Why Can Wind Delay the Shedding of Loop Current Eddies?

Y.-L. CHANG
Princeton University, Princeton, New Jersey, and National Taiwan Normal University, Taipei, Taiwan

L.-Y. OEY
Princeton University, Princeton, New Jersey

(Manuscript received 4 March 2010, in final form 4 May 2010)

ABSTRACT

It is first shown that wind in the Gulf of Mexico can delay the shedding of Loop Current eddies. A time-dependent, three-dimensional numerical experiment forced by a spatially and temporally constant westward wind stress within the Gulf is analyzed and then is compared with an otherwise identical no-wind run, and the result is confirmed with reduced-gravity experiments. It is shown that the wind produces westward transports over the northern and southern shelves of the Gulf, convergence in the west, and a returned (i.e., eastward) upper-layer flow over the deep central basin toward the Loop Current. The theory from T. Pichevin and D. Nof is then used to explain that the returned flow constitutes a zonal momentum flux that delays eddy shedding. Mass-balance analysis shows that wind also forces larger Loop Current and rings (because the delayed shedding allows more mass to be accumulated in them) and produces more efficient mass exchange between the Gulf and the Caribbean Sea. It is shown that eddies alone (without wind stress curl) can force a boundary current and downward flow in the western Gulf and a corresponding deep flow from the western Gulf to the eastern Gulf.

1. Introduction

The Gulf of Mexico is a semiclosed sea in the southwestern tropical–subtropical Atlantic Ocean (Fig. 1). The Gulf has two narrow openings. The Yucatan Channel is V shaped between the Yucatan Peninsula (west) and Cuba (east); it is approximately 200 km wide at the surface and 25 km at its sill depth of approximately 2040 m (Sheinbaum et al. 2002) and connects the Gulf to the Caribbean Sea. The Straits of Florida is between Cuba (south) and Florida (north); it is approximately 100 km wide at the surface, has a shallower sill depth of about 800 m, and connects the Gulf to the Atlantic Ocean.

The Yucatan Current is a narrow and swift western boundary current (WBC) along the eastern coast of Yucatan Peninsula in the northwestern Caribbean Sea (the Cayman Sea). Its mean speed is about 1 m s\(^{-1}\), but at any time this can exceed 2 m s\(^{-1}\) (Cetina-Heredia et al. 2006). The Yucatan Current supplies approximately 23–27 Sv (1 Sv = 10\(^6\) m\(^3\) s\(^{-1}\); Schmitz 1996; Candela et al. 2002) of warm and relatively more saline waters from the Caribbean Sea into the Gulf of Mexico (Nowlin 1972). Temperatures \(T\) (potential) in the near-surface 100–200 m are generally above 26°C with salinities \(S\) of about 36.4 psu. At \(T \approx 6.3°C\) in the subsurface depths \(z \approx -1000\) m, the salinity reaches a minimum of \(<34.9\) psu, which is a characteristic of the remnant of Subantarctic Intermediate Water (Wüst 1964). In the upper 500–1000 m, the Yucatan Current intrudes northward into the Gulf along the Campeche Bank’s eastern shelfbreak and loops clockwise into the Straits of Florida. This strong clockwise-turning current in the eastern Gulf of Mexico is called the Loop Current. The Loop Current is highly inertial; because of its generally large size of \(\approx 300\) km or larger in its south–north and west–east extents, it is significantly affected by the variation of the Coriolis parameter \(f\); that is, by the \(\beta\) effects (Hurlburt and Thompson 1980; Nof 2005). Large warm-core rings (Loop Current eddies; 200–350 km wide and 500–1000 m deep) episodically separate from the Loop Current at time intervals (“periods”) that range from a couple of

Corresponding author address: L.-Y. Oey, Princeton University, Sayre Hall, 300 Forrestal Rd., Princeton, NJ 08544.
E-mail: lyo@princeton.edu

DOI: 10.1175/2010JPO4460.1

© 2010 American Meteorological Society
months to almost 2 yr (Sturges and Leben 2000). An idealized but useful idea of why eddies are shed is given by Nof (2005). The Loop Current grows with mass influx from the Yucatan Current. Up to a certain large size, the westward Rossby wave speed (which is $\approx -\beta R^2$, where $R$ is the Rossby radius based on the matured deep Loop Current) overcomes the growth rate and the Loop Current sheds an eddy. Because of their dominance, the Loop Current and Loop Current eddies affect just about every aspect of oceanography of the Gulf of Mexico (Oey et al. 2005).

