waters towards shallow depths (< 5 m) to the north of the Delta. The bulge circulation has also changed, and by day 30 a large bulge has developed which presents a meandering pattern. Although the offshore intrusion was reduced, a fraction of plume waters expands towards the offshore region and beyond the shelfbreak (100 m isobath). The downstream coastal current now extends beyond the Atchafalaya River, and the core of the current is displaced from the coast and approximately follows the 10 m isobath. The lack of meanders indicates that coastal current instabilities are inhibited by the bottom topography. The vertical structure of the downstream coastal current (not shown) indicates that the plume is in geostrophic balance, and it exhibits a salinity front that extends from the surface to the bottom (maximum depth of 12 m). The weaker upstream coastal current (within a coastal area near the coast north of the MR Delta) represents a balance between the along-shore pressure gradient (due to the low salinity waters near the Delta) and the along-shore acceleration; the along-shore Coriolis term is near zero within
the coastal band (no cross-shore transport due to the coastal constrain, see Csanady, 1982).

3.4.2 Plume response to idealized wind forcing (wind experiments)

The idealized wind forcing had a major impact on the surface distribution of the buoyant plume waters. The response of the buoyant plume was dependent on the surface circulation induced by each wind direction, with distinct circulation patterns close to the coast and in deeper waters. Two wind scenarios enhanced the offshore transport of plume waters beyond the shelfbreak, westerly (W) and southerly (S) winds (Figure 3.3). The W winds are upwelling-favorable everywhere, while the S winds are upwelling-favorable on the northeast part of the MR Delta. After 36 hours of steady W winds (Figure 3.3, left panels), the plume structure indicates that buoyant waters are being transported offshore over the shelf and eastward close to the coast line. Because the main orientation of the coast line is in the east-west direction, the response of shelf waters to W winds consists of an offshore Ekman transport, a downwind drift at the coast and a strong coastal jet, the latter driven by a geostrophic current which is induced by a coastal sea level set-down (Csanady, 1978; Chao, 1988b Kourafalou et al., 1996a). The downwind coastal jet reverses the buoyant coastal current, and the coastal morphology of the MR Delta works as an impediment for the eastward coastal current, which is locally diverted offshore and augments the offshore removal of plume waters. To the east of the Delta, plume waters are also transported eastward and towards the shelfbreak. After 72 hours of W winds, a large fraction of plume waters that constituted the bulge (located in the vicinity of the
Delta) were transported beyond the shelfbreak and wind-induced mixing has weakened the plume low-salinity signal (Figure 3.3, left panels).

![Figure 3.3](image)

**Figure 3.3** Snapshots of surface salinity and surface velocity vectors after 36 and 72 hours of westerly (left) and southerly (right) winds. The shelfbreak (100m isobath) is represented as a solid gray line. Vectors are shown every other 8 grid points for better visualization. The solid black lines show locations of vertical across-sections in Figure 3.4.

The response of shelf waters to S winds consists of an eastward integrated Ekman transport, a small weakening of the buoyancy-driven westward coastal current, and a strengthening of the upstream coastal current (north of the Delta). The buoyant plume presents distinct responses in the coastal current and in the bulge region due to the variations in the coast line and in the bottom topography (Figure 3.3, right panels). After 36 hours of S winds, the coastal current is disrupted as the surface downwind drift pushes the plume against the coast and the local coastal morphology induces a horizontal
divergence of plume waters at 91ºW. This process also induces retention of low salinity waters on the west side of the Delta. Although the low-salinity bulge is pushed against the Delta, the plume waters are also advected eastward along and across the shelfbreak which enhances offshore removal towards the deep DeSoto Canyon. After 72 h of S winds, the plume low-salinity signal is maintained within a narrow coastal current region, as opposed to begin deformed during W winds. This is due to the accumulation of plume waters against the coast west of the Delta during S winds.

Both W and S winds induce vertical mixing of plume waters. In the vicinity of the Delta and after 72 hours of W winds, plume waters are transported offshore as a vertically well-mixed layer that is approximately 6 m thick (Figure 3.4, upper panel). The vertical mixing of the plume is attributed to shear-induced mixing generated by the vertically sheared along-shore and across-shore currents, and this process was described in detail by Fong and Geyer (2001). The depth of the plume corresponds to an equilibrium thickness $h_c$, at which the plume bulk Richardson number $R_i_b$ approaches a critical value and turbulent mixing is induced. $R_i_b = \frac{g'h^3}{4\left(\tau^w/\rho f\right)^2}$, where $g'$ is the reduced gravity, $h$ is the plume thickness, $\tau^w$ is the wind stress, $\rho$ is the plume density and $f$ is the Coriolis parameter. This equation can be rewritten as $h_c = \left[\frac{4R_i_c\left(\tau^w/\rho f\right)^2}{g'}\right]^{1/3}$, and $h = h_c$, when $R_i_b = R_i_c$, a critical Richardson number at which shear-induced mixing is triggered. In our case and for $R_i_c$ equal to 0.5 (Fong and Geyer, 2001), $h_c$ is equal to 4.8 m which
closely corresponds to the plume thickness. The plume is effective at restricting the surface Ekman dynamics to within the plume layer. Figure 3.5 presents vertical profiles of the major along-shore and across-shore momentum balance terms at one location outside the plume (A) and another location inside the plume (B) after 72 hours of W winds (the locations of the profiles are shown in Figure 3.4). The along-shore (x dir.) and across-shore (y dir.) momentum equations in Cartesian coordinates are, respectively:

\[
\frac{Du}{dt} + f u = -\frac{1}{\rho_o} \frac{\partial p}{\partial x} + \frac{\partial}{\partial z} \left( K \frac{\partial u}{\partial z} \right)
\]

\[
\frac{Dv}{dt} + f u = -\frac{1}{\rho_o} \frac{\partial p}{\partial y} + \frac{\partial}{\partial z} \left( K \frac{\partial v}{\partial z} \right)
\]

Where \( D(\quad)/dt = \partial(\quad)/\partial t + u \partial(\quad)/\partial x + v \partial(\quad)/\partial y + w \partial(\quad)/\partial z \). Terms 1 and 5 are the total rate of change, terms 2 and 6 are the Coriolis acceleration, terms 3 and 7 are the pressure gradient force and terms 4 and 8 are the vertical viscous term. \( u \) is the along-shore velocity, \( v \) is across-shore velocity, \( w \) is the vertical velocity, \( f \) is the Coriolis parameter, \( \rho_o \) is the reference density, \( p \) is the pressure and \( K \) is the vertical turbulent viscosity coefficient. The horizontal viscous terms are two order of magnitude smaller than the vertical viscous terms, and they are neglected in this calculation. The momentum balance profiles outside the plume (A) show the dominance of a balance between the Coriolis and vertical friction (i.e., Ekman balance) and an Ekman layer that is approximately 25 m deep. On the other hand, inside the plume (B), the momentum balance suggests a balance between the pressure gradient, Coriolis and vertical viscous terms. This balance is confined to the upper 5-7 m (plume layer) and represents the preexisting plume geostrophic balance with an Ekman balance superimposed.
**Figure 3.4** Vertical across-shore sections of salinity after 72 hours of westerly and southerly winds. The locations of the sections are shown in Figure 3.3. Dashed white lines show the locations of vertical profiles A, B, C and D where major momentum balance terms are calculated.

**Figure 3.5** Vertical profiles of momentum balance terms computed at points A, B, C and D (shown in Figure 3.4) after 72 hours of westerly (W) and southerly (S) winds. Upper panels show across-shore (y dir.) terms and lower panels show along-shore (x dir.) terms.
In the case of S winds (Figure 3.4, lower panel), the plume vertical structure in the vicinity of the Delta also shows the development of a well mixed plume layer. However, the thickness of this layer is more variable than in the previous case because plume waters are advected onshore by the surface downwind currents. This onshore transport is different than the onshore Ekman transport caused by downwelling-favorable winds and is restricted to the near-surface (Tilburg, 2003). Figure 3.5 presents vertical profiles of the major momentum balance terms at one location outside the plume (C) and at another location inside the plume (D) after 72 hours of S winds (the locations of the profiles are shown in Figure 3.4). The momentum balance is again an Ekman balance outside the plume (C) with an Ekman layer that is approximately 25 m thick. Inside the plume (D), the stronger stratification confines the Ekman dynamics to the plume layer and the momentum is balanced by the pressure gradient, Coriolis and vertical viscous terms.

The above scenarios demonstrate that W and S winds develop surface Ekman dynamics that promote the offshore transport of MR plume waters beyond the shelfbreak. Shelf waters respond to similar dynamics in the presence of Easterly (E) and Northerly (N) winds, although plume waters are confined to the shelf region. Downwelling-favorable E winds (Figure 3.6, left panels) promote an onshore Ekman transport, a downwind drift at the coast and a strong coastal jet that is driven by a geostrophic current, the latter induced by a coastal sea level set-up (Chao, 1988b Kourafalou et al., 1996a). After 72 hours, the buoyant plume is confined against the coast and is elongated in the downstream direction. The offshore extent of the plume is greatly reduced in the vicinity of the Delta since bulge waters are transported onshore and eastward. Despite the
fact that N winds are directed offshore in the NGoM region, they are inefficient in promoting the transport of plume waters beyond the shelfbreak. Southward currents coming from the north augment the anticyclonic plume circulation at the head of the Delta, and they push the plume westward and towards the shelf (Figure 3.6, right panels). This westward transport is also enhanced by the development of a westward surface Ekman transport in the vicinity of the Delta. In the coastal current region, the coastal morphology now induces a convergence at 91ºW that pushes the buoyant plume in the offshore direction. This process is augmented by a southward transport of plume waters by the near-surface currents (Tilburg, 2003), but the shelf to the west of the MR Delta is too broad and plume waters cannot reach the shelfbreak after 72 hours of N winds.

![Figure 3.6 Snapshots of surface salinity and surface velocity vectors after 72 hours of easterly (left) and northerly (right) winds. The shelfbreak (100m isobath) is represented as a solid gray line. Vectors are shown every other 8 grid points for better visualization.](image-url)
3.5 Discussion and concluding remarks

The process-oriented experiments demonstrated the basic response of the broad Mississippi River plume to the bottom topography and idealized winds. Although these experiments were simplified in some aspects, they provided useful information about the buoyancy-driven circulation of a large river plume in the NGoM region, how wind-driven dynamics impact the structure of the plume and what conditions favor the offshore transport of plume waters. The bottom topography proved to be an important factor determining the development of the buoyancy-driven plume. In the vicinity of the MR Delta, the steep topography suppressed the offshore expansion of the plume and promoted retention of freshwater close to the source. In spite of that topographic control, a fraction of plume waters could expand beyond the shelfbreak, which is a favorable condition for the development of interactions between the plume and the offshore mesoscale circulation. Conversely, to the west of the Delta, the broadening of the shelf maintains the plume away from the shelfbreak, and it is unlikely that plume waters will interact with the offshore eddy circulation unless other forcing mechanisms push the plume towards the shelfbreak, such as winds.

The response of the MR plume to idealized wind forcing was in agreement with previous studies on the effect of wind-driven dynamics over river plumes (Chao, 1988b; Kourafalou et al., 1996a; Fong and Geyer, 2001). Away from the coastline, over the shelf and beyond the shelfbreak, surface Ekman dynamics dictate the horizontal spreading of plume waters. Westerly (upwelling-favorable) winds promoted the largest spreading and
offshore transport of plume waters. Again, the steep bottom topography in the vicinity of the freshwater source favored the advection of plume waters towards beyond the shelfbreak. Also, the coastal morphology suggests that the MR Delta acts as an impediment to the eastward, downwind coastal current, which has to diverge offshore; this could add an extra component to the offshore spreading (Figure 3.3). This is a unique impact of a Deltaic coastal morphology that is not observed in estuary/bay systems, such as the development of the Delaware river plume in the presence of wind forcing (Whitney and Garvine, 2005), and highlights the necessity to improve the understanding of Deltaic river plumes. Southerly winds also promote offshore transport of plume waters even though they are directed onshore. The steep topography close to the Delta plays a major role as it allows the development of an eastward Ekman transport, which diverts riverine waters along the shelfbreak and towards deeper regions of the DeSoto Canyon. This eastward Ekman transport was previously suggested by Morey et al. (2003a), and it is confirmed in this study with the computation of major momentum balance terms within the plume region (Figure 3.5).

The wind direction in the NGOM region is not strictly zonal or meridional (Velasco and Winant, 1996; Wang et al., 1998), and previous studies have focused on the impact of southeasterly (SE), southwesterly (SW), northwesterly (NW) and northeasterly (NE) winds on the dispersion of the MR plume (Walker et al., 2005a Wang and Justič, 2009). Wind experiments were also performed where the idealized MR plume was subject to those wind directions, with a configuration similar to the one from the wind experiments in Section 3.3. The results (not shown) demonstrate that SW and NW winds also promote
an eastward and offshore transport of plume waters, whereas SE and NE winds advect the plume westward and towards the shelf. A brief comparison between all eight wind scenarios shows that after 72 hours of wind forcing, W, NW and SW winds generated the largest plume areas beyond the shelfbreak (Table 3.1). Therefore, one could expect that the likelihood of plume waters to interact with the offshore mesoscale circulation is larger in the presence of winds with a westerly component.

Table 3.1 Plume offshore area (beyond the shelfbreak) after 3 days of steady wind forcing from 8 different directions. Wind stress is constant at 0.075 Pa. Percentage represents an increase or decrease in the offshore area with respect to the initial offshore area at the beginning of the wind experiments.

