Mean Flow in the Gulf of Mexico

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ABSTRACT

Several independent data sources suggest that there is a net upper-layer mass flux of order 3 Sv to the west in the central Gulf of Mexico, even though the western Gulf is a closed basin. A plausible explanation is that this net flux is pumped downward by the convergent wind-driven Ekman pumping, as is typical of all mid-latitude anticyclonic gyres. The downward flux can follow isopycnals to depths of order 500 – 600 m and deeper by eddy mixing; a mechanism for forcing deep water to the south through Yucatan Channel is provided by the intrusion and ring-shedding cycle of the Loop Current. Potential vorticity maps show that a deep flow from the western Gulf back to Yucatan Channel is likely. Reconciling this weak deep flow with the observed mean density field requires that pressure surfaces near ~ 300 m in the central Gulf are level. While almost all numerical models show the observed strong flow to the west in the southern part of the basin, none that we have examined show this net upper-layer flow to the west in the central Gulf.
1. Introduction

The currents in the Gulf of Mexico are dominated by the Loop Current, by the rings that detach from it, and by the myriad other eddies that occur. Although our focus in this paper will be the mean flow, both horizontal and vertical, the eddy motions are dominant and are important to the consideration of how all the pieces fit together. Figure 1 shows a presumably typical pattern of sea surface height in a situation in which an anticyclonic ring has recently separated from the Loop Current and is drifting to the west. In this instantaneous view, rings and eddies dominate the circulation though in most means, whether from numerical models or from hydrographic data, we see a “mean Loop Current” in the east and a persistent anticyclone in the western Gulf.

The winds over the Gulf have a clear annual cycle; the near-shore currents along the western and the northern shelf regions have vigorous annual cycles as well. Although the Loop Current has a great deal of variability, for reasons that are not understood it does not have any significant energy at 12.0 months. These flow patterns have recently been discussed by many authors: see, among others, Ohlmann and Niiler (2005), Leben (2005), Schmitz et al. (2005), DiMarco et al. (2005), and by others on the basis of model results: e.g., Oey et al. (2005), Chassignet et al. (2005). W. Schmitz
has developed a web page devoted to a summary of our knowledge of the Gulf of Mexico\(^1\).

\(^1\) http://www.serf.tamus.edu/gomcirculation/

The availability of essentially continuous data from satellites, both for temperature and for sea-surface height, has given us a new understanding of events at the sea surface. Leben’s (2005) animation of sea surface height is particularly helpful in our attempts to understand the behavior of rings and the interactions between eddies. In contrast to the older view that a single anticyclonic ring existed almost in isolation, it is clear that several eddies of both signs are present at any given time.

This paper deals with several long-term data sets of winds and currents. There is a major concern in all areas of science about climate change. Many papers show data sets beginning in the 1800s (e.g., Mann and Emanuel, 2006). The only such long-term data set on surface currents in the ocean is that from the well-known archives of ship drift. One of our original motivations in this work was not only to understand the implications of the ship-drift data in the Gulf but also to get a better understanding of the errors in that data. We soon realized, however, that we were finding an unexpected
signal that was so much larger than the errors that we were obliged to try to understand the observations. Thus the primary purpose of this paper is to explain the implications of the unexpected finding of a mean flow in the central Gulf.

Further, as we began to appreciate how the different data sets all fit together, it became apparent that a mean vertical flow must be an essential component of the circulation. This result should not be surprising, as the curl of the wind stress over the Gulf is remarkably similar to the curl over the central North Atlantic; on most maps of wind curl, the contours that pass near Bermuda continue into the Gulf of Mexico. The analogous vertical component of flow that feeds the thermocline in the mid-latitude anticyclones of the ocean is well known (e.g., McDowell et al. (1982), Luyten, et al. (1983), and many others), even though it is not a component of circulation in the Gulf that is often discussed. The primary mean circulation in the upper Gulf is anticyclonic; we should not be surprised that the flow is remarkably analogous to the flow in the central N. Atlantic.

Our paper is arranged as follows: we first briefly describe the surface wind field; next, from different sets of data, we describe the observed near-surface flow. Putting these together leads to the typical internal Sverdup flow regime, from which we deduce the required vertical flow. We then
study the distribution of salinity and potential temperature in the deep water; finally we examine the density distribution in detail to estimate the mean N-S dynamic height difference in the central Gulf.

