Reports

A Western Boundary Current in the Gulf of Mexico

Abstract. *The curl of the wind stress over the Gulf of Mexico, during the winter and again in the summer, is similar to that over the central North Atlantic Ocean. An anticyclonic gyre is nearly always found in the western gulf, and we suggest that there is a typical western boundary current, similar in many important respects to the Gulf Stream. The flow appears to be strongest in winter and summer, in phase with the wind curl forcing, and there is evidence that this response is at least partially baroclinic. The deep baroclinic gyre persists when the wind curl vanishes. The winter transport is roughly half that of the Florida Current.*

We have found that many features of the observed density distribution and tidal heights in the Gulf of Mexico indicate that the gulf contains a western boundary current and an interior flow field remarkably similar to the principal midlatitude anticyclonic gyres. Western boundary currents have long been recognized in the major ocean basins and for nearly 30 years have been known to carry the return flow for the interior transport, which (at sufficiently low frequencies) is in balance with the curl of the wind stress. This concept apparently has not been applied generally for smaller oceanic areas. We point out that the Gulf of Mexico is driven by a wind-stress curl similar to that of the central North Atlantic; the available data suggest that the western boundary current, analogous to the Gulf Stream, has a transport of at least 10×10^6 m^3 sec^{-1}. The large gyre in the western Gulf of Mexico is well known; that there is a wind-driven western boundary current seems not to have been recognized.

Winds and basic geometry. Stommel (*1*) has presented maps of the wind stress and its curl over the ocean (his charts 3 and 4). Although the data (Hidaka's) extend only partially into the Gulf of Mexico, it appears that the area of the broad maximum curl in the central North Atlantic (having values of -5×10^{-9} to -10×10^{-9} dyne cm^{-3}) extends into the gulf and the Caribbean Sea. Hellerman's (*2*) data allow seasonal calculations for the central part of the gulf; the maximum, -7.5×10^{-9} dyne cm^{-3}, occurs, during December through February, with a secondary maximum (-3.8) during June through August. The curl vanishes (on the basis of these data) for the other two seasons. As a result, the wind forcing has an interesting periodicity of 6 months.

More recently, maps of wind curl have been shown by Bye *et al.* (*3*); the wind curl given by their calculations are quite consistent with Hellerman's, for our purposes. Earlier, Franceschini (*4*) presented in detail maps of monthly winds over the gulf.

The curl of the wind stress in the Gulf of Mexico, therefore, is comparable with the winds driving the Gulf Stream system for 6 months of the year. The Gulf of Mexico extends about 1500 km east-west, or about one-fourth the width of the Atlantic, hence the integrated interior transport will be reduced accordingly. It extends roughly 1000 km north-south; this dimension should not be limiting, as it is an order of magnitude larger than the width of a typical western boundary current. The Rossby number based on cross-stream scales is unaffected by the geometry in the gulf. The winds and basin geometry thus seem suitable for the development of a western boundary current.

If the width of the basin is r, the total transport in the y direction (M_y) may be estimated in the usual way as

$$M_y \approx \int_0^r \frac{1}{\beta} \operatorname{curl} \vec{\tau}\, dx \qquad (1)$$

where x is to the east and y to the north; β, the latitudinal gradient of coriolis parameter; and τ, the wind stress. For $r = 1500$ km, and $\beta = 2 \times 10^{-13}$ sec^{-1} cm^{-1}, the winter transport given by Eq. 1 is 6×10^6 m^3 sec^{-1}.

Observed features. Nowlin and McLellan (*5*) present maps of geopotential anomaly (dynamic heights) in the Gulf of Mexico; Fig. 1 shows a slightly modified version of their figure 2. In the eastern gulf, the topography is dominated by the Loop Current. Except for