<table>
<thead>
<tr>
<th>Wind direction</th>
<th>Offshore area (km$^2$)</th>
<th>% with respect to the initial offshore area</th>
</tr>
</thead>
<tbody>
<tr>
<td>W</td>
<td>13.75</td>
<td>+ 83.92</td>
</tr>
<tr>
<td>S</td>
<td>8.64</td>
<td>+ 15.57</td>
</tr>
<tr>
<td>E</td>
<td>4.05</td>
<td>− 45.74</td>
</tr>
<tr>
<td>N</td>
<td>6.73</td>
<td>− 09.99</td>
</tr>
<tr>
<td>SE</td>
<td>5.44</td>
<td>− 27.14</td>
</tr>
<tr>
<td>NE</td>
<td>4.43</td>
<td>− 40.62</td>
</tr>
<tr>
<td>SW</td>
<td>12.70</td>
<td>+ 69.88</td>
</tr>
<tr>
<td>NW</td>
<td>10.60</td>
<td>+ 41.80</td>
</tr>
</tbody>
</table>
Chapter 4

Interactions between the Mississippi River Plume and Offshore Boundary Currents

4.1 Overview

In the previous chapter, the buoyancy-driven circulation of the Mississippi River (MR) plume was characterized, as well as the impact of bottom topography and winds on the evolution and dispersion of plume waters. In this chapter, the NGoM-HYCOM model is applied in a realistically-forced simulation to address the impact of the offshore circulation on the dynamics and transport of the MR plume. The interactions between the plume and offshore boundary currents are investigated, and the role of winds and bottom topography on those interactions is also addressed. This chapter focuses on the characterization of offshore removal of plume waters in a scenario forced by multiple shelf and offshore processes, and on the identification of environmental conditions that favor the offshore exportation of MR waters.

4.2 Realistically-forced simulation

The realistically-forced experiment covers a period of 2 years, from January 1st, 2004 to December 31st, 2005. High frequency (3-hourly) atmospheric forcing is derived
from a regional coupled ocean-atmosphere simulation performed with the Coupled Ocean/Atmospheric Mesoscale Prediction System (COAMPS) (Hodur et al., 2002) with a horizontal resolution of 27 km. The surface boundary is forced by fields of wind stress, air temperature, radiation exchanges (incident solar radiation + radiation emitted by the surface) and precipitation. Surface latent heat (evaporation) and sensible heat fluxes are calculated using the parameterization scheme of Kara et al. (2005b).

Freshwater sources (Mississippi River plus other sources, Figure 4.1a) are prescribed following the freshwater parameterization described in Chapter 2 (mass inflow parameterization and spreading). Point-source discharges were included for all Northern Gulf rivers, except for the Mississippi River, where the discharge was prescribed at the locations of the Southwest Pass, South Pass and Pass a Loutre. In order to reduce the low salinity spike that may be created when all river discharge is concentrated in a few grid cells, we imposed a downward penetration of 4 m and a minimum spreading to surrounding grid points at all freshwater sources. The reproduction of smaller-scale, near-field outflow dynamics described by Wright and Coleman (1971) is beyond the scope of this study. We focus on the far-field properties of the MR plume. Daily-averaged freshwater discharges are prescribed for all sources marked in Figure 4.1a (except for Mobile Bay and Pearl River where monthly climatologies are imposed). A twin experiment without river forcing also took place to allow the calculation of freshwater fraction (Section 4.3.1).
In order to apply open boundary conditions that realistically introduce the circulation induced by the LC, LCE and other mesoscale eddies, the NGoM-HYCOM model is nested within the regional GoM-HYCOM model which has horizontal resolution of 1/25 ° and 20 hybrid layers in the vertical (Fig. 4.1b) (Prasad and Hogan,
The GoM-HYCOM simulation used herein employs the Navy Coupled Data Assimilation system (NCODA) (Cummings, 2005), which is an oceanographic version of the multi-variate optimum interpolation technique commonly employed in atmospheric forecasting systems. The NCODA system assimilates satellite altimetry track-by-track and sea surface temperature (SST) directly from orbital data using model forecasts as the first guess. Although the system assimilates more data types, the limited availability of in situ observations in the open GoM makes the assimilation scheme to rely primarily on satellite altimetry and SST measurements. Kourafalou et al. (2009) showed the realistic representation of the LC variability in the NCODA based GoM-HYCOM simulation, as compared to a free running, twin simulation. Nesting to the NGoM-HYCOM domain is done in off-line mode (archived files). A dynamical boundary condition is applied for which no distinction is made between the inflow and outflow boundaries (Halliwell et al., 2009). The method of characteristics is used to solve the barotropic flow fields (velocity and pressure) (Browning and Kreiss, 1982, 1986) and a nesting relaxation zone (10-grid points wide with e-folding time of 0.1 to 24 days) is used to relax the baroclinic structure (temperature, salinity, pressure and velocity) towards the fields provided by the GoM-HYCOM model.

4.3 Results

Although the Mississippi River is the primary source of low-salinity waters in the Northern Gulf, the contribution of several other freshwater sources results in a low-
salinity band along the northern coast. The dynamics of the broad MR plume and the low-salinity coastal band in the vicinity of the MR Delta are highly affected by the wind conditions and the bottom and coastal topography. Figure 4.2 presents examples of different states of the MR plume and low-salinity band under distinct wind conditions. During a period of easterly winds (Fig. 4.2a), plume waters are transported to the west of the MR Delta, following the westward down-wind coastal current. Because the main orientation of the coast line is in the east-west direction, the dynamics correspond to the response of a buoyant plume to downwelling-favorable winds (Csanady, 1978; Chao, 1988b; Kourafalou et al., 1996a). Easterly winds promote an onshore Ekman transport and a downwind drift at the coast, while the low-salinity band is confined against the coast. The vertical structure of the plume in the coastal current region (Fig. 4.2d) demonstrates that the downwelling-favorable winds can efficiently mix the plume in the vertical. The offshore extent of the plume is reduced in the vicinity of the Delta, since plume waters are transported onshore and westward. The transport of low-salinity waters towards the broader Louisiana-Texas shelf effectively insulates the coastal plume from the offshore circulation beyond the shelfbreak (100 m isobath). As a consequence, interactions of the MR plume with shelfbreak and offshore eddies is unlikely to happen during easterly wind events.
Figure 4.2 Sea surface salinity from selected days depicting different conditions of the MR plume (part of the model domain shown). (a): during a period of easterly, downwelling-favorable winds, (b): after a period of westerly, upwelling-favorable winds, (c): during a period of southerly winds. The gray lines represent the 100 and 1000 m isobaths. The blue vector shows the average wind direction during each event at a point in front of the MR Delta (white triangle). The black line shows the location of the vertical sections of salinity presented in (d) (day 87, easterly winds) and (e) (day 100, westerly winds). Salinity values less than 26 are not shown. The vertical layers in the upper water column at the location of the section are illustrated in (e) as grey solid lines.
The offshore transport of plume waters is enhanced in the presence of westerly, upwelling-favorable winds (Fig. 4.2b). To the west of the Delta, the eastward downwind drift at the coast reverses the buoyant coastal current, and low-salinity waters are transported offshore due to surface Ekman dynamics (Fig. 4.2e) (Chao, 1988b; Fong and Geyer, 2001). The result is a broadening of the low-salinity coastal band. The proximity of the shelfbreak to the MR Delta maximizes the offshore transport of plume waters in that region. Low-salinity riverine waters are transported eastward and offshore, towards the rim of the DeSoto Canyon, a process that facilitates the interactions between the plume and the offshore circulation. Offshore transport of low-salinity waters is also induced in the presence of southerly-southwesterly winds (Fig. 4.2c). Although the onshore winds cause retention of MR plume waters on the west side of the Delta, low-salinity waters are also transported eastward and along the shelfbreak. This process is attributed to the surface Ekman transport that develops in the vicinity of the Delta (Morey et al., 2003a). The response of the plume is different on the Louisiana-Texas shelf. Due to the variations in the local coastal topography, the onshore winds promote a divergence of waters within the low-salinity band between the Atchafalaya and the MR Deltas. This disrupts the coastal current and erases the plume signal along the Louisiana-Texas shelf.

4.3.1 Shelf transport of plume waters around the MR Delta

The pathways and the overall fate of MR plume waters are first investigated by computing the freshwater transport $Q_{fw}$ across four sections that establish a closed region around the MR Delta (m1, m2, z1 and s1, Figure 4.3). $Q_{fw}$ was computed as
\[ Q_{fw} = \int_{h}^{\eta} f w V dz dx, \]

where \( V \) is the horizontal velocity normal to the section, \( \eta \) is the sea level, \( h \) is the bottom depth and the integral with respect to \( x \) is the horizontal distance along the section. The freshwater fraction \( f_{fw} \) is equal to \( f w = \frac{S_b - S}{S_b} \), where \( S_b \) is the background undiluted salinity and \( S \) is the diluted salinity due to the river discharge. \( S_b \) was obtained from a twin experiment that was forced by the same atmospheric and lateral boundary fields but without river inflows. \( Q_{fw} \) was further separated into barotropic and baroclinic components, and the total \( Q_{fw} \) is given by

\[ Q_{f\text{, total}} = Q_{f\text{, barotropic}} + Q_{f\text{, baroclinic}} = \int_{h}^{\eta} f w V_{b\text{, trop}} dz dx + \int_{h}^{\eta} f w V_{b\text{, clin}} dz dx, \]

where \( V_{b\text{, trop}} \) and \( V_{b\text{, clin}} \) are the barotropic and baroclinic components of the across-section velocity, respectively.

Figure 4.3 Location of the sections (m1, m2, z1, s1 and iso1000m) where freshwater transport analysis is performed. The area \( A \) represents the region where wind stresses are spatially averaged. Selected isobaths are shown (solid gray lines). The dots on section iso1000m show the initial (green dot) and final (red dot) points of the along-section distance in Figs. 4.7 and 4.8.
First, we estimated the relative importance of the barotropic and baroclinic currents for the flow structure and \( Q_{fw} \) at sections m1, m2, z1 and s1. This estimate was based on the ratio of the mean baroclinic Eddy Kinetic Energy (EKE) to the total EKE (see Teague et al. (2006) for details). \( R_{eke} \) is defined as

\[
R_{eke}(z) = \frac{\overline{u_{dd}^2(z)} + \overline{v_{dd}^2(z)}}{\overline{u^2(z)} + \overline{v^2(z)}}
\]

where \( \overline{u^2(z)} \) and \( \overline{v^2(z)} \) are the total velocity variances and \( \overline{u_{dd}^2(z)} \) and \( \overline{v_{dd}^2(z)} \) are the variances of the baroclinic component of the velocity. Daily outputs of model velocities from the two years of simulation were used in this analysis. \( R_{eke} \) provides the fraction of the total velocity variance that is explained by baroclinic currents. If \( R_{eke} = 0 \), currents are mostly barotropic. If \( R_{eke} \geq 1 \), currents are mostly baroclinic.

Figure 4.4 shows the vertical structure of \( R_{eke} \) along the four sections. At section m1 (Fig. 4.4a), the \( R_{eke} \) has low values (< 0.3) in the interior of the water column, indicating that barotropic processes account for more than 70 % of the EKE in mid-depth layers. Values slightly higher (~0.5) are observed at the surface, which are associated with buoyancy effects induced by the river discharge. Close to the bottom, \( R_{eke} \) was equal to or larger than 1, suggesting strong velocity shears associated with a bottom Ekman layer and cross-shelf flows. The structure of \( R_{eke} \) at sections m2 and z1 (Fig. 4.4b and 4.4c) is similar to the one from section m1, although a larger portion of the water column has values lower than 0.3. In contrast to the other sections, section s1 (Fig. 4.4d) presents a more defined surface layer (10-15m deep) where \( R_{eke} \) was close to 0.6 and baroclinic processes contributed more to the EKE. In spite of the large layer in the interior of the
water column where EKE is mostly barotropic, a large bottom layer with $R_{eke} > 1$ exists and that represents the effect of strong velocity shears and bottom intrusions from the offshore region.

**Figure 4.4** Vertical structure of the kinetic energy ratio $R_{eke}$ at sections m1, m2, z1 and s1 (marked in Fig. 4.3). The magenta line represents the 0.3 contour. Values above 1 are not shown.

The $R_{eke}$ analysis suggests that $Q_{fw}$ is largely barotropic across sections m1, m2 and z1, whereas $Q_{fw}$ could present a stronger baroclinic signal across section s1. Figure 4.5
presents estimates of barotropic $Q_{fw}$ and baroclinic $Q_{fw}$ across sections m1, m2, z1 and s1 for the two years of simulation, together with time series of wind stress components that were spatially averaged over area $A$ (Fig. 4.3). Wind events that favor the transport of plume waters towards the offshore region, and therefore the interaction with the offshore eddy circulation, are identified as W1-6. Those wind events constitute periods of winds with a strong westerly component or periods of southerly-southwesterly winds. In agreement with the $R_{eke}$ analysis, $Q_{fw}$ across m1 (Fig. 4.5b) and across m2 (Fig. 4.5c) is mostly barotropic. The barotropic $Q_{fw}$ across both sections is well correlated with the across-section wind stress (east-west); the linear correlation coefficient $r$ is equal to 0.68 at m1 and equal to 0.45 at m2. The larger correlation at m1 is attributed to the proximity of m1 to the coast line, where the flow is constrained by the coastal topography. At m2, the open shelf introduces a larger variability in the direction of the flow.

The largest wind-driven, eastward transports of plume waters across m2 and towards the DeSoto Canyon were observed during W2 and W5, which are periods in the spring and spring-summer transition when the winds developed a southerly-southwesterly direction. W2 and W5 were prolonged periods that lasted for approximately 2 months each. With the exception of a few reversals, the freshwater transport was mostly eastward with the total $Q_{fw}$ (barotropic + baroclinic) peaking at 1.6 (W2) and 2.8 (W5) \times 10^4 \text{ m}^3\text{s}^{-1}. Shorter wind events with easterly or southerly components are also observed in the spring (W1) and the summer (W6), and they also contribute to eastward transports that were short in duration with peaks ranging from 0.5 to 1.2 \times 10^4 \text{ m}^3\text{s}^{-1}. In the autumn (W3) and
winter (W4), eastward transport of plume waters is observed in the presence of brief periods of strong westerly winds and the passage of cold fronts.