2. Surface Winds

Many analyses of surface winds are available, and useful plots can be found in many forms. The data set usually referred to as “NCEP reanalysis winds” from the National Center for Environmental Protection is widely available. Useful insights can be drawn from the set of monthly mean maps prepared by Rhoades et al. (1989), who compared the computed geostrophic winds with observations at a central Gulf buoy to find the necessary correction factors. The “FSU Winds” (Shriver and O’Brien, 1995) are widely used; the results of Isemer and Hasse (1987) are widely respected as well. Boning et al. (1991) computed the Sverdrup transport in the North Atlantic using the wind-stress values of both the Hellerman-Rosenstein and the Isemer-Hasse climatologies. They found substantially increased transports when using the Isemer-Hasse values.

In the present application (to the Gulf of Mexico), we take the point of view that the surface velocity observations are the basic results to be understood. Therefore we are concerned primarily with determining
whether the calculations of Ekman pumping (or equivalently, the interior Sverdrup transport) are reasonably consistent with the transport from surface velocity measurements. Both the wind stress and the surface velocities have errors that are difficult to quantify, so we assume that it is important primarily to ask whether the long-term mean downward pumping is consistent with the enigmatic mean surface flow into a closed western Gulf.

The basic features of the wind field are, first, the prevailing flow is from east to west and, second, the speeds decrease from south to north. The curl of the wind stress is typically negative except in the southwestern corner of the Gulf (Vasquez et al. 2000). The curl has a substantial annual cycle, with a maximum in the summer and a minimum that almost vanishes in October (e.g., Sturges 1993). The monthly mean maps of Hastenrath and Lamb show that the winds tend to be from the north-east from November through January, while they are from the east or the south-east during the rest of the year; the mean speeds are typically ~ 4 m/s in the south and ~ 2 m/s in the north. To focus on the issue of most importance here, the wind signal can be represented adequately by the simplified view in Figure 2, which shows the N-S distribution of annual mean westward wind speeds in the central Gulf from Isemer and Hasse (1995). This result is consistent with the other long-term atlas compilations, but the speeds are weaker by almost a factor of two
than the observations at the NOAA/NDBC buoy (42001) in the central Gulf. There are various ways to try to explain this discrepancy, but that is not the focus of the work here. Note that Figure 2 shows the essentially linear distribution of mean wind speed, not stress; details are given in Sturges (1993).

The Ekman transport is clearly to the north everywhere, so the flow in the upper layers is convergent across the full width of the basin. As in the North Atlantic south of Bermuda, the Ekman flow is to the north but the Sverdrup interior flow is to the south (see, e.g., Mayer and Weisberg, 1993). The near-surface flow will be at an angle to the right of due west. Price et al. (1987) found that the daily-averaged flow at ~5 m would be roughly 60° to the right; McWilliams and Huckle (2006), however, show that the angle between the surface current and wind increases with increasing wind variability. In the North Atlantic, the northward transport in the Ekman layer flow south of Bermuda runs headlong into the southward flow coming from the northern half of the central gyre. In the Gulf, by contrast, the northward flow runs into the coast of the southern United States. The effect is the same. The convergent flow must be pumped down.

This flow regime is analogous in part to the well-known Sverdrup flow as exemplified by, for example, Figure 8-4 of Cushman-Roisin’s (1994) text. It
is important to remember however that the wind pattern over the Gulf is similar to only the southern third of the wind pattern over the typical open-ocean gyre. The curl of the wind stress never goes to zero, so (in theoretical models) the flow to the north along the western boundary does not leave the coast but extends to the northern boundary, looping back around into the interior. Because the wind stress has such a large annual cycle and because the large Loop-Current rings are so often present, realizations of the ideal solution will be seen only rarely. The interior flow field is also rather like that shown by Warren (1982) in his modeled Indian Ocean. In the middle section of his Figure 18, the source on the eastern boundary can be compared with the (wind-driven) flow in the 20°-23° band here.

3. Surface Currents

Recent observations of surface currents have been made with remotely-tracked drifters. Nowlin et al. (2005) give a full description of the currents on the wide Texas shelf; Ohlmann and Niiler (2005) confirmed that the mean flow along the northern coast of the Gulf south of Texas is to the west and has an annual cycle. DiMarco et al. (2005) analyzed a long-term comprehensive set of near-surface drifter data. They show seasonal means as well as an annual mean over the entire Gulf. Weatherly et al. (2005)
examined a set of drifter data different from those of DiMarco, with values both at the surface and at ~900 m.