As the papers by Pichevin and Nof (1997), Bunge et al. (2002), Ezer et al. (2003), Nof (2005), Lugo-Fernández and Badan (2007), and Sturges and Kenyon (2008) indicate, understanding mass balances in the Gulf may hold the key to also understanding the Loop Current and eddy dynamics. A numerical model is ideal for analyzing balances for the entire Gulf in a self-consistent way; however, despite many previous modeling efforts (for a review, see Oey et al. 2005), this does not appear to have been done. Also, previous studies have focused almost exclusively on upstream influences from the Caribbean Sea (for a review, see Oey et al. 2005), although Oey (1996) and Oey et al. (2003) did note the possible effects of local, time-dependent wind in producing irregular shedding periods. In this paper, we analyze transport balances and study how these are modified by a local, steady westward wind blowing inside the Gulf only (Gutierrez de Velasco and Winant 1996). We intentionally exclude wind stress curl so that we can focus solely on eddy transports. This has important (and interesting) consequences, as we will discuss. We also intentionally exclude remote wind effects from the Atlantic Ocean. Remote winds may cause transport and vorticity variations through the Yucatan Channel, the effects of which on the Loop Current’s behaviors were discussed in Oey et al. (2003, and references therein). The goal is then to understand transport pathways in the simplest model setting possible by comparing experiments with and without wind. The numerical model [the Princeton Ocean Model (POM); Mellor 2002] is time dependent and three dimensional based on the primitive equations assuming hydrostacy and the Boussinesq approximation. The model forcing is constant inflow transport into the Cayman Sea, zero surface heat and salt fluxes, and zero or constant westward wind stress. The model produces eddy shedding with a nearly constant period of about 8 months. This is idealized
but an advantage for our purpose of understanding processes. We will attempt to provide partial answers to the following questions:

1) How is the Yucatan inflow distributed in the Gulf of Mexico, and how much of the inflow goes into the recirculation within the Loop Current and eddies?
2) How is the westward eddy transport returned into the Straits of Florida and the Yucatan Channel?
3) Does local wind alter the pathways of return flow from the western to the eastern Gulf, and does it affect Loop Current dynamics including the eddy-shedding period? If so, then why?
4) If local wind alters the return transport pathways, does it then affect the redistribution of outflow transports through the Yucatan Channel and the Straits of Florida?

The outline of the paper is as follows: Section 2 presents the numerical model. Section 3 highlights the main differences in circulation between the wind and no-wind experiments. In section 4, we analyze the transport pathways in detail, show and explain the delayed-shedding result in the experiment with wind, and confirm it using a simple reduced-gravity model. We also explain why in the experiment with wind the Loop Current and eddies are larger. Section 5 derives the transport balances in the Gulf and through the Yucatan Channel and the Straits of Florida and shows how wind can force a more vigorous exchange of mass between the Gulf and the Caribbean Sea. Section 6 is a summary.

2. The numerical model

The terrain-following (i.e., sigma) coordinate and time-dependent numerical model for this study is based on the Princeton Ocean Model (Mellor 2002). The Mellor and Yamada (1982) turbulence closure scheme modified by Craig and Banner (1994) to affect wave-enhanced turbulence near the surface is used (Mellor and Blumberg 2004). A fourth-order scheme is used to evaluate the pressure-gradient terms (Berntsen and Oey 2010); in combination with high resolution and subtraction of the mean \( \rho \) profile, it guarantees small pressure-gradient errors of \( O(\text{mm s}^{-1}) \) (cf. Oey et al. 2003). The Smagorinsky (1963) shear and grid-dependent horizontal viscosity is used with coefficient \( = 0.1 \), and the corresponding diffusivity is set 5 times smaller (cf. Mellor et al. 1994). Two domains are used on orthogonal curvilinear grids (Fig. 1). The outer grid is the regional northwestern Atlantic Ocean model (NWAOM; \( 6^\circ-50^\circ \text{N}, 98^\circ-55^\circ \text{W} \)) reported in our previous studies (e.g., Oey et al. 2003). The NWAOM has 25 vertical sigma levels and horizontal grid sizes \( \Delta \approx 6-12 \text{ km} \) in the Gulf of Mexico. The World Ocean Atlas data (‘‘climatological’’ data) from the National Oceanographic Data Center (NODC; available online at http://www.nodc.noaa.gov/OC5/WOA05/pr_woa05.html) was used for an initial condition as well as a boundary condition along the eastern open boundary at 55°W. Across 55°W, a steady transport combined with radiation using also the geostrophically balanced surface elevation \( \eta \) (Oey and Chen 1992a) specifies the Gulf Stream exiting near the Grand Banks south of Newfoundland, with a magnitude of 93 Sv (W. J. Schmitz Jr. 2001, personal communication; see also Schmitz 1996; Hendry 1982; Hogg 1992; Hogg and Johns 1995). This is balanced by transports specified as broad return flows south (the Worthington gyre; Worthington 1976) and north (the north recirculation gyre; Hogg et al. 1986) of the jet. The vertical structures of the currents (i.e., after a depth-averaged value is removed) are specified using radiation conditions. The velocity component tangential to the boundary, as well as turbulence kinetic energy and length scale, are specified using one-sided advection scheme at outflow grids and are set zero at inflow. The (potential) temperature \( T \) and salinity \( S \) are similarly advected during outflow but are specified using climatological values at inflow grids. Radiation is used for the surface elevation \( \eta \); however, because POM uses a staggered C grid and because transports are specified, the boundary \( \eta \) plays only a minor role and a zero-gradient condition on it also works well. At the sea surface, momentum, heat, and salt fluxes are set equal to zero. To prevent temperature and salinity drift in deep layers in long-term integration, the \( T \) and \( S \) for \( z < -1000 \text{ m} \) are (weakly) restored to annual-mean climatological values with a time scale of 600 days. More details are in Oey et al. (2003) and Oey (2004). The NWAOM is run for 20 yr, and the last 10 yr are then averaged to obtain the boundary values for the second domain, described next.

