The Relationship of the Curl of the Local Wind Stress to the Circulation of the Cayman Sea and the Gulf of Mexico

ROBERT L. MOLINARI

Environmental Research Laboratories, Atlantic Oceanographic and Meteorological Laboratories, NOAA, Miami, FL 33149

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ABSTRACT

The curl of the annual mean wind stress is proposed as the forcing mechanism for the anticyclonic gyre observed in the Cayman Sea. A simple wind-driven model is presented to illustrate how a steady-state gyre in the Cayman Sea and another gyre in the western Gulf of Mexico can be spun-up by the wind. The model results also indicate that the exchange of mass between the two basins can be enhanced by the wind field. Temporal changes in the upper layer temperature structure of the Cayman Sea gyre are consistent, qualitatively, with changes predicted by a simple wind-forcing model. The same model does not appear valid in the western Gulf of Mexico.

1. Background

Three large-scale anticyclonic circulation features are indicated on climatological temperature charts of the Cayman Sea and the Gulf of Mexico. Fig. 1, taken from Robinson (1973), shows a basin-wide gyre in the Cayman Sea, the Loop Current which intrudes into the eastern Gulf of Mexico, and a closed gyre in the western Gulf during January and July. These features vary in size and intensity, as inferred from changes in the observed thermal gradients.

Little is known of the annual cycle of the Cayman Sea gyre. This feature is bounded on the south and west by two intense currents, the Cayman Current and the Yucatan Current (Molinari, 1976). The surface speeds of the Yucatan Current vary with a strong annual signal. Highest speeds have been observed in June and lowest in November (Cochrane, 1963).

The intrusion of the Loop Current north into the eastern Gulf of Mexico has, on the average, an annual signal. The maximum intrusion occurs during the summer and the minimum intrusion during the winter (Leipper, 1970; Maul, 1975; Behringer *et al.*, 1977). However, the variability about this mean cycle is large, with maximum intrusions also observed in the winter (Molinari *et al.*, 1977).

Sturges and Blaha (1976) have proposed a semi-annual period for the intensity of the gyre in the western Gulf. Behringer *et al.* (1977) present temperature data which support this proposal.

Variability in both circulation and thermal patterns can be induced by variability in the distribution of the curl of the wind stress. A two-layer ocean with the lower layer at rest can be used to demonstrate the effect of the wind curl on the upper layers of the ocean. White and McCreary (1974), for instance, derive a linear, quasi-geostrophic vorticity equation for the two-layer system which can be written as

$$\frac{\partial}{\partial t}\left(\nabla^2 h - \frac{f^2}{g'D}h\right) + \beta\frac{\partial h}{\partial x} + \frac{\sigma}{D}\nabla^2 h = \frac{-f}{\rho g'D}\frac{\partial \tau_x}{\partial y}, \quad (1)$$

where h is the upper layer thickness, f the Coriolis parameter, g' reduced gravity, D the undisturbed upper layer depth, β the change of Coriolis parameter with latitude, σ the interlayer stress, and τ_x the zonal component of the wind stress vector. It is further assumed that the meridional component of the wind stress is negligible. If the upper layer thickness is decomposed into steady-state $\bar{h}$ and time-varying h' components, the equations for each component are

$$\beta\frac{\partial \bar{h}}{\partial x} + \frac{\sigma}{D}\nabla^2\bar{h} = -\frac{f}{\rho g'D}\frac{\partial \bar{\tau}_x}{\partial y}, \quad (2)$$

$$\underset{(1)}{\frac{\partial}{\partial t}\Bigg(\nabla^2 h'} - \underset{(2)}{\frac{f^2}{g'D}h'\Bigg)} + \underset{(3)}{\beta\frac{\partial h'}{\partial x}} + \underset{(4)}{\frac{\sigma\nabla^2 h'}{D}} = \underset{(5)}{\frac{-f}{\rho g'D}\frac{\partial \tau'_x}{\partial y}}, \quad (3)$$

where $h' = h - \bar{h}$, and $\tau' = \tau - \bar{\tau}$.