The largest available data set, and the one that has by far the longest records, is based on ship drift; the data are available as a CD set from the National Oceanographic Data Center. Figure 3 shows annual mean surface currents based on the most recent compilation of which we are aware, an analysis carried out at the Naval Oceanographic Office, Stennis, MS; copies were distributed widely (Dana Thompson, personal communication, 1988).

A more detailed view of this data set is in the published atlas of the Naval Oceanographic Office (1981). Each data point is from a track of either 12 or 24 hours. It is common practice for U.S. ships to use 12 hours, but many foreign vessels use 24. It is unlikely, therefore, that a ship’s result can place a data point in two (or even three) adjacent boxes.

Thus, for example, the boxes between 23° and 24° N and between 90° and 93° W, which are not along major shipping lanes, have sparse data, a combined total of ~400 observations, and these are independent. A ship’s drift velocity is commonly reported to 0.1 knot (5 cm/s); if we suspect that the error in a single data point is perhaps four times that, the standard error of the mean of 400 observations is reduced to ~ (20 cm/s)/20, or ~1 cm/sec. Most of the individual one-degree boxes have substantially more
observations than this. We therefore suspect that the fraction of the ship’s drift induced by windage on the ship, from Stokes Drift, or perhaps by the action of reflected waves will be a greater source of concern than random errors in the mean values of ship drift.

Several features in Figure 3 can be found in all these current observations. The Loop Current in the eastern Gulf is obvious but not very well resolved. Ship-drift data are Lagrangian results, based on a ship’s track over 12 or 24 hours, so structure at scales of the width of the Loop Current will be smoothed or lost. There is a concentrated flow to the west along the northern coast of Mexico. The flow to the north along the western boundary of the Gulf is strongest in July. Most noticeable is the main point we address here: all the flow in the central Gulf is to the west; where is the return flow? This last, puzzling feature is present in almost all months of all such analyses in the Gulf of Mexico; these same results are seen in the results discussed above and in the work of DeHaan (1998).

Ship-drift data contain a large store of potentially valuable observations of near-surface flow. These data extend throughout many years and are reasonably well distributed across the seasons. Where the signal is strong, Richardson and Walsh (1986) were able to use the ship-draft data very well; see also Arnault (1987). A nice example of analysis comparing ship drifts,
the Ekman-layer velocities, and the geostrophic component is given in Richardson et al. (1992; see especially their Figure 3). Richardson (1997) studied the effects of wind blowing directly on the ship to deduce a correction term that can be applied to reduce that specific error in the data. Richardson found that in the case he studied, the error in leeway was 0.6% of the wind speed for vessels steaming normal to the wind and half that for random orientations. We assume that his result (at similar latitudes) is appropriate to our data here.

Figure 4a shows the mean north-south distribution of westward surface speeds in the central Gulf from ship-drift data. Assuming that the ships are not traveling with the wind abeam, it is straightforward to apply Richardson’s correction term (0.3% of wind speed). It is immediately clear, however, that the mean flow to the west is hardly affected at all by this correction (~0.6 – 1.2 cm/sec) so it has not been applied in this figure. The primary flow to the west in the current along the north coast of Mexico has speeds of order 20 cm/sec; a correction term of order 1 cm/sec is lost in the noise. Because its sign is unambiguous, in general it should be retained.

Figure 4b shows a similar plot of westward surface velocity using the data of the ten-year study from drifters drogued at 50 m, from the work of DiMarco et al. (2005). It is important to realize that this data set is
independent of the ship-drift data, both in method and in time. While in principle the drifters also provide Lagrangian data, their analysis method chooses a smoothed velocity estimate computed from adjacent 6-hour positions that fall into a box one and a half degrees on a side. Thus these values have a greater spatial resolution than do the ship-drift data based on 12- or 24-hour fixes.

Figure 4c shows the number of observations in both data sets. DiMarco et al. chose to report the degrees of freedom of data in the individual bins. Because the decorrelation time scale of the data is ~8 days, the degrees of freedom are determined as the total number of 6-hourly observations divided by 32, the number of observations in an 8-day time window.