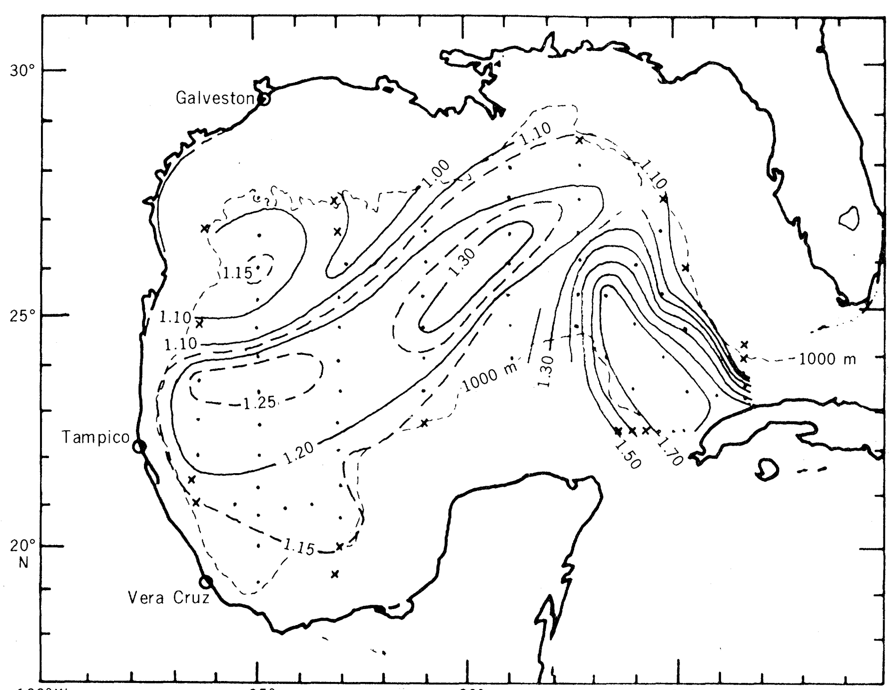

Fig. 1. Sea surface topography relative to the 1000-dbar surface (R.V. *Hidalgo*, February and March 1962) [adapted from (*5*)]. Ambiguous contours have been redrawn, giving the benefit of the doubt to the western gyre. Contours are in meters.

that feature, the pattern appears to us to resemble a basin containing a western boundary current. The area of high sea level toward the west has a relief of 10 to 15 cm. A compilation by Nowlin and McLellan of seven (north-south) sections through this high, using winter data (1958 to 1964), suggests that the usual relief as deduced from the density field is nearer 20 cm. The 1932 and 1935 winter data also show a western gyre, with reliefs of 15 and 38 cm. If the total change in sea level were 20 cm instead of 15, we would expect the transport also to increase, other factors remaining constant. The circulation is discussed further by Nowlin (*6*).

The additional central high (the 1.30 contour in the middle of the basin) (Fig. 1) is not present below 500 m. The basic shape of the large-scale gyre, however, extends below 1500 m; it remains evident in the dynamic heights between 1000 and 1500 decibars [Nowlin and McLellan's figure 21 (*5*)] and weakly evident on the maps of the 1500-dbar surface relative to 2000 dbar. The geostrophic transport has a magnitude of 10×10^6 m^3 sec^{-1} in the main pattern, in reasonable agreement with the (wind curl) computed value of 6×10^6 m^3 sec^{-1}. Whether the central high near 90°W is an extension of the main gyre's high near 95°W, or is an eddylike feature, is not clear.

Velocity distributions presented by Nowlin and McLellan, based on geomagnetic electrokinetograph (GEK) and hydrographic data, show larger shears on the inshore edge of the current, with peak speeds of 70 to 100 cm sec^{-1}.

At about 26°N there is a small cape, at the mouth of the Rio Grande, where the stream leaves the coast and is not constrained by the bottom topography. The analogy with Cape Hatteras is clear. There is another feature, however, in which comparison with the Gulf Stream shows a striking difference. The northeastern edge of the gyre coincides with, and appears to be limited by, the isobaths of the continental shelf after the current has traveled approximately 1500 km downstream. The Gulf Stream, after going this distance beyond Cape Hatteras, has not reached 60°W longitude. The region of the very large meanders appears not to be reached. Other maps suggest that at certain times—mainly in summer—the Loop Current may serve as a boundary on the eastern side of the basin. There thus appears to be control of the flow, either topographic or otherwise, at the stage where the stream might begin to develop increasingly unstable meanders that are characteristic of the far-downstream region of the Gulf Stream.

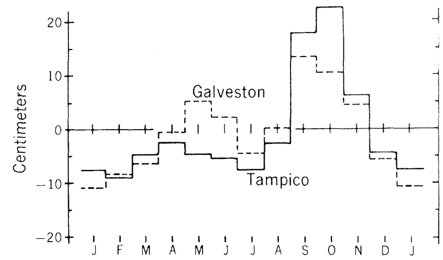

Fig. 2. Annual trend of mean monthly tidal height at Galveston, Texas, and Tampico, Mexico [data from (*7*)]. The stations are adjusted so that the annual means coincide and are corrected to uniform atmospheric pressure. Years of data are: Galveston, 1909 to 1969; Tampico, 1942 to 1950 and 1952 to 1958.