The second domain (\( 8^\circ-30^\circ \text{N}, 98^\circ-79^\circ \text{W} \); Fig. 1) contains the Gulf of Mexico and a portion of the northwest Caribbean Sea (i.e., the Cayman Sea). It has a doubled-resolution grid (\( \Delta \approx 3-7 \text{ km} \); same 25 sigma levels). It too has only one open boundary in the east (at 79°W) and the boundary specifications are similar to those described earlier for the NWAOM, except that for \( T \) and \( S \) a flow-relaxation scheme with 20 cells for the relaxation zone (Oey and Chen 1992b) is used near the Caribbean (inflow) portion of the boundary. Also, the NWAOM values (10-yr mean transports, \( T \), and \( S \)) are used instead of climatological data. This one-way nesting follows Oey and Zhang (2004). The high-resolution nested domain is then run for 10 yr, also with zero surface fluxes and weak restoring for layers deeper than \( z = -1000 \text{ m} \). During this time, the model Loop Current...
sheds eddies at almost a constant rate of one eddy every 8 months (≈ ± 20 days; cf. Oey 2004). The variables at the end of this 10-yr run are then used as initial conditions for two 5-yr experiments: the NoWind and Wind experiments. The NoWind experiment is merely the extension of the previous 10-yr run. The Wind experiment specifies a uniform westward wind stress \( \tau_{0x} = -0.1 \text{ N m}^{-2} \) roughly following the wind pattern seen in Gutierrez de Velasco and Winant (1996), but without curl, for the region west of 80°W (i.e., for the entire Gulf and the northwestern portion of the Cayman Sea). We have also conducted other experiments with seasonal surface fluxes and/or zero restoring in the deep layers. These experiments do not alter the main conclusions from the base runs of zero surface heat and salt fluxes and weak restoring.

Fig. 2. The 5-yr mean currents (m s\(^{-1}\)) depth averaged in the upper 200 m (a) without and (b) with wind. Color shows the speed, and black curves are the 20-day trajectories based on the currents. The white contour is the 200-m isobath. (c) Zonal velocity difference between Wind and NoWind experiments. Blue and black vectors indicate different vector scales for shelf and open ocean, respectively, separated by the black 200-m isobath. Red shading shows for reference the 5-yr mean Loop Current where speeds are >0.2 m s\(^{-1}\).

3. Main differences in circulation between the Wind and NoWind experiments

To understand the role of wind in the dynamics of the Loop Current and rings, we compare two otherwise identical experiments, but one is without wind and the other one is with a steady and spatially constant westward wind stress \( \tau_{0x} = -0.1 \text{ N m}^{-2} \). Figures 2a,b show the mean currents, depth-averaged in the upper 200 m, for the two experiments. An anticyclonic Gulf-wide circulation is seen in Fig. 2a when there is no wind acting in the Gulf of Mexico. The flow is westward in the southern half of the Gulf, except for a local eddy near (23°N, 94°W). A strong current is seen along the western coast; the current continues clockwise along the outer shelf and shelfbreak (≈200-m isobaths; white contour) of the
northern Gulf of Mexico to the west Florida slope. Currents on the continental shelves ($H < 200$ m) are weak. For the experiment with wind (Fig. 2b), the westward wind drives northward Ekman transport near the surface, downwelling on the northern Gulf’s shelf and upwelling on the southern shelf, and westward shelf transports. Over the west Florida shelf, the flow is north-northwestward. The northern shelf’s westward transport extends from the northern portion of the west Florida shelf westward cyclonically around Brownsville in southwestern Texas, where the shelf narrows ($25^\circ$N, $98^\circ$W). Currents over the southern shelf flow westward to the southern edge of the Bay of Campeche, where the shelf also narrows ($18^\circ$N, $94^\circ$W). Shelf currents are much stronger than those for the NoWind experiment; the maximum reaches 0.2 m s$^{-1}$.

In the western Gulf, some of the shelf transports are forced downward as the shelves narrow and flows converge; profiles of vertical velocity will be discussed later. However, most of the shelf transport is returned eastward in the central Gulf. To indicate this return flow, we plot in Fig. 2c the difference (wind minus no wind) zonal velocity in the western Gulf. There are two zonal branches of returned flows. One (southern) branch originates at approximately ($23^\circ$N, $97^\circ$W) in the northwestern Bay of Campeche and continues east-northeastward to approximately ($25^\circ$N, $90^\circ$W) near the Loop Current. The other (northern) branch originates at approximately ($26^\circ$N, $97^\circ$W), where the shelf narrows near Brownsville (southwest Texas) and continues almost due east along the continental slope of the Texas–Louisiana shelf to approximately ($27^\circ$N, $91^\circ$W). It is interesting that the southern branch agrees with the eastward return current indicated in the 50-m drifter data by DiMarco et al. (2005) as plotted in Fig. 4b of Sturges and Kenyon (2008). We will return to this later. We next calculate the strengths of the various transports and deduce their dynamical consequences.