Figure 5 is a primary result: it compares the two measures of near-surface velocity by the two methods. The N-S mean for the ship-drift is 12 cm/sec and for the drifters, 7 cm/sec. The two data sources agree remarkably well in the strong flow to the west along the coast of Mexico. Within that region of strong current the two values, when integrated across the flow, are essentially identical. The two means are noticeably in disagreement at 25.5° and 26.5° N. Because large rings that separate from the Loop Current drift through these latitudes, it is reasonable to expect that the resolutions of the two data sets would show a different result. The difference in the mean flow
between the two data sets in Figure 5 in this region is so great, however, that
the difference appears to be real. The drogues, at 50 m, are not in the
surface mixed layer in the summer, so the two methods are measuring
different quantities, but this is true in the south as well. The error introduced
by windage on the ships, according to Richardson’s (1997) algorithm, is ~1
cm/sec, but one suspects that unknown errors are at least this large. In the
band of strong flow in the region 22° -24° N along the Mexican coast, the
ships are preferentially traveling with the wind and current, so the effects of
wind and waves will be smallest there. It is possible that the reflection of
waves from the ships’ hulls could make a significant difference in the central
region, where the ships are traveling at a greater angle to the wind and
waves. This topic will be addressed in a separate paper. This difference in
the N-S mean near-surface flow is thus ~ 4 to 5 cm/sec. Whereas that
difference is troublesome and as yet unexplained, it is not central to the main
point of our work and so will not be pursued further here.

In the northern Gulf (Figure 5) the speeds from ship-drift are much
greater than from the 50-m drifters. The tracks of ships in this region are
primarily along shipping lanes from Yucatan Channel to the Texas coast;
thus because they have come preferentially from the region of the Loop
Current, we can speculate that this difference is symptomatic of a bias
arising from the remnants of strong Loop-Current speeds in a portion of the Lagrangian tracks. And because the ship-drift data in the central latitudes also could contain remnants of Loop-Current velocities, it is possible that the differences there are also evidence of this bias.

Several aspects of the ship-drift data, however, give us confidence that the data have genuine value. First, in the central Gulf at 25° – 26° N, where the large Loop Current Rings drift across to the west, the ship-drift data show an essentially bi-modal flow; roughly half the values are to the east, and roughly half are to the west. There is a small mean value which, to within error bars, is appropriate for the drift of rings.

Second, the strong flow to the west along the northern Mexican coast not only agrees with the drifters, but is consistent with classical ideas of Ekman pumping and interior Sverdrup flow. Third, the velocity values in the strong flow to the west are much larger than the suspected errors in the data or the correction terms suggested by Richardson (1997). Fourth, the concentrated flow along the western boundary is to the north while the winds are out of the east, so it is highly unlikely that errors from wind or wave effects could lead to erroneous results of that kind. Moreover, the surface Ekman flow is much too small to account for the observed ship drift.

Richardson’s (1997) work appears to be the best study to date of the errors in ship drift. Yet one issue remains. His analysis and its clever execution in the N. Atlantic shipping lanes is based on the assumption that the errors in ship drift are caused almost exclusively by the effects of winds blowing on the exposed area of the ship. It is possible however that two effects of waves should be explicitly considered: first, the well-known Stokes drift and, second, the effect of waves being reflected from the ships’ hulls. These effects are included in Richardson’s results, but they are implicitly attributed to wind forces. We therefore merely mention here brief comments about these two issues as they may affect the results of ship-drift data.

*Stokes Drift*  Our attitude toward the effects of Stokes Drift in the present case is a bit ambiguous. In most oceanographic settings we think of the large-scale, near-surface flow as extending much deeper than a few meters, whereas the Stokes Drift is limited to the upper layer in which wave orbital velocities are prevalent. In regions of weak currents, however, the Stokes Drift can be as large as the reported ship drift speeds, so inclusion of the Stokes Drift component can be important in understanding the motion of the ship through the water. Because this effect is usually small and is well
described in the literature, we merely give here the results of our calculations; see, e.g., Kenyon (1969). McWilliams and Restrepo (1999) have accorded the Stokes drift a new level of importance by incorporating it into numerical wind-driven circulation models applied over whole ocean basins. Our conclusion, however, is that under the mean wind conditions of the Gulf of Mexico, the effect of Stokes Drift on a typical ship here is the order of 4 cm/sec.

*Effects of Wave Reflections.* While Stokes Drift is often discussed, a potentially larger effect arises from the reflection of waves from a ship’s hull. We point this out explicitly because the major influence from Stokes Drift in the Gulf of Mexico is largely the result of the longer waves. By contrast, the waves that will be reflected from a hull will be those that have lengths the same order as the depth of the hull. See, e.g., Kenyon (2004), Kenyon and Sheres (2006). Thus the momentum imparted by reflected waves will largely be from shorter waves. The long waves and swell will not be reflected, of course, as the ship merely rides over them. Because the shorter waves are in equilibrium with local wind, while the longer waves tend to be preferentially from swell, it may be possible to distinguish between the two effects. This issue is not central to the main point of this paper and so will be reported separately.
5. **How can there be a net flow into a closed basin?**