**Figure 4.5** (a): Time series of wind stress components computed as spatial averages from area $A$. Selected wind periods for further analysis are delimited by vertical dashed gray lines and are identified as W1-6. (b), (c), (d) and (e): Time series of barotropic and baroclinic freshwater transport $Q_{fw}$ across sections m1, m2, z1 and s1 (marked in Fig. 4.3). Positive/negative values represent eastward/westward transport across sections m1 and m2, northward/southward transport across section z1 and onshore/offshore transport across section s1.
The freshwater transport across z1 is basically barotropic (Fig. 4.5d) and is well correlated with the north-south wind stress (linear correlation of 0.63). The lower magnitude of the barotropic $Q_{fw}$ demonstrates that a small fraction of MR waters is transported northward. The largest northward transport across z1 also corresponds to an event of southerly winds at the beginning of period W2 ($Q_{fw}$ total = $1.3 \times 10^4$ m$^3$s$^{-1}$). Southward pathways mostly correspond to the transport of freshwater from sources in the Mississippi-Alabama coast and towards the MR Delta. Finally, $Q_{fw}$ across s1 has different characteristics than across the other sections (Fig. 4.5e), since the magnitude of the baroclinic transport is much larger than the magnitude of the barotropic transport. That corroborates with the $R_{ske}$ structure from section s1, which presents a surface layer where the baroclinic currents are more than or as equally energetic as the barotropic currents. The baroclinic $Q_{fw}$ across s1 presents larger variability and magnitude than across the other shelf sections, and the correlation with the wind is less clear. That is attributed to the fact that s1 is located at the shelfbreak and is very close to the freshwater source; fast changes in the wind direction could disperse large volumes of plume waters on and off the shelf. In the absence of winds, plume waters could also expand offshore. Low-salinity waters are advected away from the Delta and towards the shelfbreak (negative baroclinic $Q_{fw}$) more frequently during the winter and spring months, when the offshore transport can reach $-2 \times 10^4$ m$^3$s$^{-1}$. In the end of the spring and during summer months, the offshore transport becomes less significant since the plume waters are pushed against the MR Delta by the southerly winds. Plume waters are then transported to the east and towards the rim of the DeSoto Canyon, where the buoyant plume can interact with the offshore circulation and be removed offshore. The above results elucidate the processes
controlling the interaction between wind and buoyancy-driven flows on the NGoM shelf. They are not intended as a study of seasonal variability, which requires a longer simulation.

### 4.3.2 Offshore transport of MR plume waters

*a) Loop Current (LC) extension during 2004-2005*

The offshore removal of the MR plume into the interior of the GoM is directly related to the proximity of the LC system and mesoscale eddies to the shelfbreak, where they may entrain riverine waters. The interactions between the MR plume and the LC system should take place when the LC is well extended and close to the shelfbreak of the NGoM region, or when an LCE is shed and it intrudes into the NGoM. In order to identify the periods of time when MR plume dynamics may be affected by the LC system, we tracked the LC northernmost position in time using a technique presented by Leben (2005). Daily sea surface height (ssh) fields from the GoM-HYCOM simulation that provided boundary conditions to the NGoM-HYCOM model were used to determine the boundary of the LC and its maximum latitude. The boundary of the LC is defined as the 17-cm ssh contour that extends from the Yucatan Channel to the Straits of Florida, and it is exemplified as a black solid line in Figure 4.1b. The 17-cm ssh contour is selected as a proxy for the high velocity core of the LC, and this parameter was previously used to track the LC evolution and shedding of LCEs with success by Leben.
(2005). On each day, we identify the maximum latitude of the 17-cm ssh contour to determine the northward penetration of the LC.

The time series of the LC maximum latitude (Figure 4.6) shows the progressive evolution of the northward penetration of the LC. In spite of short term variations, the northward penetration increases with time until it reaches a peak and drops abruptly when an LCE is shed. Events of LCE separation and reattachment are also observed as short, large negative oscillations in the maximum latitude time series. Three periods of LC impact and interaction with the MR plume are separated by vertical solid red lines and identified as LC1-3. During year 2004, the LC presented a very clear cycle of continuous northward penetration, shedding of an LCE in mid August and return to the port-to-port configuration (Yucatan to Straits of Florida). At the time of the eddy separation, the LC was well extended northward and the LCE intruded in the southern boundary of the NGoM-HYCOM model, therefore impacting the offshore circulation of the NGoM region (LC1 event). The LC does not present a clear cycle during year 2005, when a series of detachment and reattachment events took place. The LC system was well penetrated northward for several months, especially between the months of April and June when the LC was beyond 28 °N and impacted the NGoM offshore circulation (LC2 event). That period is followed by a series of eddy detachment-reattachment events with the LC also intruding in the NGoM region (LC3 event). The variability observed in 2004-2005 is representative of the LC inter-annual variability (Leben, 2005).
Figure 4.6 Time series of the maximum latitude of the Loop Current (LC) calculated from the full GoM-HYCOM regional model. The maximum latitude is based on the northernmost point of the LC ssh17-cm contour exemplified in Fig. 4.1b. Three periods of LC impact on the MR dynamics that are further investigated are shown (LC1,2 and 3) and are delimited by vertical red lines. The horizontal gray line represents the latitude of the southern open boundary of the nested NGoM-HYCOM model.

b) Impact of basin-wide circulation on the MR plume

The offshore transport of MR waters and the impact of the LC and other eddies are evaluated across a section located above the 1000 m isobath, which is just offshore of the shelfbreak in the vicinity of the MR Delta (section iso1000m, Fig. 4.3). Figures 4.7 and 4.8 present hovmöller diagrams of ssh anomaly, across-section surface velocity and total $Q_{fw}$ across the section iso1000m during years 2004 and 2005, respectively. On each day, the ssh anomaly represents the actual ssh along the section minus its average on that day. The ssh anomaly is useful to identify ssh variations associated with the passage of eddy structures. On year 2004 and during the event LC1, the passage of the tip of the LCE is observed as a strong, positive (anticyclonic) anomaly signal propagating eastward for several days in mid August (Fig. 4.7a). This is immediately followed by a strong, negative (cyclonic) signal, which indicates the passage of a cold core eddy (LCFE).
Throughout the year, several other eddy signals are observed to pass over the region, and major events are labeled as A* and C* for anticyclonic and cyclonic eddies, respectively.

Figure 4.7 Hovmöller diagram of ssh anomaly (a), across-section surface velocity (b) and total freshwater transport $Q_{fw}$ (c) across section iso1000m for year 2004. Positive/negative values for onshore/offshore velocities and transports. The solid black line in (a) represents the ± 0.1 contour. Wind periods of interest are shown as vertical magenta lines in (c). Periods of LC impact are delimited by horizontal dashed black lines. Selected anticyclonic (A*) and cyclonic (C*) eddy events are shown. The orientation of the diagrams is given by the green and red dots and is shown in Fig. 4.3.
At the time of the passage of the eddies, the response of the surface across-section velocity (Fig. 4.7b) is coherent with the eddy circulation and the ssh anomaly distribution along the section. Onshore/offshore velocities are observed on western/eastern side of anticyclonic eddies, and vice-versa for cyclonic eddies. As the eddy features propagate along the section, they induce offshore and onshore surface currents that determine the offshore removal of plume waters. Figure 4.7c shows the clear impact of the LCE and other eddies on the transport of the MR plume. During the LC1 period, streams of offshore (negative) $Q_{fw}$ are observed as the eddy impinges against the shelf and propagates along the section. Other events of MR plume entrainment are also observed as other eddies (A1, A2, C1) travel through the region. The wind periods that promoted eastward transport of plume waters across m2 are also shown in Figure 4.7c (W1-4). With the exception of W4, riverine waters were removed offshore during all wind events, especially during W2 when very large volumes of freshwater were transported offshore in the presence of southerly-southwesterly winds. In spite of winds favoring offshore removal during W4, MR waters were not removed offshore due to the onshore circulation induced by eddies C2 and A3.

Similar patterns of interactions between eddies and the MR plume are also observed during year 2005 (Figure 4.8). Large ssh anomalies are observed in the times when the LC intruded in the NGoM region (LC2-3, Fig. 4.8a). For a period of approximately 2 months (LC2), the LC imposed persistent offshore surface currents across iso1000m (Fig. 4.8b) and entrained freshwater from the shelf (Fig. 4.8c). The entrainment of plume waters decreases substantially at the end of LC2 when a LCE is shed (Fig. 4.6). The
offshore transport only resumes when the LCE intermittently reattaches to the LC (event LC3). After that period, the LC system does not impact the NGoM region and other mesoscale eddies are responsible for the entrainment process (A4-5, C3-4).

**Figure 4.8** Same as Fig. 4.7, but for year 2005.
c) Offshore removal of MR waters in the presence of southerly winds (events W2 and W5)

Southerly-southwesterly winds during the period W2 lasted for approximately 2 months. In that time, large volumes of freshwater were transported eastward across m2 (Fig. 4.5c), followed by offshore removal (Fig. 4.7c). The LC system had not approached the NGoM region yet, and the entrainment of the plume was due to ambient circulation and eddies located in the DeSoto Canyon. Wind period W5 was also characterized by southerly-southwesterly winds, large eastward transport across m2 and subsequent offshore removal, although the LC system approached the NGoM region towards the end of the wind period (Fig. 4.8c).

The spatial structure of the plume and the interactions with the offshore circulation during those two events are exemplified in Figure 4.9. During period W2 (Fig. 4.9a), the offshore circulation was characterized by a meandering westward flow (~ 0.5 m s\(^{-1}\)) and the presence of mesoscale eddies with 50-100 km in diameter. As the winds reverse the plume transport and advect the plume along the shelfbreak (days 154 and 157), the eddy circulation entrains the riverine waters and a coherent low-salinity band is formed along the offshore boundary current (day 160). The entrainment continues even after winds slightly push the plume onshore (day 163), and a few days later the offshore low-salinity band is dispersed by the offshore eddies (not shown). A similar scenario takes place during period W5 (Fig. 4.9b), when the plume is advected towards the head of the DeSoto Canyon and is dispersed by the offshore circulation. Interestingly, the tip of the
LC was located south of the MR Delta (as seen by the anticyclonic circulation on days 443 and 446, ~ 0.8 m s\(^{-1}\)), but it did not take part in the offshore removal of the plume at that time. These results suggest that when the plume is captured by the mesoscale circulation it is unlikely to return to the shelf, even if the wind direction changes.

\[ d) \text{Entrainment of MR waters by the LC system (events LC1 and LC2)} \]

The interactions between the MR plume and the LC system are exemplified in Figure 4.10. Figure 4.10a demonstrates the intrusion of the LCE during the event LC1. The anticyclonic circulation of the LCE comes into close proximity to the MR Delta, with currents reaching ~ 0.6 m s\(^{-1}\) at the shelfbreak region (day 233). As the tip of the eddy propagates eastward, it effectively entrains the MR plume and leads to the formation of a long and narrow low-salinity band that extends from the Delta to the southern boundary of the domain (day 236). The presence of a cyclonic circulation over the shelfbreak enhances the process, and the dipole circulation induces offshore currents that reach 0.8 m s\(^{-1}\). As the entrainment continues on the eastern side of the LCE, a cold core eddy advances on the western side of the LCE, and the onshore circulation shuts off the entrainment near the Delta (days 239 and 242). A few days later (not shown), the tip of the LCE has moved to the east and the low-salinity band disappeared.
Figure 4.9 Snapshots of sea surface salinity and surface velocity vectors from selected days during the wind periods W2 (a, left column) and W5 (b, right column). Part of the model domain shown. Vectors are shown every other 8 grid points for better visualization. The 100 and 1000 m isobaths are displayed as gray lines. A and C show the location of anticyclonic and cyclonic eddies, respectively. For each day, the wind stress vector averaged over the area $A$ is presented on the upper-right corner of the plot. Salinity values less than 29 are not shown.
Figure 4.10 Same as Fig. 4.9, but for LC periods LC1 (a, left column) and LC2 (b, right column).
The wind vectors in Figure 4.10a show that wind conditions were weak and unfavorable for the offshore transport, such that the removal of the plume was exclusively due to the proximity of the LC. The impact of the LC in the beginning of the period LC2 takes place in the presence of favorable winds (period W5, Fig. 4.8c), but the LC continues to entrain riverine waters when winds become irregular and unfavorable for the offshore transport. At that time the LC is again close to the shelfbreak and efficiently entrains the MR plume, with the formation of a thin low-salinity tongue on the eastern side of the LC (Fig. 4.10b, day 515).

4.3.3 Variability of MR plume offshore pathways during 2004-2005

a) Freshwater transport during selected events

In order to better establish a relationship between the wind-driven, eastward transport of plume waters and the entrainment by the offshore circulation (LC and eddies), the negative values of total $Q_{fw}$ from Figures 4.7c and 4.8c were integrated along the section iso1000m, collapsed into a single time series and plotted with the total $Q_{fw}$ across m2 (Figure 4.11a). With the exception of W4, the eastward transport across m2 was followed by an offshore transport across iso1000m during all wind events. This behavior is particularly visible during the periods W2 and W5, when consecutive bursts of eastward transport were followed by peaks in the offshore transport of approximately same magnitude. Moreover, eddy events that overlapped favorable wind periods positively enhanced the offshore removal (W3 and C1, for example). The exception is
period W4, when the three bursts of eastward transport were not followed by offshore removal. During that time, the eddy event C3-A3 negatively affected the transport of plume waters into the deep ocean (Figure 4.12a). The dipole circulation induced by the cyclone and the large anticyclone (~130 km in diameter) generated persistent onshore flows to the east of the MR Delta for approximately 30 days. When the favorable wind period started, plume waters were advected eastward, caught in the anticyclonic circulation and transported towards the head of the DeSoto Canyon, where they were wrapped around the eddy. The eddy circulation has shifted the offshore pathway from the vicinity of the Delta to head of the Canyon.

Figure 4.11 (a): Time series of total freshwater transport $Q_{fw}$ across sections m2 and iso1000m (marked in Fig. 4.3). Positive/negative values for eastward/westward transport across section m2. Only offshore (negative) portion of the transport is shown for the section iso1000m. Selected wind periods are delimited by vertical dashed gray lines. Periods of LC and eddy impacts are represented by horizontal blue lines. (b): Time series of Mississippi River discharge for the period of the simulation.
Figure 4.12 Snapshots of sea surface salinity and surface velocity vectors during two eddy events, C3-A3 (a) and A2-C1 (b). Part of the model domain shown. Vectors are shown every other 8 grid points for better visualization. The gray lines represent the 100 and 1000 m isobaths. A* and C* show the location of anticyclonic and cyclonic eddies, respectively. For each day, the wind stress vector averaged over the area A is presented on the upper-right corner of the plot. Salinity values less than 29 are not shown.