4. Transport analyses

Figure 3 shows the 5-yr-averaged sea surface height (SSH) and the depth of $18^\circ$C isotherm ($z_{18}$; black contours). The SSH and $z_{18}$ in the Wind experiment are higher in the Gulf (and the Cayman Sea), except on the Yucatan shelf because of wind-driven upwelling there (Fig. 3b). The local anticyclonic eddy near ($23^\circ$N, $94^\circ$W; Fig. 2b) in the western Gulf in the Wind experiment has a higher SSH and deeper $z_{18}$ than the NoWind experiment, indicating that the eddy is larger. A section along 90$^\circ$W (gray line in Fig. 3) is chosen to compute the transport balance in the western Gulf. The transport is integrated from surface to bottom, and it is divided into three parts separated by the 200-m isobath (white contour in Fig. 3). The “middle basin” (MB) is defined as the 90$^\circ$W section with water depths deeper than 200 m in the central Gulf. The “northern shelf” (NSh) and “southern shelf” (SSh) are then defined as the sections with water depths shallower than 200 m north and south of the middle basin, respectively. For the NoWind experiment, westward transports across the SSh and MB transects are balanced by the eastward NSh transport (Fig. 3a). The SSh and MB transports are $-0.09$ and $-0.29$ Sv, respectively, and the NSh transport is 0.38 Sv.
In the Wind experiment, westward transports over the northern and southern shelves are returned eastward in the central Gulf. The transports are \(-0.11\) and \(-0.65\) Sv in the NSh and SSH, respectively, and the MB transport is \(+0.76\) Sv (Fig. 3b). Note that the NSh transport is smaller than the SSH transport because the chosen section coincides with a narrower northern shelf. The important point here is that, comparing Figs. 3a,b, wind reverses the transports across the MB (and the NSh) sections. Is the returned eastward flow across MB predominantly confined to the “upper layer”?

### a. Upper- and lower-layer transports

To answer this and other related questions, as well as to understand the effect of the MB flow to the Loop Current, the volume transport across MB is divided into an upper 800-m layer (upper layer) and a deep layer below 800 m (lower layer). The choice of 800 m is because the Loop Current and rings are mainly confined to the upper 800 m (approximately). The 800-m depth is convenient also because the sill of the Straits of Florida is at that depth. Figures 4a,b show the 5-yr mean transports across 90°W. The dashed line denotes the depth of 800 m. In the NoWind experiment, there is a small westward transport of \(-0.09\) Sv on the southern shelf; the MB transports are also westward, \(-0.25\) and \(-0.04\) Sv in upper and lower layers, respectively (Fig. 4a), and the eastward returned flow (\(+0.38\) Sv) is through the northern shelf (cf. Fig. 3a). In the Wind experiment, the MB transport in the upper layer is \(+0.59\) Sv eastward against the wind (Fig. 4b); the corresponding lower-layer transport is also eastward (\(+0.17\) Sv) and the required westward transport in this case is supplied over the shelves (cf. Fig. 3b). Comparing Figs. 4a,b, it is clear that wind produces a middle-basin transport that is eastward toward the Loop Current. This eastward returned flow over the central Gulf is mostly confined to the upper layer. Could it therefore affect the Loop Current and hence also eddy shedding? The analyses of Pichevin and Nof (1997) and Nof (2005) suggest that it could. These authors show that the shedding of Loop Current eddies depend on the zonal momentum balance (in a control volume that includes the Loop Current, ring, and Florida outflow). Therefore, zonal momentum flux resulting from the returned flow can change the time-dependent behavior of the Loop Current. We will explore this possibility.

### b. Active- and decaying-eddy composites

Instead of temporal means over the entire analysis period (Figs. 4a,b), we now differentiate between two dynamically different states when (i) eddies are newly shed and passing through the 90°W section and (ii) eddies are decaying in the western Gulf and the Loop Current is reforming. We refer to these as “active eddy” and “decaying eddy” states, respectively. All events are used in the analysis. There are 6–7 eddies in 5 yr. We therefore composite transports according to times with and without eddy passages through 90°W (Figs. 4c–f; \(\sim 400\) days with eddies and the remaining \(\sim 1400\) days without eddies). During the active-eddy states, the westward MB transport in the NoWind experiment (Fig. 4c) is \(-1.92\) Sv which is 2 times as strong as the \(-0.93\) Sv for the Wind experiment (Fig. 4d). Because the Wind experiment
has a returned (i.e., eastward) mean flow in the central Gulf in the upper layer (Fig. 4b; also Fig. 4f below), its transport in Fig. 4d is partially cancelled and therefore is weaker than the NoWind experiment. (This does not mean, however, that eddies in the NoWind experiment transport more mass to the west than the Wind experiment, because the composite cannot separate eddy and wind effects; the appropriate calculation will be done in section 4.) Despite the different upper-layer transports, the lower-layer transports for the NoWind and Wind experiments are both eastward = +1.5 Sv. During the active-eddy state, eddies transport mass westward through the middle basin irrespective of the wind, and the compensating eastward flows in the lower layer suggest downwelling in the western basin. In other words, as eddies make their presence felt upon entering the western basin, a portion of the western basin mass is returned in the upper layer, but a substantial portion (+1.5 Sv) is downwelled and returned eastward through the lower layer.

