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A Numerical Model Calculation of the Flow in DeSoto Canyon in Response to Northerly Wind Bursts in Winter

Y. HSUEH AND YURY GOLUBEV

The continental shelf currents observed at the head of DeSoto Canyon offshore of Pensacola, FL, during Nov. 1997–Feb. 1998, are studied with the aid of a Bryan–Cox model of the entire Gulf of Mexico. The basic model circulation with no winds features a Loop Current and a weak flow in DeSoto Canyon. In contrast, the model response to northerly wind bursts produces strong canyon currents. Wind-driven southward currents on the West Florida Shelf (WFS) appear to pull a strong flow around DeSoto Canyon, giving rise to a strong up-canyon flow on the west flank. The wind-enhanced flow in DeSoto Canyon is reproduced with a model run that incorporates wind stresses derived from the NCEP–NCAR reanalysis winds for March 1997–April 1998. The model flow for the winter period (Nov. 1997–Feb. 1998) is compared with observed currents at three canyon-head locations in 100 m of water available from a Science Applications International Corporation mooring experiment. Model and observed velocities in the general around-canyon direction track each other reasonably well, with peaks synchronized roughly with but lagging slightly northerly wind bursts. The lag time appears consistent with the generation at Key West of first-mode continental shelf waves responsible for wind-driven alongshore currents on the WFS. Pressure gradient slightly in excess of that necessary for a geostrophic balance with the (eastward) alongshore flow appears to generate onshore flows, giving rise to midshelf upwelling after northerly wind bursts.

The wintertime, wind-driven continental shelf currents in DeSoto Canyon are studied using a Bryan–Cox model of the entire Gulf of Mexico. DeSoto Canyon is a broad, northward trending submarine canyon, south of the Florida Panhandle coast, in the northeastern Gulf of Mexico (NEGOM), with the 200-m isobath just 72 km away from the coast (at approximately 30°24'N) (Fig. 1a). The width of the canyon measured at the 100-m isobath along 29°30'N is about 120 km, much greater than the 24-km internal deformation radius based on climatology. It is known that warm Gulf-of-Mexico water can move shoreward along the western rim of the canyon to within 8 km of the shore during northerly wind bursts in winter (Huh et al., 1981). A recent study of the circulation in the NEGOM (Yuan, 2002) suggests that bottom friction and wind-driven pile-up of water to the south may force an onshore flow.

Additional current measurements have since become available that bear upon the intrusion issue. These data come from the Science Applications International Corporation (SAIC) mooring experiment of March 1997–April 1998, which includes three head-of-the-canyon moorings along the 100-m isobath, C1 (87°21.0'W 29°35.15'N), D1 (86°50.5'W 30°04'N), and E1 (86°19.73'W 29°42.0'N) (see Fig. 1b) (SAIC,

1998). The purpose of the present study is to present a comparison of the model current with these measurements for the wintertime intrusion flow in DeSoto Canyon and to suggest, on the basis of model dynamics, that the intrusion flow may be a wind-driven response of continental shelf water in NEGOM as a whole. This basic response seems to explain the long-suspected episodic intrusion of deep Gulf water along the western rim of the Canyon (Schroeder et al., 1994). The study is aided by the use of a Bryan–Cox model of the entire Gulf of Mexico that crudely reproduces the Loop Current and its influence on the continental shelf (Yuan, 2002).

Submarine canyon flows have been studied on both coasts of the United States, mostly with canyon width less than or on the order of the internal deformation radius (Klinck, 1996; Hickey, 1997; She and Klinck, 2000). The wind-driven continental shelf flow in the vicinity of such deep and narrow canyons exhibits characteristics consistent with intense generation of relative vorticity as a result of vortex stretching (Hickey, 1997). In the case of broad canyons, as is the case at hand, where the deformation radius is much less than the radius of curvature of the isobaths, the dynamics are predominantly linear (Klinck, 1996). In this case, the flow is expected to go approximately along iso-

Fig. 1. (a) Topography of the Gulf-of-Mexico model. Contours shown are 100, 500, 1,000, 2,000, and 3,000 m. Solid circles, F and G, mark locations where inflow-outflow model experiment time series are extracted. The solid square at the top marks Pensacola, where the model wind-stress time series is shown. (b) SAIC moorings. Data from C1, D1, and E1 form the basis for model verification.