The principal result thus far is the finding that, first, there is an observed mean surface flow to the west and, second, that the curl of the wind stress is negative over the central Gulf. In an attempt to explain the curious observation of a mean surface flow into a closed basin, DeHaan (1998) computed the monthly variation of this flow and tried various alternative correction methods. DeHaan made a series of geostrophic calculations that use the ship-drift values (corrected by Richardson’s method) as a surface boundary condition. The mean net flow to the west reaches a maximum of nearly 8 Sv in January, when the winds are strongest and the upper mixed layer is deepest. The mean flow is not significantly different from zero from April through July; the annual mean is ~3 Sv. Although the error bars are large, there is physical plausibility to the seasonal cycle. A mass transport of 3Sv in the boundary current is easily understood; if the surface flow to the west in the latitude band 21° – 24° N decreases from 20 cm/s linearly to zero at 300 m, the resulting mass transport is precisely that amount.

DeHaan tried to explain this net flow into a closed basin by assuming that there are unexplained errors in the ship-drift data. Here, however, it seems an inescapable conclusion that the ship-drift data, when taken together with
the drifter results, suggest that the apparent mean flow requires a rational explanation. This alternative explanation is readily found in the downward mass flux that results from surface convergence of the Ekman flow, as described in the next section.

6. Ekman Convergence and the Sverdrup Interior

The Gulf-specific wind values of Rhodes et al. (1989) allowed Sturges (1993) to determine the curl of the wind stress in the central basin along 24° N. There is a large seasonal variation of -5 to -18 x 10^{-9} dynes/cm^3, with a mean of ~ -10. Using the traditional linear Sverdrup relation and \( \beta = 2 \times 10^{-13} \text{ s}^{-1} \text{ cm}^{-1} \), these values lead to transports (to the south) of -2.5 to -7.5 Sv, with a mean of ~5 Sv.

The Sverdrup transport is approximately the same as the observed net transport to the west, and is no doubt within the uncertainty of both. And of course what happens to the Sverdrup transport is well understood: it is pumped down, and out of the surface mixed layer. In the main anticyclonic gyres of all oceans this feature has been well understood for a long time. We emphasize it here only because we are not aware of a discussion of it previously as it relates to the Gulf of Mexico. The three separate parts of the flow field described here are summarized in Table 1. They are computed independently but agree well within error tolerances.
7. But where does the water go?

The final issue to deal with is the question of how there can be a net downward pumping out of the bottom of the Ekman layer over essentially the whole (closed) western half of the basin. That is, even if there is a mean Sverdup flow, where does it eventually go? In a discussion of this general topic for circulation in the open ocean, the issue would usually be ignored; the water simply “goes away.” Here, however, because the basin is closed to the west, one is curious about the manner of outflow. And we must point out in advance that this is a first attempt at an explanation, and is a less quantitative or conclusive result than we would prefer.

The most straightforward answer is that water is pumped down below the westward-flowing near-surface flow so that it can return at depth and leave the Gulf either via the Straits of Florida or via Yucatan Channel. In order for the near-surface waters to reach sufficient depth to leave below the upper-layer flow, there are many potentially-operative mechanisms: winter surface cooling, flow along steeply-sloping isopycnals, the pumping of the Loop Current intrusion cycle, and the initial downward pumping of the convergent Ekman layer, where momentum and energy are supplied by wind stress.
The idea of a vertical flow, pumped down out of the upper surface layer, is well known in the major ocean basins, although it has not (to our knowledge) been discussed in the Gulf of Mexico. The speed of these vertical flows is difficult to grasp intuitively; because it is only on the order of ~100 m per year, it is clearly buried in the much larger signals of internal and inertial wave motions and horizontal flows. The downward path occurs partly as a tiny addition to horizontal flow and carried in eddies of both signs. The descent of surface water along sloping isopycnals has been studied for decades: e.g., McDowell et al. (1982), Luyten et al. (1983).

**Winter surface cooling.** Although the Gulf of Mexico is not usually thought of as a source of deep-water formation, the cold fronts that sweep off the continent often bring below-freezing (air) temperatures. Temperatures as cold as ~12° – 13° are often found on SST maps after a cold front sweeps across the western Gulf. The sea surface is usually obscured by clouds during frontal passages, however, and the coldest water sinks rapidly.