Although eastward, wind-driven transport of freshwater may initiate and promote the entrainment by the offshore circulation, offshore removal of plume waters is also observed during eddy events that are not favored by eastward transport. The best examples are during event LC1 and at the end of LC2, when the large anticyclonic circulation from the LC system dominated the offshore circulation south of the MR Delta and captured the MR plume. Other eddy events of smaller scale also entrained the MR plume during wind conditions that did not favor offshore transport. During the eddy event A2-C1 (Fig. 4.12b), winds were mostly from the southeast and that produced a westward
plume transport towards the Louisiana-Texas shelf. However, the eddy dipole formed by the anticyclone A2 and the cyclone C1 (~ 50 km in diameter) was positioned just off of the MR Delta, and the offshore currents generated by the eddy dipole were able to capture the MR plume even in the presence of southeasterly winds.

Figure 4.11b shows a time series of the MR discharge during the period of simulation, and puts in perspective the offshore transport of plume waters with respect to the variability of the MR discharge. The transport of river waters to the interior of the Gulf takes place throughout different stages of the MR outflow, during both flood and dry seasons. Continuous offshore removal happens during low discharge conditions (events LC3 and beyond), whereas no removal takes place during high discharge conditions (period W4 until the beginning of period W5). Based on the time series across iso1000m (Fig. 4.11a), we computed the total volume of freshwater that was exported to the offshore region for each removal period (wind, LC and eddies) and compared with the freshwater volume discharge by the Mississippi River (Table 4.1). LC/eddy events that are superimposed with wind events by a few days are treated as a single (combined) exportation period. During the simulation period, the volume of freshwater removed to the offshore GoM ranged from approximately 5 km$^3$ to 43 km$^3$, with the largest volumes associated with wind events. Smaller volumes were associated with eddy events and the largest removal that was exclusively due to the offshore circulation happened during event LC1, when ~16 km$^3$ of freshwater were ejected offshore.
Table 4.1 Total freshwater volume exported to the offshore region during each wind (W*), Loop Current (LC*) and eddy (A*,C*) event. Events that were superimposed in time are put together as one single exportation period (W5 and LC2, for example). The freshwater volume discharged by the Mississippi River during each period is also shown, together with the ratio between the volume exported and the volume discharged.

<table>
<thead>
<tr>
<th>Event</th>
<th>Duration (days)</th>
<th>Total volume (km$^3$)</th>
<th>Volume discharged by the MR (km$^3$)</th>
<th>Ratio exported/discharged</th>
</tr>
</thead>
<tbody>
<tr>
<td>A1</td>
<td>16</td>
<td>4.94</td>
<td>24.90</td>
<td>0.19</td>
</tr>
<tr>
<td>W1</td>
<td>21</td>
<td>8.61</td>
<td>28.76</td>
<td>0.29</td>
</tr>
<tr>
<td>A2-C1</td>
<td>18</td>
<td>7.66</td>
<td>30.44</td>
<td>0.25</td>
</tr>
<tr>
<td>W2</td>
<td>61</td>
<td>43.21</td>
<td>106.12</td>
<td>0.40</td>
</tr>
<tr>
<td>LC1</td>
<td>55</td>
<td>16.26</td>
<td>45.31</td>
<td>0.35</td>
</tr>
<tr>
<td>W3 / C2</td>
<td>42</td>
<td>18.49</td>
<td>37.61</td>
<td>0.49</td>
</tr>
<tr>
<td>W4 / C3-A3</td>
<td>42</td>
<td>0.82</td>
<td>74.91</td>
<td>0.01</td>
</tr>
<tr>
<td>W5 / LC2</td>
<td>104</td>
<td>39.63</td>
<td>140.48</td>
<td>0.28</td>
</tr>
<tr>
<td>W6 / LC3</td>
<td>41</td>
<td>18.37</td>
<td>23.28</td>
<td>0.78</td>
</tr>
<tr>
<td>C4</td>
<td>14</td>
<td>5.04</td>
<td>6.55</td>
<td>0.76</td>
</tr>
<tr>
<td>C5</td>
<td>9</td>
<td>6.12</td>
<td>4.33</td>
<td>1.41</td>
</tr>
<tr>
<td>A4</td>
<td>8</td>
<td>3.52</td>
<td>3.33</td>
<td>1.05</td>
</tr>
<tr>
<td>A5</td>
<td>18</td>
<td>13.37</td>
<td>10.07</td>
<td>1.32</td>
</tr>
</tbody>
</table>

The proximity of section iso1000m to the MR Delta suggests that most of the freshwater exported across iso1000m is of MR origin. Table 4.1 shows the ratio of the volume of freshwater exported offshore to the ratio discharged by the Mississippi River. During the period of wind influence only (W2, no LC impact), winds promoted an offshore transport that corresponded to 40 % of the MR discharge, while during the period of LC impact only (LC1, no wind contribution), the LC system captured a fraction corresponding to 35 %. In the mixed forcing periods, the fractions corresponded to 28 % (W5 / LC2) and up to 78% (W6 / LC3). In other eddy cases, ratios are larger than one and that suggests that pre-existent freshwater on the shelfbreak was also exported offshore. Although these estimates are simplified and not intended to be an accurate freshwater budget, they can be employed to elucidate the processes controlling the removal of MR waters away from the Delta. In particular, our results show that offshore pathways are an active “sink” of plume waters and should be considered in conjunction with the shelf pathways of MR waters removal.
b) MR plume pathways observed with chlorophyll a satellite imagery

We performed a qualitative comparison between the major wind (W2 and W5) and LC (LC1 and LC2) events from Figures 4.9 and 4.10 and chlorophyll a satellite images from the Indian Oceansat-1 Ocean Color Monitor (OCM). Satellite radiance measurements were used to track river water using ocean color channels with relatively high spatial resolutions of 360 m, compared with the Sea-viewing Wide Field-of-View Sensor (SeaWiFS) that has 1.1 km pixels. The SeaSpace Terascan™ software, based on standard NASA algorithms for SeaWiFS (Gordon and Wang, 1994), was used to compute chlorophyll a using the 0.49 and 0.56\( \mu \)m channels as input. Although these chlorophyll a estimates have not been validated for accuracy, they are effective for tracking the motion of near-surface, nutrient-rich river waters in the Northern Gulf of Mexico, as previously described by Walker et al. (2005a).

Figure 4.13a shows chlorophyll a images on specific days that represent the offshore removal of plume waters during the selected wind or LC event. Days were chosen based on the cloud coverage and availability of the image. Figure 4.13b shows the corresponding model surface salinity on the same days as the satellite images. The image on July 6th, 2004 (model day 188) corresponds to a time at the end of period W2, when southerly winds favored the eastward transport of plume waters. A large signal of higher chlorophyll a concentration is observed to the east of the Delta and over the shelfbreak. The signal extends into the interior of the DeSoto Canyon, and that suggests offshore removal by the mesoscale circulation. The image is in good qualitative agreement with
the eastward transport and entrainment of plume waters described during event W2 (Fig. 4.9a). The salinity distribution to the east of the Delta on the same day (Fig. 4.13b) shows the end product of the W2 period, when low-salinity waters of MR origin were transported in large volumes towards the DeSoto Canyon, then mixed and stirred by the offshore circulation. Similar agreement is observed on March 13th, 2005 (model day 438), during the wind event W5. The chlorophyll a signal in the surroundings of the Delta suggests an eastward transport of plume waters, in agreement with the model surface salinity on the same day. A pool of low-salinity waters is observed to the east of the MR Delta, which is the initial stage of the eastward transport observed during wind event W5 (Fig. 4.9b).

The satellite images during the LC events demonstrate the remarkable impact of the LC system over the transport of the MR plume. On July 30th, 2004 (model day 212, period LC1), a distinct chlorophyll a band extends from the MR Delta into the interior of the GoM, reaching as south as 24 °N. Hu et al. (2005) reported the same removal event and attributed it to the northward position of the LC, which is in agreement with the LC maximum latitude analysis (Fig. 4.6) and the model results for period LC1. The model surface salinity on the same day (Fig. 4.13b) is in good qualitative agreement and shows the corresponding low-salinity band, which extends from the MR Delta into the offshore region and to the southern boundary of the NGoM-HYCOM model. The satellite image on May 28th, 2005 (model day 514, event LC2) also reveals a distinct chlorophyll a band extending into the GoM, which agrees with the LC conditions during this period (Fig. 4.6). The oligotrophic conditions on the west side of the band suggest that the riverine
waters are being transported along the edge of the Loop Current. The model surface salinity on the same day shows the presence of the low-salinity band just to the east of the Delta, in agreement with the position of the chlorophyll α band across the shelfbreak. The satisfactory comparison reinforces the necessity to employ a modeling approach that uses lateral boundary conditions from a realistic, data-assimilative ocean model, in order to capture strong interactions between the MR plume and the Loop Current.

![Figure 4.13](image)

**Figure 4.13** (a): Oceansat-1 OCM chlorophyll α images from selected days during events W2, W5, LC1 and LC2. The chlorophyll α scale ranges from 0 to 3 mg m\(^{-3}\). The 100 and 1000 m isobaths are shown as black lines. River water is depicted in red/brown tones near the mouth of rivers, where pigment concentrations are highest and orange/yellow tones in deep water, where pigment concentrations are reduced. (b): Snapshots of model sea surface salinity on the same days as the satellite images. The 100 and 1000 m isobaths are displayed as gray lines. Salinity values less than 29 are not shown.
4.4 Discussion

The results presented in the previous section reveal the complex environmental conditions that control the evolution of the MR plume and the transport of the related low-salinity, nutrient-rich waters. The MR Delta is located in a unique environment, where complex bottom topography, variable wind forcing and strong boundary currents actively impact the dynamics and the transport of the buoyant plume. This study presents a novel analysis of the combined effects of shelf and offshore flows in tandem with topographic controls, on the evolution of a large scale river discharge.

Previous observations (Walker, 1996; Walker et al., 2005a) and numerical studies (Wang and Justić, 2009) have demonstrated the impact of different wind conditions over the structure of the MR plume. We studied the wind influence from the perspective of the plume freshwater transport, and our results are in agreement with previous observations and climatology-forced numerical modeling studies (Morey et al., 2003a, Walker et al., 2005a). With respect to the along-shelf transport, the predominant transport pathway is to the west and towards the Louisiana-Texas shelf, due to prevalent easterly wind conditions in the region (Fig. 4.5b). Easterly winds (southeasterly to northeasterly) predominate in the spring, autumn and winter months with an average annual frequency of 64 % (Walker and Hammak, 2000), therefore making the Louisiana-Texas shelf the most frequent MR plume pathway. Large eastward transport is observed when persistent southerly-southwesterly winds develop, such as towards the end of the spring and during summer months (Fig. 4.5c). During that time, the plume circulation can be completely reversed.
and riverine waters are dispersed to the east of the Delta with a small fraction towards the north, and that process significantly impacts the distribution of low-salinity waters in the region. Westerly winds in the winter and in the fall also induce eastward transport, but they are much shorter in duration, because they are associated with the passage of cold fronts and not with a change in the seasonal pattern of the wind field. Furthermore, estimates of across-shelf freshwater transport in the vicinity of the Delta demonstrate that the offshore transport can be as large as the along-shelf transport (Fig. 4.5e). The proximity of the shelfbreak to the freshwater source allows the MR plume to be constantly expanded offshore and transported on and off the shelf, and the large variability in the cross-isobath transport near the Delta suggests that changes in the wind direction are important for the offshore removal close to the source.

If in the vicinity of the Delta the steep bottom topography favors the offshore transport of plume waters, the broadening of the shelf to the west of the Delta maintains the plume away from the shelfbreak and “insulated” from the offshore circulation. To the east of the Delta, the presence of the DeSoto Canyon facilitates the interactions of the plume with the offshore circulation when the plume is transported eastward, specially in the presence of southerly-southwesterly winds. Under those wind conditions, offshore eddies effectively entrain the plume as low-salinity bands that are subsequently strained and dispersed by the mesoscale circulation (Fig. 4.9). Although this process was previously discussed by Morey et al. (2003a,b), our simulations allowed estimates of freshwater transport, which showed a new result: large volumes of freshwater can be ejected offshore through shelf to deep current interactions. Volumes as large as 40km$^3$
were removed by the mesoscale circulation during prolonged periods of winds that transport the plume eastward (Table 4.1). One characteristic of the eddy removal during periods of eastward transport is that after the plume is entrained, it is “locally” mixed and dispersed in the region of the DeSoto Canyon. When the plume is directly entrained by offshore eddies, especially by the LC system, coherent low-salinity bands extend from the Delta to the southern boundary of the NGoM-HYCOM domain, and are erased just after the passage of the eddy circulation. This distinction suggests that the isolated impact of mesoscale eddies maybe more efficient in transporting plume waters further offshore and into the interior of the GoM.