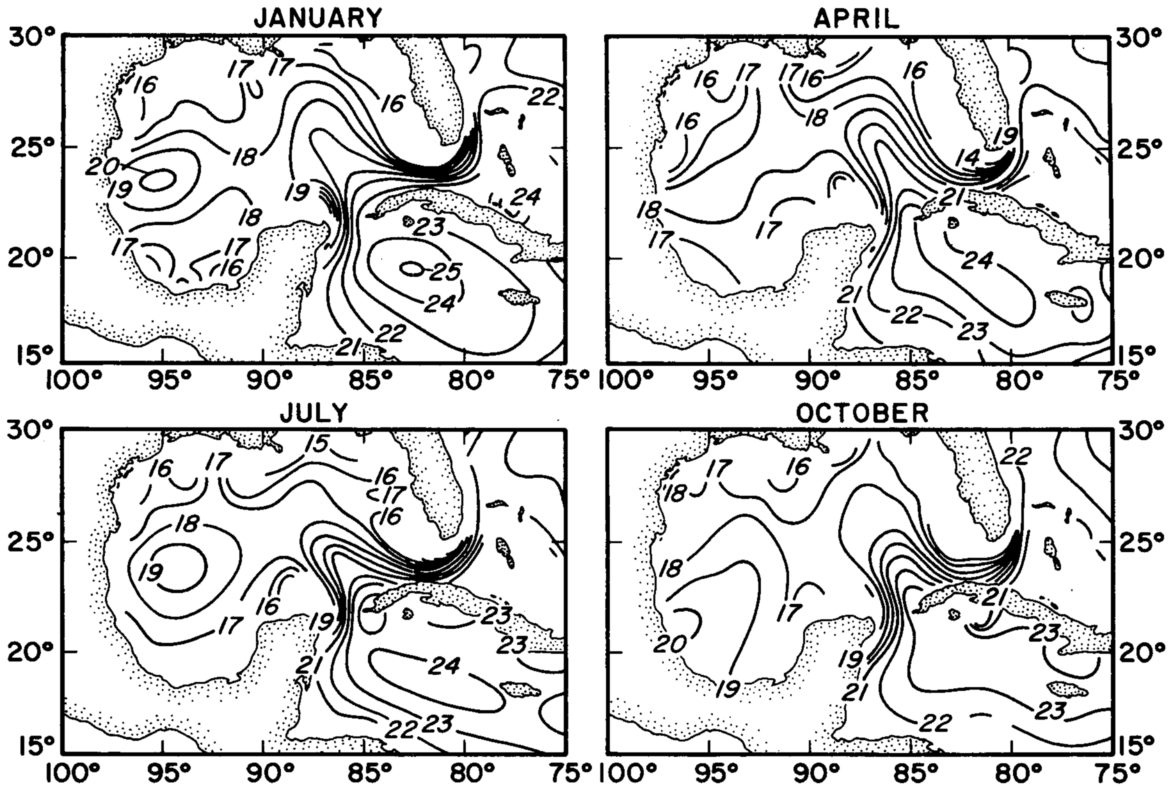

FIG. 1. January, April, July and October mean temperature distributions at 150 m adapted from the charts of Robinson (1973).

Many forms of (3) have been used to study the response of the ocean to time-dependent wind forcing. For instance, Sturges and Blaha (1976), invoking the Sverdrup balance (terms 3 and 5) have proposed that the intensity of the western Gulf gyre should vary with a semiannual period in response to a semiannual forcing by the observed curl field.

Meyers (1975), in an observational study, found that the primary balance affecting the thermal field in the tropical Pacific was between terms 2 and 5. This local effect of the curl of the wind stress on the thermal field has been called Ekman pumping. However, White (1976) found in a theoretical and observational study that term 3 may be important because of the generation of Rossby waves at the eastern boundary of the ocean. The variation of the thermal field in the Pacific is a combination of local Ekman response and Rossby wave response according to White. Finally, Anderson and Gill (1975), in a theoretical study, added term 1 to 2, 3 and 5, and found that in addition to the generation of nondispersive waves at the eastern boundary, there were dispersive short waves generated at the western boundary.