When examining records of surface water temperature at the buoys offshore of the west Texas coast (National Data Buoy Center, e.g. buoy 42035, 22 nm from Galveston TX) it is not difficult to find temperatures of ~10° in randomly-selected February data; surface temperatures can occasionally reach 8°. Nowlin and Parker (1974) reported a band of water approximately
10°-14° along the coast immediately after the passage of a cold front, which is consistent with the cold bands consistently seen in the SST images.

It is worth remembering that surface cooling is not totally restricted to winter months; Loop Current rings carry to the west a large volume of water whose dominant surface signature is its warm core – which erodes away in all seasons. Furthermore, other mechanisms, discussed next, could initiate a deepening process, with the actual cooling taking place several months later. Niiler and Stevenson (1982) suggest that much of the heat loss from these upper-layer warm waters is by turbulent diffusion that mixes the heat downward. Dewar et al. (2006) have suggested a new mechanism of vertical mixing induced by biological migrations; this mechanism has very broad implications here and elsewhere.

**Steeply-sloping isopycnal surfaces.** In the open Atlantic, winter cooling in the far north brings many deep isotherms to the surface; this is the mechanism by which the thermocline is ventilated. An analogous mechanism in the Gulf of Mexico is provided by the steeply-sloping isotherms that are found in rings and “ring pairs.” In an XBT section across a cyclonic-anticyclonic ring pair, the 14° isotherm rises above 200m and reaches down below 400 m. In strong eddy flows or in the Loop Current, isopycnals slope downward another ~ 200 m. The 10° isotherm in the
northwestern Gulf in winter is found as shallow as 200 m, and at much shallower depths along the left-hand side of the Loop Current or Florida Current. Along the right-hand side of these flows, however, the $10^\circ$ isotherm reaches well below 600 m. Thus it is plausible that the upper waters forced down by Ekman pumping can easily be carried along isopycnals to depths perhaps greater than ~600 m.

*Pumping action of the Loop Current intrusion cycle.* The final mechanism for removing this extra mass flux from the western Gulf, and possibly the most important, is the pumping action of the Loop Current as it goes through a ring-shedding cycle. Bunge et al. (2002) and Sheinbaum et al. (2002) found surprisingly strong deep exchanges between the Gulf and the Caribbean Sea. They found that when the Loop current is advancing to the north, a compensating deep flow to the south is observed well below the depth of the Straits of Florida. During a Loop Current shedding cycle, the southward deep flow had transports greater than ~5 Sv for periods longer than a month. They found a deep mean flow to the south on the Mexican as well as the Cuban side. It is crucial to recall that this “pumping action” of the Loop Current intrusion is largely a one-way process. After the Loop Current has intruded to the north, a ring separates (roughly once a year) so that most of the mass that has intruded remains in the Gulf. The essential
idea here is that mass is added in the upper layers but is removed from the deeper layers, which is the issue to be resolved.

**Comparison with Loop-Current rings.** The previous sections have been a rational if preliminary attempt to deal with getting rid of the extra mass flux to the west and, to a lesser extent, the heat flux. It is most instructive, therefore, to compare these with the mass and heat fluxes contained in the approximately annual shedding of a Loop Current ring. Using typical values (~300 km diameter, ~1000 m depth) one finds that the mass flux associated with a single ring is ~3 Sv! So far as we are aware, there has been no great concern about where this extra mass “goes,” or whether it is reasonable to expect that the extra heat contained in the warm-core rings can be dissipated.

It seems appropriate also to mention the work of Hofmann and Worley (1986) who used an inverse solution to find an estimate of the mean flow in the Gulf. They chose a three-layer scheme in which they assumed there to be no net flow in each layer; their upper-most layer reached down to ~600 m, thereby ruling out by assumption the mean flow from the ship-drift and drifter data that forms the basis of our work here.

**Evidence from water property distributions.** One traditional method of searching for weak flows is to examine water properties on density surfaces. It is well known that the Loop Current brings the high-salinity “Tropical
Water” into the Gulf in the near-surface layers. When rings detach and drift to the west, they carry this high salinity layer into the western Gulf. If this high salinity water is pumped or mixed downward and then carried back to the east by the flow postulated here, the requisite higher salinity might show up on maps of deeper density surfaces. Figure 5a shows the depth, and 5b the salinity, on the 27.4 sigma-θ potential density surface derived from a carefully-constructed long-term mean density data set. At these depths the Loop Current water has the low salinity signature of the Antarctic Intermediate Water (AAIW). In the western Gulf, however, the salinity is higher than the incoming AAIW. This higher salinity can result both from vertical mixing and from downward pumping of higher salinity water that is above this density surface. While this result is based on a collection of averaged data, a similar result, showing higher salinity in the western Gulf at these depths, was found by Nowlin (1972) based on nearly-synoptic data from a single cruise. This evidence is completely consistent with the idea of downward pumping.