In situ measurements of salinity for the quantitative evaluation of the plume surface salinity fields were not available for the study period. We performed a qualitative comparison between the modeled plume and chlorophyll δ satellite images during selected events of plume transport. The position and extent of the modeled low-salinity plume was in good agreement with the satellite high-chlorophyll plume signal, which indicates the good performance of the model with respect to the timing and extent of the transport events. Chlorophyll δ satellite imagery has been used as a qualitative proxy to river induced salinity fronts in several modeling and observational studies (McClain et al., 1988; Kourafalou et al., 1996b; Tsiaras et al., 2008; Hickey et al., 2010). In particular, the Mississippi River plume has been tracked using satellite measurements of reflectance (Walker et al., 1994; Walker, 1996; Walker et al., 2005a; Green et al., 2006; Hu et al., 2005) and using chlorophyll δ estimates (Walker et al., 2005a,b; Hu et al., 2004).
Observational studies (Walker et al., 1994; Ortner et al., 1995; Gilbert et al., 1996, Hu et al., 2005) have described transport pathways of MR waters along the Loop Current and toward the Florida Current and the Gulf Stream, but lacked information about the physical processes that initiated this unique cross-marginal removal mechanism in the NGoM region. In this study, we elucidated the specific entrainment process at the freshwater source, and showed how the intrusion of the LC system was an effective mechanism for the offshore transport of plume waters (Fig. 4.10). The entrainment by the LC happened regardless of the wind field, as the proximity of the LC to the MR Delta was a sufficient condition to capture the plume. Other eddies located in the vicinity of the shelfbreak with diameters ranging from 50 to 130 km also played important roles as sinks of riverine waters. Slope eddies (Hamilton et al., 2002; Hamilton and Lee, 2005) actively entrained plume waters, and the formation of anticyclone-cyclone pairs generated offshore-directed currents that were capable of entraining plume waters even in the presence of southeasterly winds (Fig. 4.12b). Conversely, cyclone-anticyclone pairs generated onshore currents that blocked wind-induced tendency for offshore removal of plume waters. In such cases, plume waters were advected towards the head of the DeSoto Canyon (Fig. 4.12a). Similar interactions between the MR plume and large anticyclones were observed before (Walker et al., 1996; Muller-Karger, 2000; Walker et al., 2005a). Our results demonstrate that the positioning and proximity of eddies to the shelfbreak strongly determines the characteristics of the offshore freshwater transport and imposes a large variability in the offshore pathways that the plume may take.
Hypoxia conditions on the Louisiana-Texas shelf in the summer time have been attributed to the discharge of the Mississippi/Atchafalaya River System, and a combination of biological and physical factors (Justić et al., 2007). Inner-shelf currents transport riverine waters to the Louisiana-Texas shelf, where the nutrient-rich freshwater enhances the surface biological productivity and consequently the carbon flux to sediments. This process creates a subsurface layer where bacterial decomposition of organic matter and oxygen consumption are high. At the same time, the freshwater and weak wind conditions in the summer enhance the water column vertical stratification, which decreases the flux of oxygen from the atmosphere to the bottom layers. Ultimately, the combination of these processes leads to the formation of subsurface layers where oxygen concentrations are extremely low, with adverse conditions for marine organisms. Preceding three-dimensional numerical modeling studies have focused on the transport and circulation of the MR plume waters to the west of the Delta, with emphasis on the physical aspects that promote the development of hypoxia conditions (Hetland and DiMarco, 2008; Wang and Justić, 2009). In spite of the advancement in the knowledge of the physical processes related to hypoxia, these modeling studies lacked the implementation of lateral boundary conditions that incorporate the effects of the LC dynamics and offshore eddies. Our results demonstrate that mesoscale eddies can play an important role on the cross-marginal transport of the nutrient-rich, low-salinity waters. The along-shore pathways are thus also influenced, since such eddy events reduce the transport of riverine waters to the Louisiana-Texas shelf, with possible implications on the local development of hypoxia. Our results showed that entrainment can happen year-round, and that the time scale of these events range from weeks to months (Table 4.1).
Hamilton and Lee (2005) observed that the dominant eddies in the Northern GoM can have time scales on the order of 100 days, which suggests that the eddies can interact with the MR plume for a long time if they are well-positioned in the vicinity of the Delta. Future studies employing realistically-forced simulations that are integrated for several years should provide a broader and clearer picture of the relative importance of mesoscale eddies to the dynamics and pathways of the MR plume, to the freshwater budget of the NGoM region and to the hypoxia problem on the Louisiana-Texas shelf.

### 4.5 Concluding remarks

The results presented in this chapter demonstrate that both wind-driven and eddy-driven dynamics play major roles in the transport and dispersion of the MR plume, and that the bottom topography in the NGoM region is a determinant factor in the offshore removal of plume waters. Along-shelf freshwater transport was strongly related to wind-driven, shelf currents. The prevailing easterly winds throughout the year transport plume waters towards the Louisiana-Texas shelf, where broadening of the shelf reduces the interactions between the plume and the offshore eddy field. Short term wind reversals in the fall and winter may transport the plume to the east, where the shelf is narrow and interactions between the plume and eddies are facilitated. However, our results suggest that these reversals are generally too short to enable an efficient entrainment by eddies. The offshore removal of the plume is maximized when the winds are predominantly from the south-southwest. During those wind periods, large portions of MR waters are
advected eastward along the rim of the DeSoto Canyon, where the plume can be easily entrained by the mesoscale field.

An important finding is that offshore removal by eddies does not exclusively happen during winds favoring offshore transport. The complex topography of the shelf areas around the MR plume plays a role in the offshore transport of plume waters, which can thus occur even in the presence of opposing winds. The proximity of mesoscale eddies to the head of the Delta is a sufficient condition to entrain the plume. During entrainment events, the size and position of the eddies, as well as the formation of eddy pairs, determine a variety of offshore pathways of the MR plume. The offshore transport by eddies can be as large as the along-shelf transport, and this result demonstrates that the offshore eddy field is an active component of the plume dynamics. It is concluded that the Loop Current System is an energetic pathway for the offshore transport of riverine waters. In addition, the variability of the Loop Current front and the associated eddy field are found to be a major dynamical factor in the connectivity between the Northern Gulf of Mexico and the basin interior. This finding has a variety of environmental implications, associated with the transport of low salinity waters, but also nutrients and pollutants. The latter has been evident in the recent Deepwater Horizon oil spill incident (April 22\textsuperscript{nd}, 2010), where the surface oil slick has been observed around the MR plume, but also removed from the northern Gulf and toward the interior by the surrounding eddy field.
This study highlights that in order to obtain a complete picture of the processes determining the fate of large buoyant outflows in topographically complex marginal seas, it is necessary to downscale larger scale coarser models, and to employ a nesting approach to properly reproduce complex interactions between the coastal and offshore circulation patterns. Nesting to a data assimilative model, as performed herein, is a desirable approach to ensure proper shelf to offshore interactions.
Chapter 5

Loop Current Impact on the Transport of Mississippi River Waters

5.1 Overview

In the previous chapter, the offshore removal of the Mississippi River (MR) plume was examined, with focus on the interactions with the offshore boundary currents and the interplay of winds and bottomtopography. It was demonstrated that the offshore removal is a frequent pathway, with significant variability associated with offshore eddy activity and with wind forcing, at weekly and monthly time scales. In this chapter, a long-term, realistically-forced simulation of the NGoM-HYCOM model is performed to further investigate the interactions between the MR plume and the Loop Current (LC) system. The variability and modes of interaction between the MR plume and the LC system are investigated, and the overall importance of the LC system as a plume offshore pathway is assessed.

5.2 Long-term, realistically-forced simulation

The numerical experiment in this chapter represents a 5-year realistically-forced simulation of the NGoM-HYCOM model, covering the period from January 1st, 2004 to
December 31st, 2008. Atmospheric forcing, river discharge and lateral boundary conditions come from the same data sets described in section 4.2.

5.3 Results

5.3.1 Influence of the offshore NGoM circulation on the MR plume

The offshore pathways of the MR plume are strongly related to the interactions with the NGoM offshore circulation. These interactions may be modulated by the effect of shelf currents that are primarily wind and buoyancy driven. The combination of both shelf and offshore circulation mechanisms establish a variety of offshore pathways for the MR plume (Schiller et al., 2011). Figure 5.1 shows the different examples of such pathways and interactions with a variety of offshore circulation features. The eastward, wind-driven transport of plume waters that is commonly observed during the summer time is a shelf mechanism that favors the interactions with offshore eddies (Figure 5.1a, 29/6/2004). During the summer, the seasonal shift to southerly-southwesterly winds may promote eastward transport of plume waters along the rim of the DeSoto Canyon, where the shelf is narrow and the plume is susceptible to interaction with slope eddies. The plume is captured as low-salinity bands by the eddy circulation, which subsequently strains and disperses the plume filaments in the offshore region.
Figure 5.1 Sea surface salinity and surface velocity vectors from selected days that exemplify different interactions between the MR plume and the NGoM offshore circulation (part of the model domain shown). Vectors are shown every 8 grid points for better visualization. The gray lines represent the 100 and 1000 m isobaths. A and C show the location of anticyclonic and cyclonic eddies, respectively. Salinity values less than 29 are not shown. (a): Eddy-induced entrainment after a period of eastward, wind-driven plume transport. (b): Offshore removal by a Loop Current Eddy (LCE). (c): Offshore removal induced by a cyclonic eddy field. (d): Offshore removal induced by an anticyclone in the presence of southwesterly winds.
Strong eddy-plume interactions are observed when the LC or LCE impinge against the shelfbreak near the MR Delta (Figure 5.1b, 23/8/2004). Well defined, coherent low-salinity bands are observed to extend towards the Gulf interior along the edge of the offshore anticyclonic circulation. Smaller cyclonic eddies may also interact with the larger scale anticyclonic circulation, and form eddy-dipoles that intensify the cross-shelf flows and enhance the offshore removal of plume waters. Strong interactions are also observed when mesoscale eddies not related to the LC system impact the shelfbreak and offshore circulation, and different plume offshore pathways may be observed in the presence of cyclones or anticyclones. Cyclonic eddies may transport plume waters directly southward from the MR Delta (Figure 5.1c, 7/9/2006), whereas anticyclonic eddies induce a southeastward transport, similar to that caused by the LC system (Figure 5.1d, 20/6/2008, when eastward flow due to southwesterly winds enhanced this effect). Here, the focus is on the interactions between the MR plume and the LC system, which are put in perspective with other types of offshore removal mechanisms.

5.3.2 LC dynamics and northward intrusions during 2004-2008

The interactions between the MR plume and the LC system are dependent on the intrusion of the LC or LCEs in the NGoM offshore region. The LC dynamics are characterized by a progressive northward penetration and westward bending, until the LC reaches an unstable configuration and a LCE is shed (Maul, 1977; Vukovich et al., 1979; Hulburt and Thompson, 1980; Huh et al., 1981). This is usually followed by a southward retreat of the LC, which returns to the so called port-to-port configuration (from the
Yucatan channel directly to the Straits of Florida). The LC dynamics and intrusions in the NGoM region during the period of simulation are investigated by tracking the northernmost position of the LC in time, using the technique presented by Leben (2005). The LC boundary is defined as the 17 cm sea surface height (ssh) contour that extends from the Yucatan Channel to the Straits of Florida, and it is exemplified in Figure 4.1b (solid black line). Daily ssh fields from the GoM-HYCOM model that provided boundary conditions for the NGoM-HYCOM model were used to track the maximum latitude of the LC boundary in time. This metric provides an estimate of the LC northernmost position with respect to the GoM-HYCOM model. In order to support that estimate and demonstrate the intrusion of the LC system in the NGoM region, time series of maximum ssh and maximum temperature at 100 m of depth along an offshore zonal section in the NGoM-HYCOM domain (section o2, Figure 5.2) were constructed. LC intrusions in the northern Gulf should be characterized by values of LC maximum latitude close to or above 28°N (the southern boundary of the NGoM-HYCOM model), and peaks in the maximum ssh and maximum temperature at depth, near the NGoM shelfbreak.

Figure 5.2 Location of offshore sections o1 and o2 that are employed in the analysis of results. Selected isobaths are shown. The red triangle marks the location of wind stress time series in Figures 5.4-5.7.
The time series of maximum LC latitude (Figure 5.3a) demonstrates a large variability from year to year. There are years when the LC shows a clear cycle of northward penetration (metric progressively increases with time), shedding of a LCE (large drop in the metric), and return to the port-to-port configuration (year 2004, for instance). On the other hand, there are years when the LC is somewhat extended for a long time, and a series of detachments of a LCE and successive reattachments take place before the LC returns to the port-to-port configuration (for example, year 2006). There is a clear lack of seasonality in the northward evolution of the LC, and the variability observed in the maximum latitude is representative of the LC inter-annual variability previously discussed by Sturges and Leben (2000) and Leben (2005).

**Figure 5.3** (a): Time series of maximum Loop Current (LC) latitude calculated from the full GoM-HYCOM regional model. The maximum latitude is based on the northernmost point of the LC ssh 17 cm contour exemplified in Figure 4.1b. The horizontal gray line represents the latitude of the southern open boundary of the nested NGOM-HYCOM model. LC intrusion events (LC1 to LC6) are represented by shaded areas. (b): Time series of maximum sea surface height (ssh in m) calculated at the offshore section o2. (c) Same as (b), but for maximum temperature at the depth of 100 m.
The events of LC intrusion in the NGoM region during the 5 years period of simulation are highlighted as shaded areas in Figure 5.3 and identified as LC1 to LC6. The LC intrusions in the northern Gulf region are characterized by events when an extended LC approaches the region (maximum latitude larger than 28°N) or by events when a LCE impacts the NGoM region (maximum latitude can be smaller than 28°N, event LC6 for instance). Figures 5.3b and 5.3c complement the identifications of the LC intrusion events. Overall, the intrusions are characterized by a large values in the maximum ssh (characteristic of anticyclonic circulation) and in the maximum temperature at the depth of 100 m (also a proxy for the LC frontal position) along the section o2, which demonstrate the actual intrusion of the LC system in the NGoM-HYCOM model.

Table 5.1 describes all 6 events with respect to their duration, interval between them and their characteristics (period of extended LC, LCE shedding, etc). The duration of the northward intrusion events ranges from 22 to 110 days. Shorter intervals between events happen when the LC system is close to the NGoM region for several months (year 2005, for example), and the longest intervals represent full years when the LC system did not impact the NGoM offshore circulation (years 2006 and 2008). All events are characterized by phases when an extended LC is present and by phases when a LCE is shed and it is the anticyclonic eddy that impacts the northern Gulf. The exact days of each phase were determined by tracking in time the moments when the LC boundary (Figure 4.1b) breaks and sheds an eddy, and also examining if the eddy reattaches to the LC boundary. In order to facilitate the analysis, LCE reattachment/shedding processes
that last for 3 days or less are disregarded. Some LC intrusion events are characterized by only one LCE phase (LC1, LC2 and LC5), whereas other events are marked by multiple LCE phases (LC3, for instance). The characterization of the LC intrusion events is important because they show if MR plume waters are actually entrained by an extended LC or a LCE. The offshore exportation of plume waters during the LC intrusion events is explored next.

**Table 5.1** Summary of Loop Current (LC) intrusion events. For each event, the start date, duration and interval until the next event are shown. A description and duration for each phase of the events is presented.