Figures 4e,f show the balances during the decaying-eddy states. In contrast to the active-eddy states of Figs. 4c,d, the MB transports are now reversed: eastward (westward) in the upper (lower) layers. This requires a net upwelling in the western basin, consistent with decaying eddies when warm rings' isopycnals are flattening. The eastward, upper-layer transport in the Wind experiment (Fig. 4e) is 7 times stronger than the NoWind experiment (Fig. 4f). The difference (Wind minus NoWind; +0.82 Sv) in this decaying-eddy state is approximately equal to the Ekman flux \( \tau \, \frac{x}{(\rho_0 f)} \approx 1.5 \, \text{m}^2 \, \text{s}^{-1} \) \( (f = 6.5 \times 10^{-5} \, \text{s}^{-1} \) at 26.5°N) produced by the westward wind stress multiplied by the half zonal length of the Gulf east of 90°W (=500 km).

c. Upper- and lower-layer coupling

To understand the coupling between the lower and upper layers, we apply the composite analyses (active- and decaying-eddy states) to the vertical velocity (Fig. 5). The result is relatively simple in our model because it produces a nearly periodic eddy shedding. During the active-eddy states, there is downwelling in the western Gulf in the upper 1500 m and weaker upwelling in deeper layers before becoming weakly downwelling again near the bottom (Figs. 5a,b). In the eastern Gulf, except very near the surface, the upper 2500 m is upwelling, whereas downwelling exists in deeper layers. During the eddy-decaying state (Figs. 5c,d), the sign of vertical velocity in the western and eastern Gulf is reversed. There is now upwelling in the western Gulf (except very near the surface and bottom; Fig. 5c). At the same time, as the Loop Current expands, there is general downwelling in the eastern Gulf (except near \( z = -500 \) m for the Wind experiment and also near \( z = -1700 \) m for both experiments). Together with Figs. 4c–f, these vertical-velocity composites suggest that, irrespective of wind, the Gulf of Mexico may be idealized as an oscillator forced by the Loop Current and eddy shedding, with a period = the shedding period. The downwelling in the western Gulf and the sense of upper- and lower-layer transports shown in Figs. 4c,d agree with the ideas described in Sturges and Kenyon (2008). However, the cause is very different. Sturges and Kenyon (2008) had in mind the classical Sverdrup–WBC circulation driven by wind stress curl in the western Gulf of Mexico. In the present case, the forcing is due to eddies.

d. Delayed eddy shedding

The results in Fig. 4 show clearly that the Wind experiment produces an upper-layer returned (i.e., eastward) flow in the central Gulf. Nof (2005) shows that a growing eddy detaches when it reaches a size where the westward Rossby wave speed is greater than the growth rate of the Loop Current because of inflow from the Yucatan Channel. The returned flow in the Wind experiment counteracts the westward Rossby wave speed. This tends to force the Loop Current to grow to a larger size before it sheds an eddy, prolonging the shedding period as well as making the eddy larger. To show these effects, we plot in Fig. 6 color SSH at 90°W for the NoWind (Fig. 6a) and Wind (Fig. 6a) experiments. During the 5-yr simulation, there are 7–8 eddies in the NoWind experiment and 6 in the Wind experiment. The delayed shedding in the Wind experiment results in an averaged lengthening of the shedding period of about 54 days.

e. Delayed eddy shedding in reduced-gravity experiments

To confirm the idea that altered x-momentum balance can result in longer shedding periods when a returned eastward flow (or momentum flux) exists, we conduct a number of reduced-gravity model experiments. The reduced-gravity model domain is the same as the north-west Atlantic Ocean domain shown in Fig. 1, except that the coastline is defined at the 200-m isobath, and the eastern boundary at 55°W is closed. A zonal wind stress (with curl) is then specified east of 82°W (i.e., in the Atlantic Ocean only) to drive a gyre that has a transport \( \approx 20 \text{ Sv} \) through the Yucatan Channel. Because the reduced-gravity model does not have topography, it cannot generate “shelf transports,” and the return eastward momentum flux is specified as a body force (kinematic stress \( \tau_{11} = \rho \beta g y \); \( \text{m}^2 \, \text{s}^{-2} \)) in the x-momentum equation, applied over the western portion of the model Loop Current.
only (between 90° and 86°W). Table 1 gives various model parameters and their meanings. Weak dissipative processes are included as $A_H$ for numerical stability and $\alpha_N$ and $C_b$ for dissipating eddies after they propagate into the western Gulf of Mexico. Various experiments were conducted (see later). Each was carried out for 15 yr, but a quasi steady state (when the model Loop Current sheds eddies at a regular period) was achieved in about 4 yr. The chosen $H$ and $g''$ together with the wind stress curl forcing east of 82°W give a Yucatan transport $T_{Ryuc} = 20$ Sv and an eddy-shedding period $P_{shed} = 260$ days [for $\tau^{body} = 0$; see Table 2, reduced-gravity experiment 1 (RGExp.1)]. By varying the strength of the wind stress curl over the Atlantic Ocean, we adjusted $T_{Ryuc}$ and found that, for the range $15$ Sv $\leq T_{Ryuc} \leq 25$ Sv tested, the $P_{shed} = 260$ days is unchanged (cf. Hurlburt and Thompson 1980). We exclude the case of small $T_{Ryuc} \approx 5$ Sv, when the solution is diffusive (i.e., linear, $P_{shed} = \infty$).