In the present study, the mean annual curl of the wind stress is proposed as the forcing mechanism for the steady-state component of the Cayman Sea and western Gulf of Mexico gyre. Eq. (2) is solved numerically to illustrate how the Cayman Sea and western Gulf gyres may be formed by the wind stress, and how the Cayman gyre may supplement the current which enters the Cayman Sea through the passage separating Jamaica and Honduras and flows into the Gulf of Mexico. Finally, the time-varying component of the thermal field in the two gyres is considered in relation to the temporal variability in the wind stress curl distribution.

2. Data sources

Surface wind stress components over the Gulf of Mexico and Cayman Sea were computed by Mr. A. Bakun of the Pacific Environmental Group of the National Fisheries Service, NOAA, from mean monthly atmospheric pressure data for the period between January 1946 and June 1974. Geostrophic wind speeds were computed from the surface pressure distribution. The wind stress was assumed equal to the wind speed squared times a drag coefficient of 0.0013. Zonal and meridional stress components were determined for each month at 3° square grid spacing.

Monthly and annual wind stress means were computed for each 3° square. The model monthly stresses shown in Fig. 2 were consistent with those of Hellerman (1965), except for the region of the Bay of Campeche. There the 3° squares centered at 18°N include land, and the stresses appear to be high. These values were not included in this study.

3. Steady-state components of the Cayman Sea and western Gulf of Mexico gyres

The simple wind-driven model of Stommel (1948) can be used to show how the permanent gyre in the Cayman Sea and western Gulf of Mexico can be established by the curl of the wind stress. By setting $\psi = (g'/f)\bar{h}$ and assuming a cosine distribution for the wind stress, Eq. (2) becomes Stommel's (1948) vorticity relation

$$\frac{D\beta}{\sigma}\frac{\partial\Psi}{\partial x} + \nabla^2\Psi = \frac{F\pi}{\sigma b}\sin\frac{\pi y}{b}, \qquad (4)$$

where F is the maximum wind stress and b the north-south basin dimension. The numerical relaxation scheme which is used to solve (4) was tested by reproducing Stommel's solution for a rectangular closed basin.

Sets of model solutions were computed using representative parameters for the Gulf of Mexico and Cayman Sea. Two types of forcing were used, the curl of the wind stress, and the specification of the mass transport inflow at the passage separating Jamaica and Honduras and the outflow at the Straits of Florida. Fig. 2 gives the wind stress curl used in the model and the mean annual curl computed from the observed data. The undisturbed upper layer thickness D is taken as 200 m. The frictional coefficient σ is taken as 0.02. The maximum transport in this portion of the water column is arbitrarily taken as 5×10^6 m^3 s^{-1} at the Jamaica-Honduras passage. This transport is approximately one-sixth of the mean annual transport for the entire column observed at the Straits of Florida

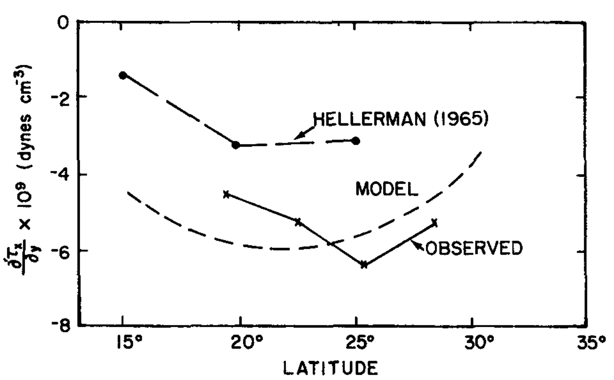

FIG. 2. The curl of the annual mean wind stress, as approximated by $-\partial\bar{\tau}_x/\partial y$, over the Gulf of Mexico and Cayman Sea determined from the 3° wind stress grid discussed in text (solid line) and the 5° wind stress grid given by Hellerman (1965) (dot-dashed line). Also shown is the curl of the wind stress approximation used in the numerical model (dashed line).