baths and to exhibit the familiar linear dynamics (Hsueh, 1980). In the concept of such a flow, DeSoto Canyon thus becomes an extension of the West Florida Shelf (WFS). It has been shown that wind-driven sea-level heights over the WFS are reasonably explained by the linear long-wave theory (Mitchum and Clarke, 1986). It stands to reason that the flow in DeSoto Canyon is just a part of this wave response. In addition, in the long-term, Sturges and Evans (1983) find that the 8-mo band sea-level fluctuations along the west Florida coast, interpreted as southward currents, are coherent with, but lead, the northward penetration positions of the Loop Current, indicating Loop Current influence.

To study the intrusion in DeSoto Canyon with both these forcing possibilities, the model in Yuan (2002) of the entire Gulf of Mexico is used (see Fig. 1a for the model domain). The model is of the Bryan-Cox type (Cox, 1984), with a rigid lid with increased vertical resolution in the upper water column to resolve the topography of the continental shelf and slope west of the Florida peninsula. [For future ref-

erence, the rigid-lid assumption produces a slight error in the phase speed of the continental shelf wave. The error is on the order of the square of the ratio of shelf width to the barotropic deformation radius, thus quite negligible even for a shelf as wide as the WFS (Clarke, 1977).] The forcing in the model includes wind stress at the surface and open-boundary conditions in the Caribbean Sea and in the Straits of Florida. The southern boundary of the model cuts through the Caribbean Sea so that the structure of the inflow through the Yucatan Straits is not specified as in Hurlburt and Thompson (1980). A volume transport of 30 Sv is imposed across the southern open boundary and through the Straits of Florida to represent the mean Loop Current transport. No attempt has been made to study the model sensitivity to the flow specification at these open boundaries. The purpose is to achieve a Gulf-of-Mexico model, with a broad-scale representation of the circulation, to serve as background for a few process studies in continental shelf waters.

A perspective of the model against the background of numerous models of the Gulf of Mexico can be found in Yuan (2002). It suffices to say that the inertial nature of the Loop Current is better preserved here than in most previous models. Oey (1996) uses improved vertical resolution to include a continental shelf and successfully simulates the irregular shedding of Loop Current rings. The Loop Current and the migration path of its rings in Oey (1996), as in the present model, bend unrealistically to the west, hinting at an overly diffusive representation, perhaps partly attributed to a lack of horizontal resolution. Even though the present model uses a biharmonic formulation, the flow is still too frictional so that there is a complete absence of filaments and parasite eddies often observed around the Loop Current (Perez et al., 1999), and the model Loop Current never reaches so high a latitude as the observed 28°N (Sturges, 1992).

The outline of the article is as follows. In the next section, a description is given of the model configuration, and then in the section on An Inflow-Outflow Experiment are reported model flows driven by the Loop Current. The section A Hindcast for Nov. 1997–Feb. 1998 considers the response of the WFS and the DeSoto Canyon waters to winds during Nov. 1997–Feb. 1998, the SAIC experiment winter period. The dynamics of the flow are discussed in the discussion section. The last section contains the conclusions.

THE MODEL

As mentioned earlier, the model used in this study is the z-level model of Bryan-Cox (Cox, 1984) described in Yuan (2002). The model domain covers the entire Gulf of Mexico, from 18°N to 31°N and from 100°W to 80°W, with topography obtained from the 5-min ETOPO global relief model (see Fig. 1a). The horizontal grid resolution is $1/6^\circ$ by $1/6^\circ$. The total number of levels in the vertical is 30, with larger vertical spacings in the deeper ocean. The smallest vertical spacing is 10 m at the sea surface, small enough only to give the Ekman veering and transport crucial for the Ekman-geostrophic response in coastal waters, which is more important here than the detailed Ekman layer structure not resolved. The number of levels above the 200-m depth is 8, and the number of levels above the 1,000-m depth is 16. With this grid, both the continental shelf and the continental slope are reasonably resolved.