<table>
<thead>
<tr>
<th>LC intrusion event</th>
<th>Start date</th>
<th>Duration (days)</th>
<th>Interval until next event (days)</th>
<th>Phases of the event</th>
<th>Number of days</th>
</tr>
</thead>
<tbody>
<tr>
<td>LC1</td>
<td>July 26th, 2004</td>
<td>52</td>
<td>146</td>
<td>Extended LC</td>
<td>26</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>LCE shedding</td>
<td>26</td>
</tr>
<tr>
<td>LC2</td>
<td>February 8th, 2005</td>
<td>22</td>
<td>18</td>
<td>Extended LC</td>
<td>17</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>LCE shedding</td>
<td>5</td>
</tr>
<tr>
<td>LC3</td>
<td>March 19th, 2005</td>
<td>100</td>
<td>22</td>
<td>LCE</td>
<td>11</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>LCE attachment, extended LC</td>
<td>45</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>LCE shedding*</td>
<td>9</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>LCE attachment, extended LC</td>
<td>24</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>LCE shedding</td>
<td>11</td>
</tr>
<tr>
<td>LC4</td>
<td>July 18th, 2005</td>
<td>36</td>
<td>571</td>
<td>LCE*</td>
<td>36</td>
</tr>
<tr>
<td>LC5</td>
<td>March 16th, 2007</td>
<td>32</td>
<td>7</td>
<td>Extended LC</td>
<td>20</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>LCE shedding</td>
<td>11</td>
</tr>
<tr>
<td>LC6</td>
<td>April 23rd, 2007</td>
<td>110</td>
<td>-</td>
<td>LCE</td>
<td>29</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>LCE attachment, extended LC</td>
<td>10</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>LCE shedding</td>
<td>68</td>
</tr>
</tbody>
</table>

*Phase is characterized by short periods (less than 3 days) when the LCE reattaches and detaches from the LC.
5.3.3 Freshwater analysis and MR plume entrainment by the LC system

The entrainment of MR waters by the LC system was estimated in terms of the freshwater transport across section o1 (Figure 5.2). Daily model outputs of salinity and horizontal velocity along o1 were interpolated in the vertical to a grid with levels spaced every 2 m and the analysis was restricted to the upper 100 m of the water column. At each vertical level $k$, the freshwater fraction was estimated as:

$$f_w(k) = \frac{S_w(k) - S(k)}{S_b(k)}$$  \hspace{1cm} (1)

where $S$ is the model salinity (diluted due to the river discharge) and $S_b$ is a background, undiluted salinity. $S_b$ was obtained from a twin experiment that was forced by the same atmospheric and lateral boundary fields but without river inflows.

Horizontal velocities ($u$ and $v$) were rotated by 18.7° in order to obtain along-section and across-section components ($u'$ and $v'$, respectively). Only in the offshore (negative $v'$) freshwater transport across o1 is of interest. The offshore freshwater transport was estimated as

$$Q_{fw} = \sum_{n}^{n} \sum_{k}^{k} f_w(n,k) \times v'_{o f f s h o r e}(n,k) \times dz \times dx$$  \hspace{1cm} (2)

where $n$ is the number of grid points in the along-section direction, $dz$ is the distance between vertical levels (2m) and $dx$ is the distance between grid points along the section (approximately 1800m). $v'_{o f f s h o r e}$ is a function of $v'$ and defined as

$$\begin{cases} v' < 0, v'_{o f f s h o r e} = v' \\ v' > 0, v'_{o f f s h o r e} = 0 \end{cases}$$  \hspace{1cm} (3)
Therefore, only consider the one-way transport towards the offshore region is considered.

For each LC intrusion event, the days of *direct* interaction between the MR plume and the LC system were determined. These periods represent days when the LC system is directly entraining the MR plume, and the model sea surface salinity field shows low-salinity bands from the plume region being ejected offshore by the LC system. Table 5.2 presents statistics on the removal of MR plume waters by the LC system, for each LC intrusion event. A basic characteristic that allows us to establish a distinction between the LC intrusion events is the number of days of direct interaction with the MR plume, in comparison with the duration of the events. Two events (LC1 and LC4) are characterized by interactions with the MR plume that lasted for practically the whole duration of the events (49 out of 52 days and 36 out of 36 days). Two events (LC3 and LC6) were very long in time, and plume to LC interactions constituted approximately half of the durations of the events (48 out of 100 and 61 out of 110 days). Finally, two of the events (LC2 and LC5) have none or very little number of days of plume to LC interaction (0 out of 22 and 4 out of 32).

Estimates of the *plume* offshore freshwater transport directly induced by the LC system \( Q_{fw\text{-}plume\text{-}LC} \) (Table 5.2) show that average values range from approximately \(-4,000 \text{ m}^3\text{s}^{-1}\) to \(-6,000 \text{ m}^3\text{s}^{-1}\), and that the maximum \( Q_{fw\text{-}plume\text{-}LC} \) can reach up to \(-13,000 \text{ m}^3\text{s}^{-1}\). The *plume* freshwater water volume entrained by the LC system \( V_{fw\text{-}plume\text{-}LC} \) is very much dependent on the number of days of plume to LC interaction; three events had
$V_{fw\text{-plume-LC}}$ close to 18 km$^3$, while the LC intrusion event with the longest interaction presented $V_{fw\text{-plume-LC}}$ up to 26 km$^3$.

Table 5.2 Statistics of the offshore freshwater transport across section o1 for each LC intrusion event. Days of direct interaction between the MR plume and the LC system represent days when low-salinity bands are observed to extend from the MR plume beyond the shelfbreak, along the edge of the LC system.

<table>
<thead>
<tr>
<th>Duration of LC intrusion event (days)</th>
<th>LC1</th>
<th>LC2</th>
<th>LC3</th>
<th>LC4</th>
<th>LC5</th>
<th>LC6</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>52</td>
<td>22</td>
<td>100</td>
<td>36</td>
<td>32</td>
<td>110</td>
</tr>
<tr>
<td># of days of direct interaction between the MR plume and the LC system</td>
<td>49</td>
<td>0</td>
<td>48</td>
<td>36</td>
<td>4</td>
<td>61</td>
</tr>
<tr>
<td>Average $Q_{fw\text{-plume-LC}}$</td>
<td>-4127.4</td>
<td>0</td>
<td>-5967.3</td>
<td>-5472.7</td>
<td>-4722.9</td>
<td></td>
</tr>
<tr>
<td>Standard deviation $Q_{fw\text{-plume-LC}}$</td>
<td>3501.3</td>
<td>0</td>
<td>3036.8</td>
<td>1937.6</td>
<td>2597.6</td>
<td></td>
</tr>
<tr>
<td>Maximum $Q_{fw\text{-plume-LC}}$</td>
<td>-13056.2</td>
<td>0</td>
<td>-11648.7</td>
<td>-7792.8</td>
<td>-11258.0</td>
<td></td>
</tr>
<tr>
<td>$V_{fw\text{-plume-LC}}$</td>
<td>17.89</td>
<td>18.81</td>
<td>18.56</td>
<td>1.89</td>
<td>26.17</td>
<td></td>
</tr>
<tr>
<td>$V_{fw\text{-total}}$</td>
<td>18.70</td>
<td>6.35</td>
<td>34.26</td>
<td>18.56</td>
<td>10.49</td>
<td>33.46</td>
</tr>
<tr>
<td>$V_{fw\text{-plume-LC}} / V_{fw\text{-total}}$</td>
<td>0.95</td>
<td>0</td>
<td>0.54</td>
<td>1</td>
<td>0.18</td>
<td>0.78</td>
</tr>
</tbody>
</table>

$Q_{fw\text{-plume-LC}}$ Offshore plume freshwater transport directly induced by the LC system (m$^3$ s$^-1$)

$V_{fw\text{-plume-LC}}$ Plume freshwater volume directly entrained by the LC system (km$^3$)

$V_{fw\text{-total}}$ Total freshwater volume exported offshore for the whole duration of the LC intrusion event (km$^3$)

With the exception of LC4, the duration of the LC intrusion events is longer than the number of days of plume to LC interaction. In the days of no direct plume to LC interaction, freshwater may also be removed offshore by other processes. The total freshwater volume exported offshore for the whole duration of each LC intrusion event ($V_{fw\text{-total}}$) is also shown in Table 5.2. With the exception of LC4, $V_{fw\text{-total}}$ is always larger than $V_{fw\text{-plume-LC}}$ during all LC intrusion events. This is an indication that during the LC intrusion event, freshwater is also removed offshore when the LC does not directly interact with the MR plume. For example, freshwater accumulated on the shelf could be
entrained by the LC system or by other eddies. The MR plume could also be removed offshore when the LC is far from the Delta, but near ambient eddies. When the interactions between the LC system and the MR plume were inexistent or negligible (LC2 and LC5), $V_{fw\text{-}total}$ was mostly a result of other processes that removed freshwater from the shelf. The fraction $V_{fw\text{-}plume\text{-}LC} / V_{fw\text{-}total}$ is close to or equal to one when plume to LC interaction dominated the LC intrusion event (LC1 and LC4). That fraction drops when plume to LC interactions represented just a portion of the LC intrusion event (LC3 and LC6).

5.3.4 River discharge and wind conditions during the LC intrusion events

Figure 5.4a shows the daily river discharge for the Mississippi, Atchafalaya and other rivers combined from 2004 to 2008. All rivers present a clear seasonal signal, with highest discharge conditions occurring in the end of winter and during the spring (February to May), while lowest flow conditions occur in the end of summer and during the fall season (August to November). The discharge of the Mississippi River presents significant inter-annual variability, with the peak discharge ranging from approximately 20,000 m$^3$s$^{-1}$ (year 2006) to 40,000 m$^3$s$^{-1}$ (year 2008). The climatology of the MR discharge is also shown for comparison. Variability with respect to climatology is larger during months of high discharge and lower during low-discharge conditions. The discharge of the Atchafalaya River follows the trend of the MR discharge and is smaller in magnitude. The contribution from all other rivers is even smaller and suggests a secondary importance to the offshore exportation of freshwater. The timing of the LC
intrusion events show that the events tend to happen during low or high-to-low discharge conditions. The effect of the inter-annual variability of the MR discharge with respect to the plume to LC interactions is also observed. For example, events LC3 and LC5 took place during periods of high climatology discharge. However, the inter-annual variability is such that discharge conditions were below climatology values during those events. Year 2008 was a year with extremely high discharge peak, but the LC did not interact with the MR plume.

The wind conditions during each of the LC intrusion events are shown in Figures 5.4b–g, and the moments of direct plume to LC interaction are represented as shaded areas. Some patterns can be observed with respect to the plume to LC interactions and the wind direction. Events LC1 and LC4 suggest that winds did not play an important role during the interactions between the plume and the LC system. Winds were weak during event LC4, when the plume interacted with the LC system for the whole duration of the event. Wind changes observed during event LC1 apparently did not impact the plume entrainment by the LC. Events LC2 and LC5 were characterized by absent or negligible interactions (see Table 5.2). That could be related to the prevailing easterly winds (negative east-west wind stress component, especially during LC5), which promote along-shelf plume transport toward the Louisiana-Texas shelf. Finally, events LC3 and LC6 were the longest in duration, and a relationship between the wind direction and the days of plume to LC interaction is better observed. Overall, the plume was not entrained by the LC when winds had a strong easterly component (negative east–west wind stress
component), but the opposite happens when the winds were weak in the east–west direction or presented a westerly component (positive stress).

Figure 5.4 (a): Time series of daily river discharge from the Mississippi River (MR), Atchafalaya River and all other rivers combined for years 2004 – 2008. The monthly climatology of the MR discharge is also shown for comparison. Periods of Loop Current (LC) intrusion events (LC1 to LC6) are shown as gray shaded areas. S-S-F-W stands for Spring-Summer-Fall-Winter seasons. (b) to (g): Time series of wind stress components at a point in front of the MR Delta (red triangle, Figure 5.2) during each of the LC intrusion events. In each plot, the gray shaded areas represent periods of direct interaction between the MR plume and the LC system.
5.3.5 Interactions between the LC and the Northern GoM shelf

The dynamics and environmental conditions during each LC intrusion event allow us to sort and classify the LC intrusion events into three different cases. This classification is based on the duration of the event, actual interaction with the MR plume and wind conditions.

a) Case I: Strong LC intrusion, direct interaction with MR plume without support of winds (events LC1 and LC4)

This case of LC intrusion and interaction with the MR plume is characterized by a strong intrusion of the LC system. The LC system comes into close proximity to the MR Delta, directly entraining plume waters and transporting them to the offshore region, regardless of the wind conditions. Events LC1 and LC4 are examples of this case. Event LC1 was studied by Schiller et al. (2011) (their Figure 10a), who showed how the passage of the LC system entrains the MR plume and concluded that the proximity of the LC system to the MR Delta was a sufficient condition to capture plume waters.