TABLE 1. Forcing used in model runs.

Model run	Straits: Jamaica-Honduras	Straits: Yucatan	Straits: Florida	Forcing
A	Closed	Closed	Closed	Wind-stress curl
B	Closed	Open	Closed	Wind-stress curl
C	Open	Open	Open	Mass inflow at Jamaica-Honduras Passage
D	Open	Open	OPen	Wind and mass inflow

(Niiler and Richardson, 1973). The model runs for which steady-state solutions were obtained are listed in Table 1.

The solutions are presented in Fig. 3. The contours represent transport streamfunctions, i.e., ψ multiplied by D. It should be stressed that the dynamics of the currents in the western Cayman Sea and eastern Gulf of Mexico are not adequately represented by Eq. (4). The Yucatan and Loop Currents are inertial flows in which the nonlinear advective terms are significant. However, it is assumed that linear dynamics may be sufficient to describe the steady-state conditions in the central Cayman Sea and western Gulf of Mexico.

The curl of the wind stress establishes two gyres, one in the Cayman Sea and the other in the western Gulf of Mexico (Fig. 3a). As in the Stommel case, the interior balance in both basins is between the β term and wind curl term in (4). A western boundary current provides the necessary return flow to the north. The balance in this frictional boundary layer is between the lateral friction term and the wind curl term. In the case of an open Yucatan Straits (Fig. 3b), there is an exchange of mass between the two basins.

Western intensification caused by the β effect also occurs in the Cayman Sea when the driving mechanism is only mass inflow through the Jamaica-Honduras passage and outflow through the Straits of Florida (Fig. 3c). When both wind stress and flow through drive the system, the western boundary current is a linear sum of the two independent boundary currents. Therefore, in Fig. 3d, the flow through the western section of the Yucatan Straits into the Gulf is intensified, but the total transport into the Gulf remains the same. Other cases (not shown) in which the curl induced transport is greater than the flow through transport, show an exchange of mass between the two basins (as in Fig. 3b).

Sturges and Blaha (1976) computed the Sverdrup transport for the western Gulf gyre by assuming that the feature extended across the entire Gulf, as in the model results of Fig. 3a. However, the presence of the Loop Current in the east complicates the circulation pattern in a manner which cannot be accounted for in the Sverdrup theory.