At the southern open boundary across the Caribbean Sea, a 30-Sv ($1\text{ Sv} = 10^6 \text{ m}^3 \text{ s}^{-1}$) volume transport is specified across a 100-km section next to the Yucatan Peninsula to represent nominally the western boundary current of the Caribbean Sea. The same volume transport leaves the model domain through a 100-km section in the northern Straits of Florida at the eastern open boundary. A five-grid-interval buffer layer is used along the open boundaries to relax salinity and temperature fields to their climatological values (Levitus and Boyer, 1994; Levitus et al., 1994) and the velocity to the geostrophic velocity. The geostrophic velocity at the open boundaries is obtained by combining the barotropic velocity based on the specified transport with the thermal-wind velocity calculated from the climatological tracers fields. The relaxation coefficients at the open boundaries correspond to time scales of a few hours. Weak relaxation with a time scale of 50 d is used at the sea surface for temperature and salinity. Radiation boundary conditions for temperature, salinity, and velocity similar to those of Carmelengo and O'Brien (1980) are used on the eastern open boundary. The strategy of the open-boundary condition specification has been explained in detail in Schultz (1994) and in Hsueh et al. (1997). It must be

noted that the inflow and outflow conditions are probably far too simple to represent the full reality. Again, the aim here is to achieve in a broad sense an inflow-outflow configuration for the Gulf of Mexico for process studies. The effect of open-boundary conditions on the flow is a worthy study subject in its own right but is outside the present scope.

At closed boundaries, the no-slip velocity and no-normal-flux temperature and salinity conditions are employed. A biharmonic horizontal mixing formula is used in the momentum and tracer equations, with a biharmonic coefficient of $-4 \times 10^{18} \text{ cm}^4 \text{ s}^{-1}$. The use of a biharmonic formulation is intended to reduce damping and preserves the nonlinear characteristics of inertial currents (Holland, 1978). [The model, however, still proves too diffusive (see later).] Vertical mixing and diffusion are embodied in a Laplacian with a coefficient of $1 \text{ cm}^2 \text{ s}^{-1}$. The quadratic-law bottom friction is used in the momentum equations, with a friction coefficient of 10^{-3} . The model is initialized with climatological temperature and salinity fields and with a geostrophic velocity field obtained from adding the climatological thermal-wind velocity to the barotropic velocity forced by the imposed transports (Hsueh et al., 1997).

AN INFLOW-OUTFLOW EXPERIMENT

In order to assess the influence of the Loop Current on the circulation over the WFS and inside DeSoto Canyon, the output is examined of a model run (referred to as GOM-01) without the wind stress, as in Yuan (2002). Under the forcing of the open-boundary conditions, the inflow-outflow combination gives rise to a Loop Current in the Gulf of Mexico, which sheds anticyclonic rings of 300–400-km radius at a period of around 13 mo. The lack of a spread in separation periods points to the fact that the model leaves out important factors that contribute to the Loop Current behavior and ring separation (Sturges, 1994; Sturges and Leben, 2000). It does, however, conform to what is anticipated in an analytical theory (Pichevin and Nof, 1997).

Figure 2a,b shows, in eight panels, the evolution of the 5-m velocity over NEGOM

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Fig. 2. Evolution of velocity at the 5-m depth in the inflow-outflow experiment in eight panels dated in integration days and numbered in the upper right corner. The 200-m and 1,000-m isobaths are plotted to identify the shelf area.

Day 2070

Day 2190

Day 2130

Day 2250

Day 2310

Day 2430

Day 2370

Day 2490

Fig. 3. Time series of north-south velocity component (dashed) at 5-m depth at 84°W , 28°N (G in Fig. 1a) on the WFS in the inflow-outflow experiment. The upper panel shows the time series during more than 5 yr of integration. The lower panel shows the boxed portion of time series in the upper panel. The expanded portion covers the time window of Figure 2. The horizontal lines are zero velocity lines.