Following Pichevin and Nof (1997), a control region $S_{EG}$ is taken enclosing the eastern Gulf and the following is obtained by integrating the steady-state (inviscid) $x$-momentum equation over $S_{EG}$:

---

**Figure 5.** Composite-mean vertical velocity (m day$^{-1}$) for (a),(b) active-eddy state and (c),(d) decaying-eddy state for the (left) western and (right) eastern Gulf of Mexico (west and east are separated by the 90°W longitude) and for the Wind (solid) and NoWind (dashed) experiments. The calculations were done for regions where the water depth is $>200$ m.
FIG. 6. The SSHs (color, m) at 90°W as a function of year and latitude for the (a) NoWind and (b) Wind experiments, showing delayed shedding and stronger eddies for the Wind experiment.

\[
\int \text{Florida} \, hu^2 \, dy \approx \int S_{\text{EG}} \, \tau_{\text{body}} \, dS_{\text{EG}},
\]

which is Pichevin and Nof’s (1997) Eq. (2.3) (without a \(\beta\) term on the lhs: it is an order of magnitude smaller than the \(\int \text{Florida} \, hu^2 \, dy\) term) with an extra term involving \(\tau_{\text{body}}\) on the rhs. Here, \(h\) is the layer depth, \(u\) is the \(x\)-component velocity, and “Florida” means that the integral is across the Straits of Florida (from Cuba to Key West). Without the \(\tau_{\text{body}}\) term, (1) is not balanced: hence Pichevin and Nof’s “momentum imbalance paradox,” meaning that there can be no steady state. Pichevin and Nof (1997) resolved the difficulty by allowing eddies to shed. When \(\tau_{\text{body}} \neq 0\), the westward force due to momentum outflux through the Straits of Florida may be balanced by an eastward \(\tau_{\text{body}} > 0\), which then results in a steady-state (i.e., non-eddy-shedding) Loop Current. However, the required \(\tau_{\text{body}} > 0\) turns out to be unrealistically large. For more realistic values, shedding is delayed (i.e., smaller \(\partial / \partial t\) term), as shown next.

To estimate the equivalence of \(\tau_{\text{body}}\) in the 3D simulation, it is convenient to express \(\tau_{\text{body}}\) in terms of a zonal velocity \(u_{\text{body}}\). The \(u_{\text{body}}\) may then be thought as the influx of an equivalent eastward momentum in the central Gulf in the 3D simulation. In Fig. 2c, this influx would be near 90°W from 25° to 26°N, and it has a mean transport \(\sim 0.05 \text{ m s}^{-1} \times 200 \text{ m} \times 100 \text{ km} \approx 1 \text{ Sv}\), which is approximately equal to westward wind-induced shelf transport. From (1),

\[
u_{\text{body}} \approx (\tau_{\text{body}} x_{\text{body}} / H)^{1/2},
\]  

where \(x_{\text{body}} \approx 100 \text{ km}\) is taken to be approximately the same as the 25°–26°N (near 90°W; Fig. 2c) distance upon which the eastward momentum flux acts.

Table 2 shows \(P_{\text{shed}}\) for various experiments with \(\tau_{\text{body}}\) varied from 0 (RGExp.1) to \(8 \times 10^{-4} \text{ m}^2 \text{ s}^{-2}\) (RGExp.4). These show that \(P_{\text{shed}}\) increases (shedding is delayed) as \(\tau_{\text{body}}\) increases. The \(u_{\text{body}} \approx 0.26 \text{ m s}^{-1}\) (RGExp.3) is approximately the upper limit of that found for the eastward momentum (per unit mass) near 90°W in the Wind experiment just before (~60 days before) an eddy is shed. The resulting lengthened period of 40 days is shorter than but consistent with the lengthened period of about 54 days previously noted for Fig. 6. As another test, RGExp.5 shows that the \(P_{\text{shed}}\) decreases as \(\tau_{\text{body}}\) is reversed \(-2 \times 10^{-4} \text{ m}^2 \text{ s}^{-2}\). In this case, Eq. (1) cannot be balanced because \(\tau_{\text{body}}\) and the lhs term \(\int \text{Florida} \, hu^2 \, dy\) both contribute to the outflux of momentum (from the control region \(S_{\text{EG}}\)) or a westward force, which tends to hasten eddy shedding as Pichevin and Nof’s (1997) theory suggests. As a corollary to the theory, these model experiments suggest that eddy shedding can depend on an externally imposed momentum flux through the Straits of Florida, which is an interesting possibility that may be worth pursuing in a future study.

5. Eddy-propagation tracks and mass balance in the Gulf

Figure 6 also shows that eddies in the Wind experiment are larger and stronger than the NoWind experiment, as