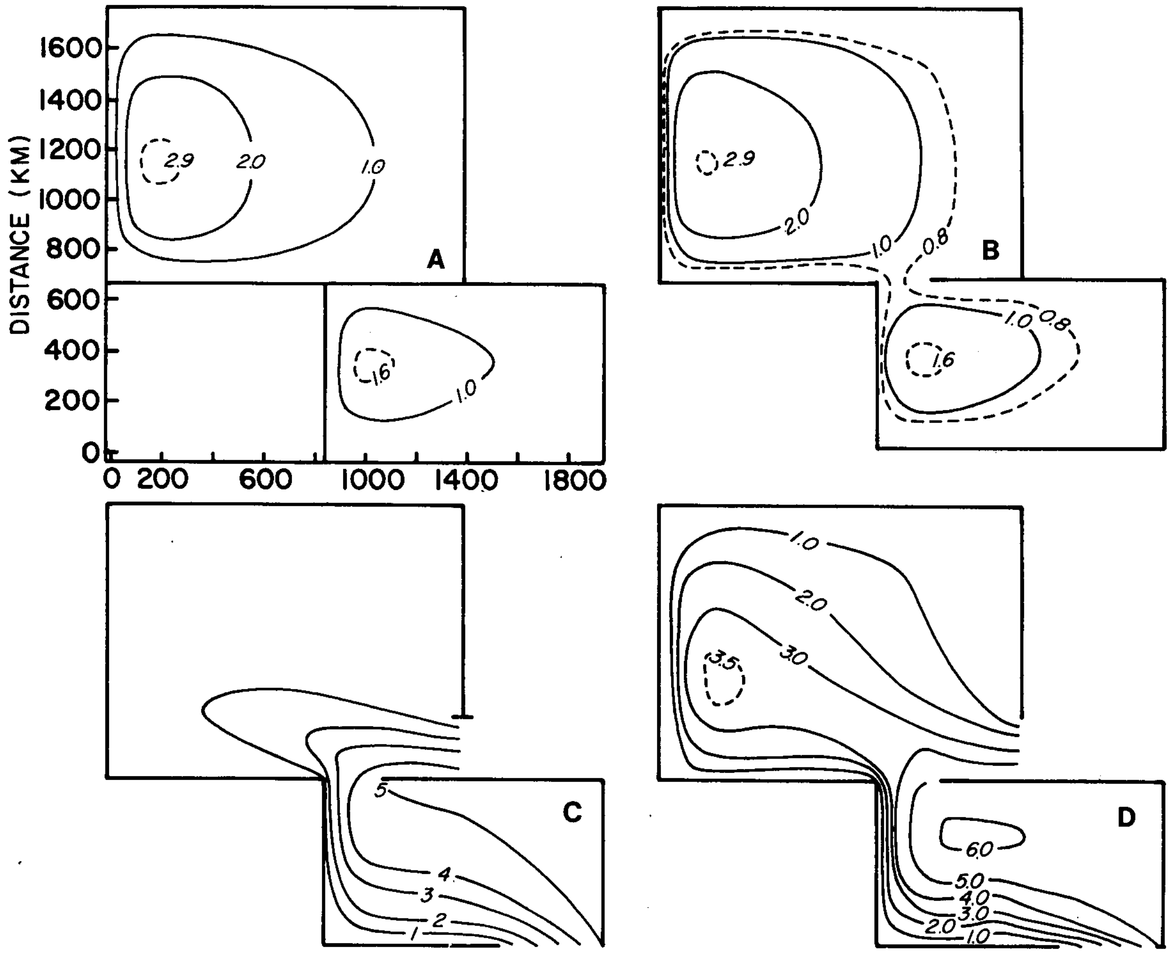

FIG. 3. Transport streamfunction contours (10^6 m^3 s^{-1}) determined from the Stommel model. The insets are lettered to coincide with the model runs listed in Table 1.

Therefore, if the Sverdrup transport is computed from only the wind data in the western Gulf alone, the annual transport for the western Gulf gyre is 3.1×10^6 m^3 s^{-1}. In comparison, Surges and Blaha (1976) computed a summer transport of 3×10^6 m^3 s^{-1} and a winter transport of 6×10^6 m^3 s^{-1}.

4. Time-dependent components of the temperature distributions in the Cayman Sea and western Gulf of Mexico gyres

The seasonal variations of the temperature at 150 m within the Cayman Sea and western Gulf of Mexico gyres (Fig. 4) are taken from the charts of Robinson (1973) as the warmest temperature within the respective gyres at 150 m (except when this isotherm appears to be defined by a single grid point). The time-dependent components of the curl of the wind stress (Fig. 4) are computed from the data of Bakun, described previously. In the Cayman Sea, the curl data represent the average curl along 19.5°N from 87 to 81°W, and in the western Gulf, the average curl along 22.5°N from 96 to 93°W.

The curl of the wind stress has two maxima during the year in both basins. In the Cayman Sea, the maxima (cyclonic curls) are in May and October and the minima (anticyclonic curls) are in January through March and July. These extrema are approximately in phase with periods of greatest temperature change. For instance, the July anticyclonic minima coincides with a temperature increase, while the October cyclonic maxima coincides with a temperature fall, which is consistent with simple Ekman pumping [terms 2 and 5 of relation (3)].