through a typical ring-shedding cycle of the model integration. On Day 2,070 of the run (panel 1), an anticyclonic ring is clearly seen separated from the main stream of the Loop Current, and the Loop Current has retreated to near its southernmost position. The isolated current ring proceeds to move to the west and southwest at a speed of about 4 cm s^{-1} as is found elsewhere (Sturges et al., 1993). In the meantime, the Loop Current begins to grow northward. In early stages in this growth (panels 3–4), a shelf-break jet is seen formed, apparently from the impingement of the Loop Current on the WFS (see later). As the Loop Current reaches nearly its maximum northward penetration, with its axis at its closest to the WFS (panels 5–6), a weak cyclonic eddy is gaining in strength southwest of Key West. This cyclonic eddy grows and moves to the west and causes a “necking-down” of the Loop Current, leading to the eventual separation of an anticyclonic current ring (panels 7–8) (Vukovich et al., 1979; Pratt and Stern, 1986; Vukovich, 1986). The spawning of this cyclonic circulation cell and its possible triggering effect for ring separation have been widely acknowledged (Hurlburt and Thompson, 1982; Dietrich and Lin, 1994; Frantantoni, 1998). It takes about a year for the model Loop Current to regain its full penetration and start another round of ring shedding. The model ring-shed-

ding cycle repeats itself and constitutes the anticipated simplified periodic solution found in theoretical studies (Pichevin and Nof, 1997). In reality, the ring shedding is quite aperiodic, and the appearance of the Loop Current is irregular with filaments and parasite eddies, exposing the overly diffusive nature of the present model run (see Oey, 1995 for a successful model in this regard).

The influence of this simple model Loop Current evolution on the WFS circulation is shown in Figure 3. The upper panel shows the evolution of the north-south velocity component (dashed) for more than 5 yr at the 5-m depth in 40 m of water at the solid circle marked G in Figure 1a (84°W 28°N) off Tampa. The solid curve is for the east-west velocity, which is much weaker. The velocity at this location is qualitatively representative of the velocity on the WFS. The mean current over the time span is southward, suggesting that, on the average, the Loop Current generates a shelf flow to the south. The oscillation on the ring-shedding period appears phase-locked to the ring-shedding cycle. The time span of each trough (large southward velocity) roughly begins with the time when a ring has just separated (panels 2 and 3 in Fig. 2).

The lower panel shows an expanded portion of the north-south velocity time series. The expanded portion corresponds to the time win-

Fig. 4. Time series plot same as in Figure 3 but at $87^{\circ}\text{W } 29.5^{\circ}\text{N}$ (F in Fig. 1a) inside DeSoto Canyon. Solid curve represents the east-west velocity component, and dashed curve represents the north-south component. The horizontal lines are zero velocity lines.

dow spanned by Figure 2. Notice that, at the time (marked 2370) when the Loop Current front is closer to the WFS, the southward drift over the shelf is weaker than, say, that on Day 2,250 when the front is farther away. This suggests that the flow over the shelf is not simply generated frictionally by the Loop Current. The forcing of the Loop Current on the shelf circulation must be through other mechanisms. As alluded to in Yuan (2002), it turns out that the drop in dynamic pressure brought about by the Bernoulli effect as the Loop Current careens along the continental slope-shelf edge, after it collides with the continental mass, effects just such a flow in accordance with the arrested topographic wave (ATW) theory (Csanady, 1978; Csanady and Shaw, 1983). The manner in which this along-shelf-edge pressure change is imposed varies with the Loop Current position. This change in shelf-edge pressure as a function of the Loop Current position induces a corresponding change in shelf flow, making the southward flow to appear to lead the northward position of the Loop Current (Sturges and Evans, 1983). The detailed steady-state physics of this pressure forcing process is discussed elsewhere (Hetland et al., 1999). Based on the evidence so far presented, it suffices to say that the model Loop Current generally forces a southward flow over the WFS, and the low-frequency var-

iations of the flow occur primarily at the Loop Current ring-shedding frequency.

The influence of the model Loop Current variation on the circulation inside DeSoto Canyon is shown in Figure 4, in which are plotted, over the same time period as in Figure 3, the velocity time series at 5-m depth in 324 m of water at the solid circle marked F in Figure 1a ($87^{\circ}\text{W } 29.5^{\circ}\text{N}$) at the center of DeSoto Canyon. The solid curve represents the east-west velocity component, and the dashed curve represents the north-south velocity component. The east-west component is particularly comparable to the north-south component shown in Figure 3, both being roughly alongshore. The east-west component time series in Figure 4 shows more irregularities, but the phase-lock with the Loop Current variation emerges again. Again, larger velocity tends to occur at a time when the Loop Current is farther away to the south (Day 2,250 vs Day 2,310). The alongshore flow in DeSoto Canyon appears simply to respond to the southward flow over the WFS. This is again consistent with the ATW theory and attributable to the insulating effect of the continental slope topography, confining the shelf flow to roughly follow the isobaths. The east-west velocity component at the solid circle F thus varies similarly with the north-south velocity component at G, 180° out of phase with the extent of the northward pene-