This anticyclonic gyre is well documented as a persistent feature in the historical data. The bathythermographic files have been studied by Whitaker (*7*) and by Robinson (*8*). A series of hydrographic cruises throughout the year is discussed by Vasquez (*9*).

Variability of wind stress and circulation. The available sources of wind data (*2–4*) indicate that the wind curl has a strong 6-month periodicity, being strongest in winter and summer. In response to this forcing, the flow field should also be stronger in winter and in summer if the response time of the system is sufficiently rapid. This variability is apparent in the tide-gauge records. Anticyclonic flow will lower sea level at the coast, compared with the value it would have in the absence of the gyre. Figure 2 shows monthly mean tidal heights at two stations in the western gulf, based on the work of Whitaker (*7*).

There are three separate effects of wind variability in the data shown in Fig. 2; the essential feature connected with wind curl forcing is the abnormal low in summer. First, the usual annual trend at temperate latitudes is a steady rise from spring to fall [see Schroeder and Stommel (*10*)] resulting from seasonal heat storage. Data in Fig. 2 suggest that the magnitude of the summer low, contrasted with a "normal" trend, is about 10 to 15 cm at Galveston.

The July low in sea level is a characteristic feature of every tidal station we have examined around the edge of the Gulf of Mexico—a feature discussed by Whitaker. The amplitude decreases away from the western boundary, being approximately 5 cm at Pensacola and Havana and 1 cm (barely noticeable) along the west Florida coast [see (*11*), figure 7, St. Petersburg].

The second most striking feature of Fig. 2 is the abrupt rise of sea level from August to September. The tidal data at Vera Cruz (not given in Fig. 2) show a trend very similar to that at Tampico. Two other gauges on the Mexican coast, east of Vera Cruz (given by Whitaker), also show this abrupt rise.

Franceschini's (*4*) data (monthly wind stress) show an abrupt change in the wind stress over the entire Gulf of Mexico, from low wind stress (not curl) in summer to higher values in September. We propose that the windstau effect (which implies a balance between the surface wind stress and a horizontal pressure gradient in the direction of the wind) is an adequate mechanism. The slope across the Pacific Ocean at low latitudes is over half a meter [see (*12*)] and about 25 cm across the Atlantic (*13*). The apparent effect (in Fig. 2) seems accentuated by the combination of three factors: the increase in wind stress, the peak in the seasonal cycle of stored heat, and the rise of sea level that accompanies the relaxation of the summer wind (curl)–driven gyre.

There is a third feature in Fig. 2 that results from wind variability. In the spring (May and June), sea level at Galveston is 10 cm higher than that at Tampico, whereas in the fall (September and October) it is lower at Galveston. This effect, again, is caused by direct windstau. In May and June, Franceschini's maps show that the winds are directed nearly toward Galveston. In October and November, however, they blow toward the southwest. A simple calculation of the magnitude of the windstau effect, in which 0.5 dyne cm^{-2} is used and an average upper mixed-layer depth of 50 m is assumed, yields a surface displacement of 10 cm, in accord with the observation.

Response times. The periodic wind forcing raises interesting and important questions about the spin-up and spin-down times of the system. Although the exact mechanism for generating the Somali Current system is in some doubt [see (*14*)], the rapid response of the Somali Current to the onset of the monsoon winds suggests that western boundary currents are capable of developing rapidly. For the Gulf of Mexico this idea is supported by the tidal data of Fig. 2, in which the sea level response (and hence, the flow field) appears to be approximately in phase with the wind forcing, at least to the extent we can deduce from the data we have.

The response of sea level to forcing by local winds is known to be rapid. Smith

(15) shows that on the Oregon coast the initial response time is about an inertial period and is barotropic. The baroclinic density field remains unchanged (near the coast) after rapid wind reversals and sea level changes of ~10 cm.