However, the temperature and curl distributions in the western Gulf of Mexico appear 180° out of phase. The maxima in wind stress curl (April–May and September) occur at times of minimum temperature, while minima in wind stress curl (July and December) are coincident with maximum temperatures.

5. Discussion

The results of a simple wind-driven model indicate that the curl of the wind stress could spin-up a gyre in the Cayman Sea. Furthermore, the linear results suggest that the wind-induced gyre would serve to enhance the exchange of mass between the two basins by increasing the transport into the Gulf above that which enters the Cayman Sea from the eastern Caribbean Sea. In this simple model, this additional flow into the Gulf must be returned on the eastern side of the Yucatan Straits. However, in reality, deeper compensating flows are possible.

Although the major currents of the Cayman Sea are nonlinear and time-dependent (Molinari, 1976), a steady-state dynamics might suffice in the interior of the basin to describe the presence of a permanent gyre and its associated transport. Leetmaa *et al.* (1977) have proposed a similar interior dynamics to account for the mean annual transport of the Gulf Stream at Miami.

Sturges and Blaha (1976) argued for a wind-driven gyre in the western Gulf. Other investigators, such as Cochrane (1972), have proposed that the gyre is a remnant of Loop Current eddies which have drifted west after separation from the Loop. A crude energy calculation is presented to show that both mechanisms can provide comparable amounts of energy to the western Gulf.

If Eq. (1) is multiplied by $(g')^2Dh/f^2$ and some algebraic manipulation is performed, an energy equation for the two-layer geostrophic system results. The total energy in this system consists of a kinetic energy component $(u^2 + v^2)D/2$ and an available potential energy component $g'(h - D)^2/2$, where u and v are the geostrophic velocity components and D the undisturbed upper layer depth. The total energy for a circular eddy in this system can be computed by assuming that the distribution of the upper layer thickness h is Gaussian, i.e., $h = D + h_0 \exp(-r^2/2s^2)$, where h_0 is the maximum disturbance of the interface, r radial distance, and s the half-width of the feature. The eddies energy can be computed by multiplying the two energy components by $2\pi r$ and integrating the resulting expression. A detached eddy can be characterized by $D = 200$ m, $h_0 = 100$ m, $s = 100$ km, and $g' = 2.0 \times 10^{-2}$ m s^{-2}. The energy input from one such eddy entering the western Gulf per year is of the order 6×10^{15} J.

The energy input by the wind is given by $(g'h/f)$ curl $\bar{\tau}$, or using the streamfunction defined previously, ψ curl$\bar{\tau}$. From Figs. 3a or 3b, a suitable value for ψD is 2×10^6 m^3 s^{-1}, or for ψ, 10^4 m^2 s^{-1}. A mean curl of 5×10^9 dyn cm^{-3} acting over a $4° \times 4°$ area for one year gives an energy input of about 3×10^{15} J. These crude estimates are sufficiently close so that neither of the mechanisms, wind or detached eddy, can be eliminated as the forcing function for the circulation in the western Gulf.