Fig. 5. Horizontal velocity charts at 16.4 m in the DeSoto Canyon area in the inflow–outflow experiment. The times of the panels correspond to those in Figure 2. The 100-m isobath is drawn to identify DeSoto Canyon.

tration of the Loop Current. A time–distance contour plot (not shown) of east–west velocity at 5 m along the meridian through F from 30°N southward to 26°N shows that not until the latitude of 27°N is reached does the east–west velocity begin to vary in phase with the northward penetration of the Loop Current.

Figure 5 shows the velocity vector plots at 16.4 m in the DeSoto Canyon area at the time of the eight panels in Figure 2. The Loop Current–driven flow inside DeSoto Canyon remains weak and basically follows the isobaths. Because the observed intrusion in DeSoto Canyon suggests strong currents, causes in addition to the low-frequency influence of the Loop Current must be found to explain the intrusion. It must be stressed that the results of weak direct Loop Current influence in both the WFS and DeSoto Canyon are probably a model artifact because of the inability of the model in resolving Loop Current filaments and parasite eddies. Model runs with better resolution and mixing representation than used in the present run are needed to address the influence of these small-scale Loop Current structures. With their influence sup-

pressed, the model presents an opportunity for studying, in almost a default isolation, other factors influencing continental shelf flows in NEGOM, such as the wind.

A HINDCAST FOR NOV. 1997–FEB. 1998

Direct current measurements made by the SAIC in NEGOM, particularly at the head of DeSoto Canyon, for March 1997–April 1998, have become recently available. To take advantage of the opportunity thus provided, a hindcast (GOM-09) for currents is conducted with the imposition at the model sea surface of 6-hourly wind stresses converted from the 10-m surface winds from the NCEP–NCAR 40-yr reanalysis data (Kalnay et al., 1996), using a drag coefficient of 1.3×10^{-3} . [Time series plots (not shown) of NCEP–NCAR winds together with those observed at Pensacola and at a number of buoys in the northeastern Gulf of Mexico show that the two track each other closely most of the time.] The initial condition is the GOM-01 output at Day 2,070 (panel 1 in Fig. 2). A TOPEX/Poseidon sea-surface height anomaly map for March 1, 1997

Fig. 6. Plot of the model-used 6-hourly wind stress at Pensacola in Nov. 1997–Feb. 1998. The upper panel shows the time series of the east–west (thin) and north–south (heavy) components of the wind stress. The lower panel shows the stick plot of the daily averaged Pensacola wind stress. Times of northerly wind bursts discussed are marked with heavy tick marks. In the middle panel, the model north–south wind-stress component time series at a point (84.45°W 27.07°N) on the WFS, south of G, is shown.

(not shown) shows approximately the same Loop Current position as in the chosen frame of the GOM-01 output.

Figure 6 shows the model wind-stress time series at Pensacola for the selected hindcast period. The middle panel shows the model north–south wind stress at a point south of G at (84.45°W 27.07°N). A number of northerly wind bursts are clearly identifiable (Nov. 7, 15, 23, Dec. 5, 12–15, 27, and Feb. 4, for example). Each burst lasts for several days and then collapses. Figure 7 presents a comparison between the hindcast and the inflow–outflow experiment (GOM-01) in terms of DeSoto Canyon area velocity vector plots at a time about 1–2 d later of several of these wind burst events (with Day 2,070 in GOM-01 as March 1, 1997). The hindcast clearly produces strong onshore flows on the west flank of DeSoto Canyon not otherwise present. To verify this hindcast model response, use is made of SAIC direct current velocity measurements during Nov. 1997–Feb. 1998 at C1 ($87^{\circ}21.0'\text{W}$ $29^{\circ}35.15'\text{N}$), D1 ($86^{\circ}50.5'\text{W}$ $30^{\circ}04'\text{N}$), and E1 ($86^{\circ}19.73'\text{W}$ $29^{\circ}42.0'\text{N}$) (see Fig. 1b) (SAIC, 1998). Figure 8 shows time series comparisons, at two interior depths 16 and 32 m, of the around-canyon and crosswise velocity components from data