So far as we are able to tell, the existing database of typical tracers (oxygen, nitrogen, or isotopes) in the Gulf is not yet adequate to allow their use in a long-term averaged sense.
8. Inferences from potential vorticity

Examining the potential vorticity between two density surfaces allows us to determine whether a deep mean flow is dynamically possible. Figure 7 shows potential vorticity between the sigma-theta surfaces 26.8 and 27.65, at depths of ~250 to 1000 m; the exact choice of these two surfaces does not significantly affect this result. ((For the convenience of the reviewers, Appendix A shows the depths of the 26.8 and 27.65 surfaces as well as potential temperature and salinity on those surfaces. It is not our intention to include this in the published paper however. ))

The contours 6.3 – 7.3 found in the western Gulf are also seen in the Caribbean Sea near and south of Yucatan Channel with some eddy noise in between. Thus, because geostrophic flow takes place along PV contours, it is likely that deep water from the western Gulf does indeed flow to Yucatan Channel and back into the Caribbean.

9. Can we find these results in Dynamic Height?

Most oceanographers familiar with the Gulf will know that the mean surface flow to the west as described here, as well as the deep return flow, is not a feature that is “known” from conventional maps of dynamic height. The reason, of course, is that maps of dynamic height typically assume that a reference surface (such as 1000 db) is level, or, equivalently, that the flow at
those depths is vanishingly small. There is an equally plausible alternative assumption.

If we assume that the surface flow suggested by the data here is correct, and that this flow is restricted to a relatively shallow surface layer, then we must reverse the traditional logic. If the upper flow is restricted to perhaps the upper 300 m, then the 300 db surface in the central Gulf is approximately level and dynamic height maps should be interpreted with that in mind. That is, if we assume that the 300 db surface is level, then we can use dynamic height maps to determine the slope of the 1000 db surface. Admittedly this is not the way it is usually done, but we are using the observed surface current data to determine the procedure. To compute the dynamic height is straightforward, but the maps (300 – 1000 db) have a small signal and large noise. Table 2 shows the values of density at the north and south edges of the basin in the center of the Gulf, interpolated at longitudes ~ 90° – 93°W, at a series of depths as determined from maps similar to Figure 5a, except plotted at high resolution and with fine-scale contours.

There is a net difference of ~ 10 units of sigma-theta at almost every level, with the deeper value at the southern edge of the basin. Thus, density surfaces slope up from south to north. The net dynamic height difference across the basin between 300 db and 1000 db is ~10 cm. (The uncertainty is
~3 cm.) While the deeper data are “noisy,” the persistent difference shown in Table 2 is unmistakable. The conclusion here, therefore, is that IF the 300 db surface is approximately level, then the 1000 db pressure surface must slope down from south to north. The slope of these pressure surfaces means that the deep flow is eastward, providing the return flow necessary to balance the westward near-surface flow. A deep mean velocity of order 1 cm/sec, and a net transport of ~3 Sv is a straightforward result. Because the nearly-constant passage of rings provides much larger velocities, such a result will only emerge in a long-term mean. The alternative is to assume that the deep surfaces are level. This assumption gives ~3 Sv of deep flow to the west, doubling the mass transport problem, which is an untenable result.

10. Discussion

It is important to be aware of the very small values of velocity that are able to support a mass flux out of the western Gulf. The vertical shear in the western Gulf below ~300 m is quite weak, and the observed deep flow to the south in Yucatan Channel penetrates very deep. If we assume that a deep, broad flow leaves the western Gulf over the full N-S extent of the Gulf with a vertical extent of ~1000 m, as discussed in the previous section, the associated average velocity required to provide 3 Sv is only ~0.5 cm/sec.
The early work of Fofonoff (1954) suggests that while strong flows to the west are allowable, strong flows to the east would be unstable. Thus we suspect that a deep flow to carry the transport of ~3 Sv back to the east should be distributed in a wide area in the southern part of the basin and probably not concentrated along the boundary.

Sturges (2005) deduced a deep mean southerly outflow from the Gulf back into the Caribbean Sea at depths below ~1100 m. There is obviously a net upper-level inflow in Yucatan which is more than adequate to supply the mass transport carried to the west. Most of the upper-layer flow, of course, goes out through the Florida Straits, but Sturges concluded that at least ~1 Sv leaves the Gulf in this deep flow in Yucatan. The mooring results in Yucatan show a deep outflow along the western wall (on the Mexican side) below ~500m that is completely consistent with these results.