Event LC4 is another example of this case of LC intrusion, and is exemplified in Figure 5.5. Winds during this event were irregular and weak overall (Figure 5.5a), and offshore freshwater transport $Q_{fw}$ was basically characterized by the impact and passage of the LC system close to the MR Delta (Figure 5.5b). At the moment of strongest intrusion (Figure 5.5c), the tip of the LC system induced an anticyclonic flow of up to 2
m s$^{-1}$ in the offshore region, with current speeds reaching 0.5 m s$^{-1}$ about 40 km south of the MR Delta. Figures 5.5e and 5.5f show that the impact was performed by a LCE, and that the anticyclonic circulation near the MR Delta was induced by the large scale, eddy currents.

b) Case II: Strong LC intrusion, interaction with the MR plume modulated by winds (events LC3 and LC6)

In this scenario, there is a strong intrusion of the LC system in the vicinity of the MR Delta, and the interactions with the MR plume are modulated by wind forcing. The pathway for offshore transport is facilitated by the offshore circulation, but eastward plume transport by the wind is essential for the LC entrainment. Event LC6 is an example of this case and is shown in Figure 5.6. Due to the long duration of this event, winds in Figure 5.6a are smoothed with a 15 days, low-pass filter using a Lanczos filter (Duchon, 1979) for better visualization. The LC system impacted the offshore circulation of the NGoM region for a long period of time (110 days), but the interaction between the MR plume and the LC was dependent on periods of southerly and southwesterly winds (Figure 5.6a). During periods of easterly winds, the plume is transported to the west, and that minimizes the offshore freshwater transport $Q_{fw}$ (Figure 5.6b). Although the LC may be close to the shelfbreak, the westward, wind-driven circulation prevails and transports the plume away from the shelfbreak (Figure 5.6c).
Figure 5.5 Example of Case I of Loop Current (LC) intrusion event. (a): Time series of wind stress vectors at a point in front of the MR Delta (red triangle, Figure 5.2). (b): Time series of $Q_{fw}$ across o1 (Figure 5.2). The red and black dashed lines show the days of the snapshots on (c) and (d). Shaded areas represent periods of time when the LC system interacted with the MR plume. (c) and (d): Snapshots of sea surface salinity and surface velocity vectors on selected days (part of the model domain shown). Vectors are plotted every other 8 grid points for better visualization. The gray lines represent the 100 and 1000 m isobaths. Salinity values less than 29 are not shown. LC and LCE show the locations of the Loop Current and Loop Current Eddy, respectively. (e) and (f): Snapshots of sea surface height (ssh in cm) from the GoM-HYCOM model on the same days as (c) and (d). The LC boundary is shown as a solid black line.
Figure 5.6 Same as Figure 5.5, but for an example of Case II of LC intrusion. Wind is smoothed with a 15 days, low-pass filter for better visualization.
When the wind reverses and comes from the west, the westward coastal current and the whole plume circulation may be reversed. The plume offshore transport is restored (Figure 5.6b, negative values), which represents the direct entrainment by the LC system (Figure 5.6d). The plume pathway has changed from an along-shelf to an across-shelf regime, with the MR outflow being exported to the interior of the Gulf. Figures 5.6e and 5.6f show the different states of the LC system during event LC6. The last 68 days of LC5 are characterized by a LCE state (Table 5.1), therefore one could expect that the large volume of MR waters that is entrained in Figure 5.6d will recirculate in the Gulf, and not be transported downstream along the LC. This aspect is further explored in the Discussions (section 5.4).

c) **Case III: Strong LC intrusion far from the MR Delta, negligible interaction with the MR plume (events LC2 and LC5)**

This scenario is characterized by a strong intrusion of the LC system in the NGoM region, but relatively far from the MR Delta and with negligible interaction with the MR plume. Events LC2 and LC5 are examples of this case, and event LC5 is depicted in Figure 5.7. During this period, winds were basically from the east and unfavorable for the offshore removal of plume waters (Figure 5.7a). The plume to LC interaction takes place only in the beginning of the event, when the LC circulation intrudes in the NGoM region (Figure 5.7b). As the tip of the LC drifts away from the MR Delta, the interactions with the MR plume are completely shut off (Figures 5.7c and 5.7d). The offshore transport of freshwater observed during this time happens after the wind relaxes for a short duration.
(Figures 5.7a and 5.7b). That event of offshore removal is due to shelfbreak currents not directly related to the LC system.

Figure 5.7 Same as Figure 5.5, but for an example of Case III of LC intrusion.
5.3.6 Variability of offshore exportation of Northern GoM riverine waters

The offshore transport of plume waters during LC intrusion events is put into perspective with respect to the year-round offshore exportation of freshwater. The long term offshore freshwater transport $Q_{fw}$ across $o1$ is investigated in relationship with the content of freshwater at $o1$. That is estimated by computing the equivalent freshwater depth $fw_{ed}$ at each grid point along the section:

$$f^{w_{ed}} = \sum_{k}^{k} f_{w}(k) \times dz$$

where $fw_{ed}$ has units of meters. If the water column could be locally “unmixed” into two layers of salinity zero and $S_{b}(k)$ (from (1)), the freshwater layer would be $fw_{ed}$ thick. $fw_{ed}$ is integrated along the section in order to estimate the amount of freshwater that is present in $o1$ ($fw_{ed-int}$, unit m$^2$). Figure 5.8a shows the time series of $fw_{ed-int}$ at $o1$ for years 2004 – 2008. The content and availability of freshwater at $o1$ presents both seasonal and inter-annual variability. Overall, there is an increase in the freshwater content in the end of the spring and summer seasons (clear signal in years 2004, 2006 and 2008), which indicates the eastward transport of freshwater associated with the seasonal shift in winds during the spring – summer transition (Morey et al., 2003b; Schiller et al., 2011). Freshwater content can be close to zero in the fall and winter (year 2004), when northeasterly to southeasterly winds prevail and transport the plume toward the Louisiana – Texas shelf. But, overall, there is a minimum portion of freshwater present at $o1$ year-round.
The long term offshore freshwater transport $Q_{fw}$ across o1 (Figure 5.8b) shows a seasonal signal that is in agreement with the larger freshwater content in the end of the spring and in the summer season. The larger freshwater content during that period (Figure 5.8a) increases the likelihood for offshore eddies to entrain and transport freshwater toward the offshore region. $Q_{fw}$ can reach up to $-3 \times 10^4$ m$^3$ s$^{-1}$ during that time of the year (years 2004 and 2008, for instance), but large exportation of freshwater is not restricted to the end of the spring and the summer season. Events of offshore removal of freshwater are observed year-round, and that reflects the episodic entrainment of freshwater by offshore eddies.

The LC intrusion events are shown as shaded areas in Figure 5.8. One can observe that the contribution of the LC entrainment events to the year-round offshore removal of
freshwater does not readily stand out from the other events in the time series. Table 5.3 shows the total freshwater volume exported offshore in each year ($V_{fw\text{-exported\text{-year}}}$), which is compared to the total MR plume freshwater volume entrained by the LC system in each year ($V_{fw\text{-plume\text{-LC\text{-year}}}}$). The latter represents the summation of $V_{fw\text{-plume\text{-LC}}}$ (Table 5.2) for each year. $V_{fw\text{-exported\text{-year}}}$ ranges from ~123 to 210 km$^3$ with an average value of ~168.8 km$^3$ year$^{-1}$, while $V_{fw\text{-plume\text{-LC\text{-year}}}}$ ranges from 0 (years of no LC intrusion) to 37 km$^3$, with an average value of ~16.6 km$^3$ year$^{-1}$. The ratios between $V_{fw\text{-plume\text{-LC\text{-year}}}}$ and $V_{fw\text{-exported\text{-year}}}$ indicate that there can be years when up to 30% of the total freshwater exported offshore is due to the direct interactions between the MR plume and the LC system (as in year 2005). Taking into account the years of no LC intrusion in the NGoM region, a yearly average fraction of the freshwater exported offshore as the result of MR plume to LC interactions is 11.6%.

**Table 5.3** Estimates of the total volume of freshwater exported to the offshore region across O1 for years 2004 – 2008 ($V_{fw\text{-exported\text{-year}}}$). The volume of plume freshwater directly entrained by the LC system during the same years is also shown ($V_{fw\text{-plume\text{-LC\text{-year}}}}$). Volume unit is km$^3$.

<table>
<thead>
<tr>
<th>Year</th>
<th>$V_{fw\text{-exported\text{-year}}} / V_{fw\text{-exported\text{-year}}}$</th>
<th>$V_{fw\text{-plume\text{-LC\text{-year}}}} / V_{fw\text{-exported\text{-year}}}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>2004</td>
<td>131.57</td>
<td>0.13</td>
</tr>
<tr>
<td>2005</td>
<td>123.24</td>
<td>0.30</td>
</tr>
<tr>
<td>2006</td>
<td>210.98</td>
<td>0</td>
</tr>
<tr>
<td>2007</td>
<td>185.17</td>
<td>0</td>
</tr>
<tr>
<td>2008</td>
<td>193.16</td>
<td>0.15</td>
</tr>
</tbody>
</table>
5.4 Discussion

The results presented in the previous section demonstrate considerable variability in the characteristics of the interactions between the MR plume and the LC system. The plume to LC interactions present strong inter-annual variability. There may be years when the total time of interaction lasts for almost 3 months (year 2005), and also years when the interactions are non-existent (years 2006 and 2008). The interactions are also dependent on the characteristics of the intrusion of the LC system in the northern Gulf. If the anticyclonic circulation of the LC system is very close to the MR Delta, the plume will be captured and removed offshore regardless of other environmental factors, such as wind (Case I). If the LC intrusion is strong but the anticyclonic circulation is located farther offshore, winds may play a very important role in modulating the availability of plume waters for LC entrainment (Case II). The interplay of cyclonic eddies (shelfbreak eddies and LCFEs) is also important during plume to LC interactions, and may add an extra degree of variability to that process.

Ortner et al. (1995) proposed that (1) northward position of the LC and (2) high stratification in shelf waters are two mechanisms required for the entrainment of MR waters by the LC system. They also suggest that contributing factors that enhance plume to LC interactions are (1) high river discharge rates and (2) persistent eastward winds. Although the northward position of the LC system is indeed necessary, the results presented herein demonstrate that if the intrusion of the LC system in the northern Gulf happens over the DeSoto Canyon and far from the MR Delta, interactions with the MR
plume are unlikely to happen (Case III). Northward intrusion of the LC system is a necessary condition, but not sufficient, because it depends locally on the proximity to the MR Delta.

High stratification of shelf waters can facilitate the offshore spreading of riverine waters, since they remain as a freshwater cap over the denser, shelf and offshore waters. In this study, all events of plume to LC interaction happened in the spring or summer seasons. The seasonal higher stratification of shelf and surface waters during the summer could, indeed, facilitate the LC entrainment process. Although that aspect was not explored in this study, results suggest that persistent eastward winds are more important than shelf stratification. When the plume to LC interactions were modulated by winds (Case II), eastward, wind-driven transport was essential for the entrainment process. Persistent eastward plume transport is more common in the summer, as a results of the seasonal change in the wind pattern to southerly – southwesterly winds (Morey et al., 2003a,b; Schiller et al., 2011). At the seasonal time scale, that increases the availability of freshwater beyond the shelfbreak (Figure 5.8a) and increases the likelihood of freshwater offshore removal (Figure 5.8b) by eddies and the LC system.

Although none of the LC intrusion events happened during periods of very high river discharge (Figure 5.6), model results suggest that the discharge rate of the Mississippi River plays a secondary role in determining the “strength” of the plume to LC interaction. The strength of the interaction is determined, in first place, by the number of days of direct interaction between the plume and the LC system, the proximity to the MR
Delta and possible wind effects. Higher discharge rates would increase the volume of freshwater exported offshore by the LC and enhance remote downstream effects (larger freshwater volume reaching the Straits of Florida, for example). At the seasonal time scale, the content and availability of freshwater beyond the shelfbreak does not exactly follow the trend of the MR discharge. For example, year 2006 presented the lowest discharges of the period of simulation (Figure 5.6), but presented the largest signal of freshwater content along section o1 (Figure 5.8a). It is suggested that freshwater availability for eddy and LC entrainment is also dependent on other environmental conditions, such as winds. However, high frequency river discharges (as opposed to monthly or seasonal means) are essential for the accurate representation of the physical conditions that determine the transport and fate of the riverine waters on shorter time scales. This has important implications in the understanding of ecosystem dynamics and ocean health issues, which are locally controlled by river inputs.

When the LC system interacts with the MR plume, the estimates of freshwater entrainment by the LC system are not extremely different from each other ($Q_{fw-plume-LC}$, Table 5.2). An overall average rate of LC freshwater entrainment is $\sim 4,150 \text{ m}^3\text{ s}^{-1}$. However, interactions between the plume and the LC are episodic, and the total volume of plume freshwater capture by the LC will depend on the amount of river discharge and the duration of the interactions.

The 5 years average of plume freshwater volume entrained by the LC is 16.6 km$^3$ per year (section 5.3.6), which gives an average, year-round LC entrainment rate of 526.4
m$^3$ s$^{-1}$. Considering an average value of MR discharge to be 13,500 m$^3$ s$^{-1}$ (Hu et al., 2005), it can be estimated that, on average, 3.9% of the MR discharge is exported offshore by the LC system every year. This seems to be a small number, but the large inter-annual variability of the plume to LC interactions must be taken into account. Table 5.4 compares the volume of freshwater discharge by the MR ($V_{MR}$) with the plume freshwater volume directly entrained by the LC system ($V_{fw-plume-LC}$, from Table 5.2), during each LC intrusion event. The ratio $V_{fw-plume-LC} / V_{MR}$ shows that there can be events when $V_{fw-plume-LC}$ corresponds to 40% and 90% of the freshwater volume discharged by the Mississippi River (events LC1 and LC4, respectively). Although longer numerical simulations would be beneficial to confirm these estimates, the analysis presented herein provides new insights on the role of the LC system over the fate of MR waters and the overall MR plume freshwater balance. This methodology can be useful for freshwater balance and budget studies in the NGoM that may need to parameterize the impact of the LC in the transport and fate of MR waters, and associated materials such as nutrients, sediments and pollutants.

**Table 5.4** Freshwater volume discharged by the MR during each LC intrusion event ($V_{MR}$, in km$^3$). $V_{fw-plume-LC}$ is shown for comparison, as well as the ratio $V_{fw-plume-LC} / V_{MR}$.

<table>
<thead>
<tr>
<th>LC intrusions event</th>
<th>$V_{MR}$</th>
<th>$V_{fw-plume-LC}$</th>
<th>$V_{fw-plume-LC} / V_{MR}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>LC1</td>
<td>42.32</td>
<td>17.89</td>
<td>0.42</td>
</tr>
<tr>
<td>LC2</td>
<td>45.46</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>LC3</td>
<td>120.42</td>
<td>18.81</td>
<td>0.15</td>
</tr>
<tr>
<td>LC4</td>
<td>20.05</td>
<td>18.56</td>
<td>0.92</td>
</tr>
<tr>
<td>LC5</td>
<td>47.50</td>
<td>1.89</td>
<td>0.03</td>
</tr>
<tr>
<td>LC6</td>
<td>135.14</td>
<td>26.17</td>
<td>0.19</td>
</tr>
</tbody>
</table>

The entrainment of plume freshwater by the LC system was also put into perspective with the overall offshore freshwater transport near the MR Delta, and it was
estimated that plume to LC interactions corresponded to approximately 11.6 % of the yearly freshwater offshore transport. Once in the offshore region, the distribution and transport of riverine freshwater is subject to the chaotic behavior of the interacting eddies and the LC system. As discussed by Schiller et al. (2011), the LC system is more efficient at transporting freshwater farther into the open Gulf than cyclones and anticyclones with smaller spatial scales. Although eddies with diameter ranging from 50 to 150 km may promote large removal of freshwater (for example, Figures 5.1c and 5.1d), the significance of the LC system is that it acts as a connectivity pathway to remote ecosystems. Low salinity waters of MR origin may be transported for long distances along the edge of the LC, and eventually reach the Florida Current and the Straits of Florida (Ortner et al., 1995; Gilbert et al., 1996; Hu et al., 2005).