with the corresponding calculation from the model. (Because of the bottom slope, 48 m represents the last depth for which model velocities are calculated at D1.) The local around-canyon directions (positive y-axis) are 45° , 75° , and 170° clockwise from north for C1, D1, and E1, respectively. The crosswise direction (positive x-axis) is generally offshore. At all locations, there is a measure of similarity in the around-canyon component between the observation and the hindcast. In particular, the strong model flow, as indicated by the around-canyon velocity peaks shortly after northerly wind bursts of Nov. 7, 15, 23, Dec. 5, 12–15, 27, and Feb. 4, seems well substantiated by the direct measurements. In all events, the current slightly lags the wind. This is consistent with the notion that the DeSoto Canyon flow responds to the southward flow on the WFS, which is established by the passage of continental shelf waves originated at the end of the WFS after the onset of each wind burst (Mitchum and Clarke, 1986). Figure 9 shows the coherence squared and phase between the north–south wind component and the around-canyon current at station D1 at the depth of 16 m. In the 3- to 10-d band, a lag of a couple of days is detected for the around-canyon cur-

Fig. 7. Model velocity vector plots at 16 m in the DeSoto Canyon area during northerly wind bursts. Plots are for Nov. 8, 17 (Fig. 7a), Dec. 7, 16 (Fig. 7b), and Dec. 31 and Feb. 5 (Fig. 7c). For each case, the corresponding plot from the inflow-outflow experiment is shown in the top panel. The plot from the hindcast is in the middle. The plot of the vertical velocity in contours of $2 \times 10^{-3} \text{ cm s}^{-1}$ appears in the bottom panel, where the zero contour is dashed, and negative values are contoured in dotted curves. Areas of particularly large positive values are marked "U." The 100-m isobath is drawn to identify DeSoto Canyon.

Fig. 7c. Continued.

rents, both observed and modeled, consistent with the first-mode continental shelf wave phase speed for the WFS (Mitchum and Clarke, 1986). There is little similarity in the crosswise component, which is much weaker. The difficulty is typical of that encountered in modeling cross-shore flows on the continental shelf (Hsueh, 1989).

There remains the question of the degree to which the around-canyon flow is locally driven by the east–west wind bursts of which there are several at the time of north–south wind bursts (for example, Dec. 31 and Feb. 4). In order to address this issue, three model runs with a constant wind stress to the southeast (at 1 dyne cm^{-2} for the components) are conducted from a common initial condition: one with the wind over the entire Gulf of Mexico, one with the wind covering only the domain east of 87°W , and one with the wind covering the domain east of 88°W , with the wind stress ramping from zero at 88°W to the constant value of 1 dyne cm^{-2} at 87°W for both components. Against these runs, a control run is made with no wind forcing. In each case, the model is run for 15 d. (The continental shelf flow pattern becomes steady at about the 10-d mark.) The around-canyon velocity at station D1 at 32 m from the run with no wind is then subtracted

from that from the three experimental runs. The wind-driven around-canyon velocity, thus obtained, differs very little (3 cm s^{-1} in 40 cm s^{-1}), with the one from the experiment with the forcing ramp the greatest. This result indicates empirically that the local wind-driven around-canyon velocity is small. Theoretically, one could estimate the amplitude of the first-mode continental shelf wave at D1 by integrating along the characteristics (in this case, the 100-m isobath) the along-characteristics wind stress from Key West to D1 (see, for example, Mitchum and Clarke, 1986). Because the east–west running portion of the isobath is relatively short compared with the north–south running portion on the WFS, it stands to reason that the amplitude contribution from east–west wind forcing in the neighborhood of D1 will be small. A few words for the event around Feb. 4 are in order. Although the winds at Pensacola are northerly, those on WFS are actually southerly a couple of days earlier (see Fig. 6, middle panel). This southerly wind drives a northward coastal flow, leading to a negative around-canyon velocity on Feb. 4 (Fig. 8a).