| Table 2. Shedding period \(P_{\text{shed}}\) and Yucatan transport \(T_{\text{Yuc}}\) from the RGExp. |
|---|---|---|---|---|
| RGExp | \(\tau_{\text{body}}\) (m^2 s\(^{-2}\)) | \(u_{\text{body}}\) (m s\(^{-1}\)) | \(P_{\text{shed}}\) (days) | \(T_{\text{Yuc}}\) (Sv) |
| 1 | 0 | 0 | 260 | 20 |
| 2 | \(2 \times 10^{-4}\) | 0.18 | 280 | 19 |
| 3 | \(4 \times 10^{-4}\) | 0.26 | 300 | 19 |
| 4 | \(8 \times 10^{-4}\) | 0.37 | 360 | 18 |
| 5 | \(-2 \times 10^{-4}\) | -0.18 | 245 | 21 |
seen in the higher peaks of the SSH contour in the former. In this section, we explore some of the consequences of the larger-size eddies in the Wind experiment. Figure 7 shows eddy-propagation paths for the experiment with wind (red) and without wind (green) over 5 yr. In the central basin shown (removed from the coast and/or significant topographic slopes), anticyclonic eddies move southwestward because of the $\beta$ (westward) and nonlinear self-advection (southward; Smith and O’Brien 1983). Figure 7 shows clearly that eddies in the Wind experiment tend to move more southward, suggesting a stronger nonlinear self-advection that is consistent with the result discussed in the previous section and Fig. 6 that wind forces a stronger Loop Current and correspondingly stronger rings. Do larger and stronger eddies in the Wind experiment increase the westward transport of mass? Let the (northward only) volume influx into the Gulf through the Yucatan Channel be $Q$ and the corresponding (eastward only) outflux from the Gulf through the Straits of Florida be $q$ ($\leq Q$; Fig. 8). In the model, the $Q$ is dominated (99%) by the WBC (i.e., model Yucatan Current) influx from the Caribbean Sea into the Gulf on the western side of the Yucatan Channel in the upper 800 m (as sketched in Fig. 8; see Oey 2004; Oey et al. 2005). It is in fact the flux that feeds (and grows) the Loop Current. When the Loop Current sheds an eddy, a portion of $Q$, $Q_{\text{eddy}}$, is transported westward. If the eddy propagates into an open ocean, $Q_{\text{eddy}}$ is “lost” (Nof 2005). For the Gulf of Mexico, $Q_{\text{eddy}}$ is returned as outflux from the Gulf into the Atlantic Ocean and the Caribbean Sea, as $Q_f$ and $Q_y$, respectively (Fig. 8). The sill depth at the Straits of Florida is limited to 800 m so that the $Q_f$ consists of the upper-layer transport only.
The $Q_f$ will be called the “eddy outflow” through the Straits of Florida to distinguish it from the main Florida outflow $q$ that is meant to be a continuation of $Q$ (less $Q_{eddy}$; see Fig. 8). The sill depth in the Yucatan Channel is deeper than 2000 m so that $Q_y$ consists of both upper- and lower-layer outflows. The term $Q_y$ will be called the Yucatan outflow. Irrespective of the wind, a WBC is formed along the western continental slope of the Gulf of Mexico after model integration to a statistically equilibrium state (i.e., the volume integrated kinetic and potential energies are quasi steady). This is similar to the wind-forced $\beta$ dispersion and WBC adjustment problem.

**Fig. 8.** Schematic sketches of volume transport pathways (arrows) in the Gulf of Mexico (a) without wind and (b) with wind. Dashed box is the control volume. Contour is 200-m isobath.
(Anderson and Gill 1979), except that the forcing is due to eddies impinging from the east. The cross sections of temperature and northward velocity across the WBC are plotted in Fig. 9a at 24°N (blue line in Fig. 9b). The width and depth of the WBC are 100 km and 250 m, respectively, and the maximum $v$ velocity is 0.47 m s$^{-1}$ near the surface (Fig. 9a). The volume transport integrated from the western coast shows that the WBC maximum transport is 5.4 Sv (Fig. 9c). The predominant return route for the NoWind experiment in the upper layer is along this WBC route; it continues along the northern Gulf and the west Florida slopes into the Straits of Florida and contributes to the $Q_f$ in Fig. 8a.

Under the wind forcing, the westward wind-induced transport over the northern shelf (Figs. 2b,c) weaken $Q_f$, and this alters the transports through the Yucatan Channel and the Straits of Florida. From Fig. 8b in the control volume (black dashed box), we have

$$Q - Q_y = q + Q_f \quad \text{and} \quad (3a)$$

$$Q_{eddy} = Q_y + Q_f = Q - q, \quad (3b)$$

where the last equality in (3b) is from (3a). Assuming that the $Q_{eddy}$ is approximately the same for both the Wind and NoWind experiments (which will be shown to be true later), from (3b), the weakened $Q_f$ for the Wind experiment must imply a correspondingly larger returned outflow $Q_y$ from the Gulf of Mexico to the Caribbean Sea through the Yucatan Channel (Fig. 8b).

Table 3 gives the various 5-yr mean transports. As defined previously, $Q$ is integrated in the Yucatan Channel taking the inflow only (i.e., northward, $v > 0$), whereas $Q_y$ is for the outflow only ($v < 0$). The term $Q_{eddy}$ is calculated for each eddy (later), and the values are then ensemble averaged. In the NoWind experiment, the Yucatan inflow $Q$ is 30.6 Sv and the outflow $Q_y$ is 6.9 Sv. In the Wind experiment, the Yucatan outflow $Q_y$ increases to 9.2 Sv and the inflow $Q$ also increases to 33.2 Sv. We will shortly explain why; for now,