While an extended LC “opens” the pathway to the South Florida region, the shedding of a LCE may close that path. One can expect that plume waters that are entrained by a LCE are most likely to circulate around the LCE and not reach the LC. However, if the LCE is still close to the LC after being shed, MR waters may find their way to the LC via the interaction of the LCE with other mesoscale eddies. This scenario is exemplified in Figure 5.9. Figure 5.9a shows a snapshot of the interactions between the MR plume and a LCE during the event LC6. A strong intrusion of the LCE just offshore of the MR Delta captures the buoyant plume, which is transported along the anticyclonic circulation and southward. This southward transport is enhanced by a cyclonic eddy to the east, which forms a dipole circulation with the LCE. The ssh snapshot from the GoM-HYCOM model on the same day (Figure 5.9b) shows that the LC is to the south of the
LCE, but connectivity between the MR plume and the LC could still be made via the LCE – cyclone dipole circulation.

**Figure 5.9** (a): Snapshots of sea surface salinity and surface velocity vectors on June 11th, 2007, during event LC6 (part of the model domain shown). Vectors are plotted every other 8 grid points for better visualization. The gray lines represent the 100 and 1000 m isobaths. Salinity values less than 29 are not shown. (b): Snapshot of sea surface height (ssh in cm) from the GoM-HYCOM model on the same days as (a). The LC boundary is shown as a solid black line. LC, LCE and C show the locations of the Loop Current, Loop Current Eddy and a cyclonic eddy, respectively. (c): MODIS Aqua chlorophyll a image on June 9th, 2007.
The connectivity during the LC6 event is confirmed with a chlorophyll \textit{a} (chl-a) satellite image from the Moderate Resolution Imaging Spetroradiometer (MODIS – Aqua satellite) on June 9\textsuperscript{th}, 2007 (Figure 5.9c). The image was constructed from 1 km resolution NASA data (http://oceancolor.gsfc.nasa.gov/cgi/browse.pl). River water is represented in red/brown tone near the mouth of rivers and next to the coast line, where pigment concentrations are highest. River waters are depicted in green/light blue tone in deep water, where pigment concentrations are reduced. The MODIS image shows the existence of a chl-a band that extends from the MR Delta toward to the Straits of Florida, with a shape that corresponds remarkably well with the location of the LCE – cyclone dipole and the position of the LC. The southward circulation induced by the eddy dipole transports the chl-a band (MR waters) first southward, directly into the Gulf. When the chl-a band encounters the northern portion on the LC, it is carried eastward and then toward the Straits of Florida by the LC circulation. Understanding the phase of the LC system when it interacts with the MR plume, as well as the eddy configuration in the Gulf, is important to evaluate the availability of MR waters for remote ecosystems.

The position and extent of the modeled plume around the LCE in Figure 5.9a was in very good qualitative agreement with the shape of the chl-a band near the MR Delta (Figure 5.9c), which shows a good performance of the model with respect to the timing and extent of the plume to LC interaction. Previous model validation with respect to plume offshore transport and interactions with the LC was presented by Schiller et al. (2011). Chl-a is an effective proxy to track the motion of nutrient-rich, river waters in the
northern Gulf, and has been employed to track the MR plume before (Hu et al., 2004; Yuan et al., 2004; Walker et al., 2005a,b).

5.5 Concluding remarks

In this chapter, the interactions between the MR plume and the LC system were investigated in a long-term, realistically-forced numerical simulation. The temporal and spatial characteristics of the interactions were studied in the context of across-shelf freshwater transport, LC dynamics and other environmental conditions (river discharge, winds). The plume to LC interactions present significant temporal and spatial variability. Specific events may last for months, with intervals between events that can go up to a year. During years of active LC intrusions, up to 30% of the freshwater exported offshore is due to plume to LC interactions. Wind is an important agent for the LC entrainment when the LC intrusion in the NGoM region lasts for months. If the LC circulation approaches the MR Delta very closely, it becomes a sufficient condition for entrainment and dominates over other environmental factors.

The results presented herein expand the previous understanding of the entrainment of MR waters by the LC system. They confirm that this phenomenon is not anomalous, but it is an important component of the MR plume dynamics with a large degree of variability. It was established that three different combinations of the circulation forcing mechanisms that determine the cross-marginal and cross-basin transport of waters of MR origin. The connectivity pathways that the LC (through its large scale variability)
establishes for MR waters to reach the Gulf interior and downstream remote regions depends on both local (NGoM shelf) and regional (GoM) processes. After the low-salinity bands have reached the Gulf basin, LC and mesoscale eddy dynamics will dictate the downstream transport of plume waters. Studies to address this second component are needed in order to obtain a complete picture of all processes that affect the transport of MR waters toward the South Florida Region. In a broader context, the findings presented herein contribute to the understanding of the role of large-scale current systems as pathways for riverine waters, and stimulate the comparison with other basins where offshore boundary currents are distinctly different from the Loop Current.
Chapter 6

Summary and Conclusions

River plumes are an important component of the coastal ocean environment. The buoyancy inflow from rivers has a broad range of impacts on the continental shelf environment, with implications to the circulation, biological productivity, dispersion of anthropogenic contaminants and to the society needs in general. Understanding the mechanisms controlling the dynamics and dispersion of coastal buoyant plumes is extremely important for coastal management and water quality purposes in the coastal region, especially in complex environments where multiple shelf and offshore processes act simultaneously to change the plume dynamics. This is a challenge in the study of river plumes. The present work sought to improve the understanding of how a large river plume develops in a complex environment, where the impact of offshore boundary currents, wind-driven dynamics and complex coastal and bottom topography are active players in the plume dynamics. The dynamics of the Mississippi River plume were the focus of this study. Specific objectives were set to understand the interactions between the plume and the energetic offshore circulation of the Gulf of Mexico. The approach here was to employ an ocean general circulation model (HYbrid Coordinate Ocean Model – HYCOM) in a series of model experiments that began with idealized conditions and progressively increased in complexity towards full-blown, realistically-forced simulations of the Mississippi River plume. Idealized river plume experiments targeted to
understand the impact of discharge conditions on plume development were performed in Chapter 2. A numerical model of the Northern Gulf of Mexico (NGoM) region was developed, and process-oriented numerical experiments were conducted to explore the isolated impact of bottom topography and winds (Chapter 3). A realistically-forced simulation was employed to address the impact the offshore Gulf of Mexico circulation (Chapter 4). Finally, aspects of seasonal and inter-annual variability of the along-shelf and across-shelf transports of Mississippi River waters were addressed in Chapter 5. The main results are reviewed in this concluding chapter, with a discussion of their importance to the understanding of the dynamics of large river plumes, in particular the Mississippi River plume.

The first step of this dissertation was the application of the HYCOM model to the study of river plumes in an idealized setting, and the investigation of how the development and structure of an estuarine buoyant plume is affected by outflow properties and river discharge conditions. New model parameterizations for the freshwater inflow were developed and applied to modify the mixing of riverine inflow at the point of discharge. These parameterizations were closely related to physical processes, especially lateral and vertical mixing of the river discharge within the idealized estuary. It was found that changes in the estuarine vertical mixing changed the dynamics of the flow prior to reaching the receiving basin. These changes significantly impacted the near (bulge) and far (coastal current) fields of the buoyant plume in flat bottom conditions. The development of plumes varied from having large circular bulges to cases with no bulge development and complete deflection of the outflow in the
downstream direction. Additional experiments in sloping bottom conditions demonstrated a two-way interaction between the plume and the estuarine dynamics. It was shown that the plume vertical structure was insensitive to various choices of vertical coordinates allowed by the hybrid approach in HYCOM. From the scientific point of view, the process-oriented results expand on previous findings about the importance of estuarine dynamics on the dynamical balance of outflows and of the buoyant plume, and they provide a contribution to realistic studies. From the technical aspect, the results demonstrate that the HYCOM model is able to reproduce major processes in river plume dynamics. The findings offer a benchmark for coastal studies with HYCOM, where plume dynamics should be examined in tandem with additional shelf and offshore circulation dynamics that can benefit from a flexible vertical coordinate system.

The second part of this study addressed the specific dynamics of the Mississippi River (MR) plume. A high resolution numerical model of the Northern Gulf of Mexico region was developed and the objective was to investigate the buoyancy-driven and wind-driven dynamics of the MR plume. Experiments were performed in “controlled” scenarios where the river and winds were the only forcing mechanisms, and focus was given to the mechanisms that favor the offshore dispersion of the plume. The buoyancy-driven circulation of the broad MR plume was in accordance with the development of a large-scale river plume that is influenced by the earth’s rotation, and the plume circulation was affected by the complex coastal and bottom topography of the region. Although the steep bottom topography near the Delta promotes retention of freshwater on the shelf, a fraction of the large outflow is able to expand offshore, and that is a favorable
condition for offshore dispersion. Winds significantly impacted the dynamics and spreading of the MR plume. The response of the plume to wind forcing was in agreement with previous studies, and new insights on the impact of bottom and coastal topography over the wind-driven dispersion were obtained. Upwelling-favorable (westerly) winds promoted the largest offshore spreading of the plume, and results suggest that the Deltaic morphology enhanced the offshore spreading by diverting the coastal current to the offshore region. The steep bottom topography near the MR Delta also plays a major role in the wind-driven offshore removal of the plume, since a surface Ekman layer is able to develop. In those conditions, onshore (southerly) winds promote an eastward Ekman transport towards the deep regions of the DeSoto Canyon, and therefore also enhance the dispersion of the plume towards the offshore region. These findings highlight the necessity to expand the understanding of river plumes dynamics where Deltas and intricate topography are present, which may modify the traditional perspective of the buoyant plume circulation, namely a dominant coastal current in the direction of Kelvin wave propagation.

The above findings provided the background to evaluate the dynamics of the MR plume in a long-term, realistically-forced scenario. The NGoM model was nested in a data-assimilative regional Gulf of Mexico model which provided realistic boundary conditions that incorporated the circulation of the Loop Current (LC) and associated frontal eddies. Here, the core of this dissertation was addressed, which is the interactions of the MR plume with the Gulf of Mexico offshore circulation. It was found that interactions between the MR plume and offshore eddies is a frequent, year-round process,
and that offshore removal is a frequent pathway for the plume. Eddies with diameters ranging from 50 to 150 km impinge against the shelfbreak and effectively entrain the water of the plume, with the formation of low-salinity bands that expand into the Gulf interior. The formation of eddy dipoles (with the contribution of shelf and shelfbreak eddies, in addition to LC frontal eddies) also affected the interactions with the plume. Offshore currents generated by anticyclone-cyclone pairs enhanced the offshore entrainment, whereas onshore currents generated by cyclone-anticyclone pairs could completely eliminate the offshore transport. The proximity of eddies to the shelfbreak is a sufficient condition for eddy entrainment, and shelf-to-offshore interactions near the MR Delta are facilitated by the steep bottom topography. The interactions between the plume and eddies are enhanced in the presence of eastward, wind-driven currents that transport the plume towards the deep DeSoto Canyon. During such wind events, the plume shelf circulation may be completely reversed towards the east, where offshore eddies entrain, stir and disperse the plume waters. Different than the interactions with “smaller” eddies, the interactions between the MR plume and the LC system produced the formation of strong, narrow low-salinity bands that reached the southern open boundary of the model. Such findings demonstrate the efficacy of the LC system in transporting MR waters into the interior of the Gulf, which is in agreement with previous observational studies.

Estimates of freshwater fluxes across the shelf demonstrated that the plume offshore transport may be as large as the along-shelf transport. It was found that largest volumes of freshwater were removed offshore in the presence of eastward, wind-driven currents. Winds that favor eastward plume transport may last for 1-2 months during the summer,
and up to 40 km$^3$ of riverine freshwater may be removed offshore during their duration. It was estimated that isolated eddy events could entrain up to 16 km$^3$ of freshwater. The actual volume of freshwater exported offshore is also a function of the MR discharge; eddies could entrain more than 50% of the MR discharge at the time of the entrainment event, given that the eddy was “correctly” positioned near the freshwater source. Here, it is found that the most effective conditions for the offshore exportation of MR waters are the proximity of large eddies in the presence of eastward, wind-driven transport.

The LC system is an important component of MR plume dynamics. Here, it is shown that the entrainment of MR waters by the LC is not an anomalous phenomenon, but presents considerable inter-annual variability. Plume to LC interactions are determined by episodic northward intrusions of the LC system in the NGoM region. If the intrusion is very close to the MR Delta, plume waters can be entrained regardless of wind conditions. If the intrusion is farther offshore and lasts for several months, the LC entrainment may be modulated by wind forcing and plume transport can switch from an along-shelf to an exclusive across-shelf regime. Other environmental factors, such as river discharge and shelf stratification, also play a role in the plume to LC interactions, but the dominant factors are: proximity to the MR Delta (primary) and wind influence (secondary). The local (NGoM shelf) plume to LC interactions are the first step in the across-basin plume transport, and the above findings are very relevant for the understanding of the connectivity between the MR waters and remote coastal systems, such as the South Florida region and the Florida Keys.
It was estimated that, on average, ~ 12 % of the year-round offshore freshwater transport near the MR Delta is due to plume to LC interactions. However, this percentage can go up to 30 % in individual years. During the interactions between the plume and the LC, it was estimated that an average value of LC freshwater entrainment was ~ 4,150 m$^3$ s$^{-1}$. This is the first time such estimates are made, and they provide valuable information on the role of the LC in the MR plume freshwater balance. Those estimates are also useful for studies in the NGoM region that may need to parameterize the role of the LC on the fate of MR waters and associated nutrients, sediments and pollutants.

A major conclusion from this study is that the Gulf of Mexico offshore circulation is an integral part of the dynamics of the Mississippi River plume. The effect of offshore boundary currents on the dynamics of the plume must be considered in conjunction with the shelf circulation, in order to obtain a complete picture of the processes determining the fate of plume waters. The findings of this work can be employed for the understanding of plume dynamics in the presence of energetic boundary flows in general, and for the design of an integrated observational and modeling system in the Northern Gulf of Mexico, in particular.
References