The idea that the DeSoto Canyon flow is but a part of the continental shelf flow over the NEGOM is demonstrable with the hindcast model output. Figure 10 shows a time–distance

Fig. 8. Time series plots of (a) around-canyon and (b) crosswise velocity components at 16 m and 32 m at C1, D1, and E1 for Nov. 1997–Feb. 1998. The model results are dashed, and observations are solid. The wind-stress component time series shown in Figure 6 is reproduced in the upper panel for easy comparison.

Fig. 9. Coherence squared (γ^2) and phase (θ) with cycles per day (cpd) between the 6-hourly north-south wind component and 6-hourly around-canyon velocity at station D1 at the depth of 16 m for (a) modeled currents and (b) observed currents. The dashed line in the coherence squared plots indicates the 95% confidence level. Negative phase indicates the wind leads. Phase is calculated only for the frequency band in which the coherence squared is above the 95% confidence level.

Fig. 10. Time-distance plot of 16 m north-south velocity components along the segment of the parallel at $29^{\circ}40'N$ spanned by C1, D9, and E1. Model results are in the middle panel, and observations are in the lower panel. The wind-stress component time series shown in Figure 6 is reproduced in the top panel. The contour interval is 4.5 cm s^{-1} .

plot of the north–south velocity at 48 m for the hindcast period along the parallel at 29°40'N for the segment spanned by C1, D9, and E1 (see Fig. 1b). (No data return at shallower depth than 48 m at D9.) During this winter period, it is clear that the flow is predominantly northward (red) on the west half and southward (blue) on the east. The contrast is particularly strong for the times of northerly wind bursts. The plot from interpolation from observed currents at C1, D9, and E1 shows a similar tendency. Thus, the flow around DeSoto Canyon is clockwise under these northerly wind bursts, connecting the flow off the Florida Panhandle with that of the WFS.

An important consequence of strong currents in DeSoto Canyon after northerly wind bursts is that it leads to upwelling. In the bottom panels of Figure 7 are found model vertical velocity contour plots in the DeSoto Canyon area during the six intrusion events in response to the northerly wind bursts. Areas of particularly large positive vertical velocities are marked “U.” Clearly, there is a strong upwelling center at the head of DeSoto Canyon as a result of onshore flows associated with strong around-canyon currents. It has been suggested that upwelling here could be caused by the bottom Ekman flow under a strong clockwise current around DeSoto Canyon. Unfortunately, Ekman layer over a sloping terrain in a z -level model is decimated and not well resolved (Winton et al., 1998). As a consequence, north–south velocity at a near-boundary grid-point like the one at D1 is set to zero at the next level (72 m) below 49 m to satisfy the non-normal flow condition, leading to an exaggerated frictional effect at midcolumn depth. An effect that is represented reasonably correctly in the model that also leads to upwelling is the production of onshore flow as a result of a slight excess in pressure gradient in the geostrophic balance associated with a clockwise around-canyon currents (see later in Discussion). The onshore flow over the shoaling topography then leads to upwelling. Because of this upwelling associated with strong around-canyon currents and flow to the southwest on the east flank, it is perhaps not surprising that peak monthly mean pigment concentration along the 40-m isobath over the WFS should be found consistently for November through April (Gilbes et al., 1996).

DISCUSSION

It is argued here that strong currents in the DeSoto Canyon region result from the estab-

lishment of coastal currents over the WFS by the passage of continental shelf waves originating from the end of the WFS after each northerly wind burst. To substantiate this argument, the model term balance (with all terms moved to the right side of the equation) at 16 m in both the around-canyon and crosswise momentum equations at D1 (Fig. 1b) during the hindcast period of Nov. 1997–Feb. 1998 is shown in Figure 11 (nonlinear terms are negligibly small and not shown). The crosswise momentum is nearly in geostrophic balance, with the crosswise pressure gradient nearly balancing the Coriolis term arising from the around-canyon flow. In the around-canyon direction, the balance must include the local time derivative of the around-canyon velocity, suggesting a continental shelf wave–like dynamics. Continental shelf waves of the lowest four modes have been found to explain low-frequency coastal sea-surface height variations from Key West to Apalachicola (Mitchum and Clarke, 1986). Here, it is perhaps the first-mode continental shelf wave at the several-day period with a phase speed of about 800 km d⁻¹ that gives rise to the phase lag found in Figure 9.