**Table 3. Mean transports (see Fig. 8 for meanings and transport directions).**

<table>
<thead>
<tr>
<th>Transports</th>
<th>No Wind (Sv)</th>
<th>Wind (Sv)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Yucatan inflow $Q$</td>
<td>30.6</td>
<td>33.2</td>
</tr>
<tr>
<td>Yucatan outflow $Q_y$</td>
<td>6.9</td>
<td>9.2</td>
</tr>
<tr>
<td>Westward eddy $Q_{eddy}$</td>
<td>13.1</td>
<td>13.2</td>
</tr>
<tr>
<td>Florida outflow $q = Q - Q_{eddy}$</td>
<td>17.5</td>
<td>20.0</td>
</tr>
<tr>
<td>Florida eddy outflow $Q_f = Q_{eddy} - Q_y$</td>
<td>6.2</td>
<td>4.0</td>
</tr>
</tbody>
</table>
note that the net inflows at Yucatan \((= Q - Q_f)\) in both experiments are very nearly the same (23.7 Sv for no wind versus 24 Sv for wind). These values agree quite well with Sheinbaum et al.'s (2002) observations (see also Candela et al. 2002; Hamilton et al. 2005) but are lower than those reported by Schmitz (1996) and Mooers et al. (2005). To estimate \(Q_{\text{eddy}}\), we first define the edge of an eddy as the SSH = 0.1 m. The depth of the eddy is then set to 7°C isotherm, which closely coincides with speeds under the core of the eddy of about 0.05 m s\(^{-1}\). This gives \(h \approx 770\) m in both the NoWind and Wind experiments. The (ensemble; same later) mean width of the eddy in the Wind experiment is then 235 km, and it is 195 km for the NoWind experiment. The mean eddy-propagating speeds are determined and checked by (i) plotting time and along-track contours (i.e., Hovmöller plot) of SSH along the tracks shown in Fig. 7 and then averaging for all eddies the slope of distance over time and (ii) using a control volume to time the entries and exits of eddies and then computing the speeds. 

The two methods give nearly identical numbers, and the averaged speeds are 7.3 cm s\(^{-1}\) for the Wind experiment and 8.7 cm s\(^{-1}\) for the NoWind experiment. Eddies propagate approximately in the south-westward direction. Although the eddy volume in the Wind experiment is larger than the one in the NoWind experiment, the slower propagating speed (due to the eastward return flow in the central basin; Fig. 4) in the former results in similar transports in both experiments: 13.2 Sv for the Wind experiment and 13.1 Sv for the NoWind experiment. We check the model \(Q_{\text{eddy}}\) against the formula given in Nof and Pichevin (2001; see also Nof 2005). They show that, when an eddy separates from the Loop Current, a fraction of \(Q, 2\alpha Q/(1 + 2\alpha),\) goes into \(Q_{\text{eddy}}\), where \(\alpha\) is the Rossby number \((=|c|/f)\). The \(\alpha\) is estimated from the gradient of the tangential velocity of the newly shed eddy’s inner core (approximately enclosed by the SSH = 0.2 m contour) and take the ensemble mean. For example, for the NoWind experiment, \(\alpha \approx 0.35\), so that the theoretical \(Q_{\text{eddy}}\) is 12.6 Sv in good agreement with the \(Q_{\text{eddy}}\) calculated directly from the numerical model (Table 3). From Eq. (3b), because \(Q_{\text{eddy}}\) is about the same with or without wind (Table 3), the larger Yucatan outflow \(Q_y\) in the Wind experiment therefore produces a smaller eddy outflow \(Q_f\) through the Straits of Florida. In the Wind experiment, the anticyclonic along-shelfbreak current that originates in the western Gulf’s WBC, which flows around the northern Gulf and along the west Florida slope, has been weakened by the along-shelf transport so that \(Q_f\) is also weakened. An interesting conclusion is that, because the net \(Q - Q_y\) through the Yucatan Channel is nearly the same irrespective of the wind, the wind forces a more vigorous exchange of mass between the Gulf of Mexico and the Caribbean Sea (both \(Q\) and \(Q_y\) are larger).

The mass balance results also explain why the composite-mean westward transport in the upper layer of the middle basin at 90°W during eddy passage is weaker in the Wind experiment (cf. Figs. 4c,d). The reason is that, of the two eastward returned transports \(Q_f\) and \(Q_y\), the former is predominantly over the northern slope and outer shelf, whereas the latter is through the middle basin. The wind redistributes the Gulf-wide returned flow such that \(Q_f\) is larger and \(Q_y\) is weaker in the Wind experiment. The corresponding composite-mean westward transport through the middle basin is therefore also reduced.

6. Summary

This paper explores the role of the local wind in the Gulf of Mexico in altering the behavior of the Loop Current and shedding of rings. We force a model (POM) of the Gulf of Mexico with steady transport, temperature, and salinity in the Caribbean Sea to produce a Loop Current that sheds eddies at a nearly constant rate. We compare this NoWind experiment with an otherwise identical experiment forced by a spatially and temporally constant westward wind stress. The main findings are as follows:

1) The wind forces westward transports over the shelves and a returned (i.e., eastward) flow in the central basin toward the Loop Current. The returned flow delays eddy shedding.
2) The eastward returned flow and delayed shedding allow more mass to be accumulated in the Loop Current, which therefore grows larger and sheds a correspondingly larger ring.
3) Influx of eddies into the western Gulf produces a WBC.
4) Wind produces a more efficient mass exchange between the Gulf and the Caribbean Sea, because a larger portion of the (westward) eddy transport is then returned via the central Gulf instead of along the WBC and slopes of the northern Gulf and west Florida; therefore, more mass exits the Gulf via the Yucatan Channel.

The simplicity of the wind and other forcing used in the model (e.g., constant inflow transport from the Caribbean Sea) makes these findings robust.

Future work

Some interesting questions remain. The model WBC is due to rings that force themselves upon the western Gulf; the 3–5-Sv transport (Fig. 9c) is close to Sturges’s