Although the baroclinic response time of the large-scale central ocean circulation is usually considered in terms of years, it should be remembered that the baroclinic adjustment time near the coast seems sometimes to be in terms of weeks (for example, in upwelling situations). In the annual cycle of transport of the Florida Current, Niiler and Richardson (16) show that the baroclinic and barotropic components are in phase, even though the transport maximum at Miami lags the maximum wind curl by about 6 months. Reid and Mantyla (17), using monthly mean data, have recently shown very good agreement between tidal heights and steric sea level along the U.S. northern Pacific coast. There, the longshore flow reverses direction from summer to winter, yet the baroclinic data agree in phase and amplitude at the annual period with the tidal data (which contain baroclinic plus barotropic information).

When the wind curl is applied to the ocean, the essential mechanism for forcing the baroclinic response (that is, the deepening of the main thermocline) in the center of the ocean is the convergence of the Ekman drift. We suggest that the Gulf of Mexico appears to have a relatively rapid baroclinic spin-up time. Presumably this is because the north-south scale is short and also because the gyre (Fig. 1) fills almost the whole deep basin.

The data appear to be contaminated with high-frequency effects (that is, noisy) but we feel that the evidence is indicative of a baroclinic response that is rapid. The temperature patterns in the western gyre, as shown by Robinson (8), are strongest in winter and summer, but weaken in April and weaken again slightly in September and October. This feature is in agreement with the implications of the tidal signal in Fig. 2. The baroclinic adjustment appears to take place, in the summer, at 100 m and deeper. Whitaker's maps of monthly dynamic height are referred only to 150 dbar, but they suggest that the circulation in April is weaker than in the winter months. The data base, unfortunately, is marginal and is poor in the critical months.

Robinson (8) and Whitaker (7) show the mean vertical temperature variation with depth for average areas in the gulf. These variations show a reversal from the normal seasonal trend in midsummer, at depths of 100 to 150 m, again suggesting baroclinic agreement with the tidal variations.

Our results in one respect are in disagreement with those of Whitaker. He concluded that the tidal heights do not accord with steric level calculations. Unfortunately, for his comparison near Galveston (his figure 32) Whitaker had no hydrographic data below 150 m for the summer months. His results were based mainly on bathythermograph data, taken 100 km from the coast, where the tide gauge was inside a wide shallow shelf. He did find, however, that the inclusion of data nearer the coast (as in the study near Pensacola) improved the comparison significantly. In fact, Reid and Mantyla (17) find that for the Pacific coast small steric changes near the bottom, all across the continental shelf, contribute to changes in nearshore steric sea level and have to be taken into consideration.

Vasquez (9) has studied the western gyre in detail for data in May and June 1971 and in November 1971. He has mapped the topography of four density surfaces for both sets of data. The average change in depth of these density surfaces is 100 m in November, but only 50 m in the spring. This is a weak test, but the difference seems significantly above the noise level.

During the late fall and winter, there is a narrow region near the Mexican border where the winds blow from the north and contribute to an upwelling-like situation. These winds could presumably raise the isotherms on the inshore edge of the flow field and contribute to the rapid spin-up of the gyre.

The implication in some previous studies is that the large anticyclonic gyre in the western Gulf of Mexico is driven by the Loop Current, either directly or by an accumulation of pinched-off eddies. We suggest, instead, that the flow is primarily the result of forcing by the curl of the wind stress. Whether there is any interaction with the Loop Current obviously remains to be investigated. The western gyre is well known in winter, but our suggestion that the currents should increase in the summer in response to the wind curl appears to be essentially a "prediction."

It has been shown (11) that in the eastern gulf, local sea level—and hence the longshore current—is dominated by local longshore winds, at periods of a few days and longer, with coherence becoming low at about a month. Our present results indicate that the coastal currents, away from the Loop Current, will be strongly influenced by the curl of the wind stress over the entire Gulf of Mexico for periods near 6 months.

Another feature apparent in Fig. 1 is the downstream slope of sea level along the inshore edge when the boundary current is constrained by the coast. Sea level falls in the downstream direction. The total change appears to be about 5 cm between 21° to 26°, or less than the slope [2 cm per degree (13)] along the U.S. Atlantic coast. Drift currents reported by Murphy *et al.* (18) are also consistent with the present idea. Their results show a surface drift between the central Gulf of Mexico, near the Loop Current, and the far southern Texas coast.

The most appropriate name for a western boundary current in the Gulf of Mexico would certainly be the Gulf Stream, but this name is already firmly in use elsewhere. An obvious next-best choice would be the Mexican Current, a name which we now propose.

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