Figure 12 shows at 16 m at D1 (in the lower panel) the time series of the excess crosswise pressure gradient (sum of the first two curves in the crosswise momentum balance in Fig. 11), negative of local derivative, and the (exaggerated) vertical friction. The north–south component of wind stress at Pensacola (upper panel) and the model around-canyon velocity at 16 m at D1 from Figure 8 (middle panel) are reproduced here for easy reference. It is interesting to note that after northerly wind bursts, when the around-canyon velocity is generally eastward, the excess pressure gradient force is onshore, balancing a crosswise local onshore acceleration. This is particularly clear around Dec. 7, 31, and Feb. 4. The excess is perhaps because of the pile-up of surface water offshore under the northerly wind. [Figures 8 and 9 in Yuan (2002) showing in the DeSoto Canyon area model surface pressure gradient diagnosed for the Nov. 7 northerly wind event seem to confirm this in that the greatest offshore gradient (high pressure offshore) is found just after the passage of strong northerly winds.] It is the crosswise acceleration caused by this pressure gradient that gives rise to a flow across the shoaling topography, leading to upwelling in Figure 7. The contribution from vertical friction is small but not always negligible. A model with terrain-following coordinates is needed to properly assess the effect of the bottom Ekman flow, which can also pro-

Fig. 11. Around-canyon (left panel) and crosswise (right panel) momentum balance are shown at the 16.4-m depth at D1 at the head of DeSoto Canyon. Here, positive y direction is eastward, and positive x direction is offshore. From top to bottom in the left panel, the curves represent $-fu$, $-\partial p/\partial y$, $-\partial u/\partial t$, nonlinear term, and $A_v(\partial^2 v/\partial z^2)$. From top to bottom in the right panel, the curves represent fu , $-\partial p/\partial x$, $-\partial u/\partial t$, nonlinear term, and $A_v(\partial^2 u/\partial z^2)$. The unit is $10^{-4} \text{ cm s}^{-2}$. The horizontal line marks zero for each curve.

Fig. 12. Bottom panel. Crosswise pressure gradient excess [sum of top curves in right panel of Fig. 11, $fu - (\partial p/\partial x)$] at 16 m at D1 for the hindcast period. The two curves in Figure 11, $-\partial u/\partial t$ and $A_v(\partial^2 u/\partial z^2)$, are reproduced in the same order here as the lower curves. Upper panel. North-south wind-stress component at Pensacola. Middle panel. Around-canyon velocity component at 16 m at D1, reproduced from Figure 8.

duce upwelling under an eastward around-canyon current.

CONCLUSIONS

A numerical study of both the Loop Current-driven and wind-driven circulation in the northeastern Gulf of Mexico is conducted. The Loop Current of the Gulf of Mexico generally imposes an eastward drift in DeSoto Canyon and a southward drift over the WFS. The Loop Current-driven drift inside DeSoto Canyon produces only a weak flow. But this picture of a suppressed role of the Loop Current could be attributed to the fact that the model is deficient in representing filaments and parasite eddies of the Loop Current that could extend the Loop Current influence to the continental shelf. Given this model deficiency, northerly winds are found to generate large around-canyon water movements in DeSoto Canyon. The dynamics of the wind-driven flow appears primarily to be that of the setup of coastal currents by continental shelf waves all along the NEGOM shelf.

An important aside is the discovery that strong upwelling is found in the model at the head of DeSoto Canyon during the northerly wind bursts. This is apparently, in part, the result of excess in pressure gradient in the cross-wise momentum balance brought by the northerly wind pile-up (Yuan, 2002). The pressure gradient excess accelerates flow onshore across the shoaling topography, leading to upwelling. Upwelling could also be caused by bottom Ekman flow under the strong clockwise circulation around DeSoto Canyon after northerly wind bursts. Unfortunately, the present model does not resolve the Ekman layer structure over the sloping shelf. Further clarification of the contribution of bottom friction to upwelling in DeSoto Canyon must await future studies using a model with terrain-following coordinates.

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