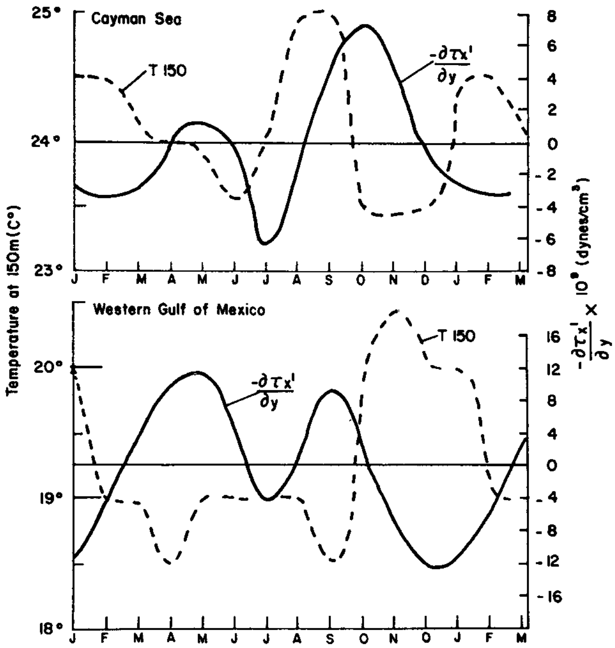

FIG. 4. Mean monthly temperatures at a depth of 150 m in the center of the Cayman Sea and western Gulf of Mexico gyres from the charts of Robinson (1973). Also shown are the mean monthly wind stress curls approximated by $-\partial\tau_x'/\partial y$ (see text for definition of τ_x').

If terms 2 and 5 are dominant in Eq. (3), the upper layers of the ocean will respond locally to the effect of Ekman pumping. In this case, an anticyclonic wind stress curl causes convergence and sinking of the upper isothermal surfaces. Conversely, a cyclonic wind stress causes divergence and rising of these surfaces. This movement of the isotherms is induced by variations in the local curl rather than the mean curl (Meyers, 1975).

Although the thermal response in the Cayman basin appears to be consistent with the dynamics of local Ekman response, the analysis is insufficient to hypothesize a balance between terms 2 and 5 in Eq. (3) to explain this correlation. Meyers (1975) found a similar balance in the Pacific, but White (1976) found that boundary-induced Rossby waves [term 3, Eq. (3)] and local forcing act together to cause the observed thermal response. Since the curl of the wind stress (Fig. 2) has the same structure across the entire length of the Cayman Sea, it is likely that Rossby waves are excited in this basin also. The lack of sufficient ocean data precludes an observational study such as completed by White (1976) in the Pacific.

If the temperature structure in the center of the western Gulf gyre is related to the intensity of the gyre, i.e., the warmer the central temperature, the

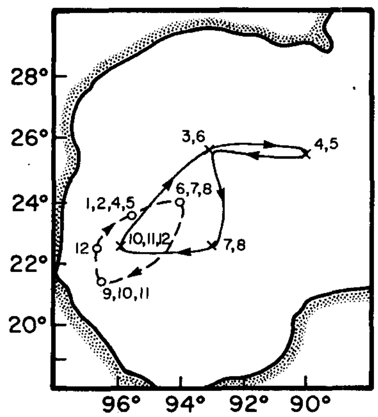

FIG. 5. The migration of the center of the western Gulf gyre (dashed line) determined from Robinson (1973). The month is given by number, i.e., 1 = January, 2 = February, etc. A March position is not plotted because this center position is not consistent with the locations observed in the surrounding months. The positions of the maximum curl of the wind stress are connected by the solid line. September is not plotted because of the small curls throughout the region.

more intense the gyre, then the curl and temperature data presented in Fig. 5 support the contention of Sturges and Blaha (1976). They suggest that the circulation in the western Gulf of Mexico responds almost instantaneously to changes in the intensity of the curl of the wind stress, i.e., terms 3 and 5 are dominant in Eq. (3). Molinari *et al.* (1978) have computed geostrophic transports for the upper 200 m of the southern limb of the western Gulf gyre. In the western Gulf, as in the Pacific, the geostrophic transports are in phase with the wind stress curl. The average geostrophic transport of the upper 200 m layer is 2.2×10^6 m^3 s^{-1}, with maxima in transport in January and August.

As in the case of the Cayman Sea, the exact phase relation between the wind and circulation cannot be determined without further study. A final piece of evidence suggests that such a relation does exist, and warrants further effort. The center of the gyre in the western Gulf of Mexico as determined from the 150 m temperature charts of Robinson (1973) migrates in a clockwise loop (Fig. 5). A position is not given for March because the center is not well-defined during this month. Also shown on this figure are the positions by month of the maximum wind stress curls determined from the $3° \times 3°$ grid of wind data in the region of the gyre. This position migrates in a clockwise loop very similar to the loop made by the gyre center.

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