One other mechanism should perhaps be mentioned, if only for completeness: the loss of surface waters by evaporation. The magnitudes of evaporation minus precipitation over the Gulf are orders of magnitude smaller than the ~3 Sv values described here (e.g., Schmitt, 1998). And the inflow from rivers tends to offset this loss.

The observations of Sheinbaum et al. (2002) from deep moorings showed southward-directed mean flow on the Mexican side in Yucatan at
depths of ~600 m and deeper. The standard deviation is large, however, so flow to the south occurs often at shallower depths. Thus it is abundantly clear that the water pumped down out of the Ekman layer can leave the Gulf through the Yucatan Cannel and flow back to the Caribbean. The possibility that some of the water pumped down leaves the Gulf with the flow through the Straits of Florida is also equally plausible but is not pursued here.

The results here are based heavily on ship-drift data. Most of our early understanding of ocean surface currents came from these observations. So far as we are able to learn, ship-drift data are no longer being collected. Considering the remarkable accuracy available with modern navigation techniques, ship-drift data could provide a continuing major source of surface current data that would nicely complement the data taken from modern methods.

While almost all numerical models show strong flow to the west in the southern part of the basin, none that we have examined show the net upper-layer flow to the west in the central Gulf. Whether this is from imperfect physics and forcing, inadequate resolution, or some other cause, we cannot say.
Acknowledgements.

We are grateful to many colleagues for discussions over many years: in particular, D. Nof, L. Oey, W. Schmitz, D. Sheres, M. Stern, and M. Tsuchiya have been most helpful; the work of C. DeHaan was essential. Steve DiMarco was helpful and generous in sharing the drifter data and Mia Shargel provided admirable editorial help. In particular, we are happy to acknowledge the helpful comments from Phil Richardson over many years. During this work W.S. had support from NSF grant 0326233 and the Minerals Management Service, Cooperative Agreement 1435--1-04-CA32645 for which I am most grateful.
Arnault, Sabine, 1987: Tropical Atlantic geostrophic currents and ship drifts


Naval Oceanographic Office, 1981: Surface Currents: SW North Atlantic


Figure 1. Sea-surface height from satellite altimetry (courtesy R. Leben), Jan 9, 2000, from http://argo.colorado.edu/~realtime/gsfc_gom-real-time_ssh/. The contour interval is 5 cm. The anticyclonic ring centered at ~ 91° W has recently separated from the Loop Current. Data are shown at depths >50m.
Figure 2. Mean wind speeds to the west in the central Gulf of Mexico, 90°–92° W, from Isemer and Hasse (1995). Note that speeds to the west are shown as positive.
Figure 3. Mean annual surface currents in one-degree boxes from ship drift, from an unpublished internal distribution at the Naval Oceanographic Office (see text). A velocity scale value is shown at the top of the figure.
Figure 4a. The N-S distribution of the westerly component of mean ship-drift speeds in the central Gulf of Mexico. The full, black curve is the mean of the three individual longitude bands.
Figure 4b. The N-S distribution of westerly component of mean 50-m drifter speeds (from DiMarco et al.) in the central Gulf of Mexico. The full, black curve is the mean of the three individual bands.
Figure 4c. The N-S distribution of the number of observations of surface currents in the central Gulf of Mexico. For the ship-drift data, the number of observations is given. For the drifters, the degrees of freedom are shown, as by the authors. Note the log scale on the x-axis.
Figure 5. Mean speeds of ship drift in the central Gulf of Mexico compared with the mean speeds from drifters.
Figure 6 a, upper. The depth of the potential density surface where sigma theta equals 27.4. The individual “data points” represent the mean from a cluster of 8-10 near-by hydrographic stations available in the NODC data base after quality control and the removal of outliers. The isobaths shown are 100, 500, and 1000 m.
Figure 6 b, lower. Salinity on the potential density surface of Figure 5 a.
Figure 7. Potential vorticity between the sigma-theta surfaces 26.8 and 27.65. In the western Gulf these surfaces lie at depths of ~250 and 1030 m.
Table 1. Transport Estimates of the three flow components

<table>
<thead>
<tr>
<th>Component</th>
<th>Mean, Sv</th>
<th>Annual cycle, Sv</th>
</tr>
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<tbody>
<tr>
<td>N-S Integrated Flow to west</td>
<td>3 +/- 0.9</td>
<td>1 – 8</td>
</tr>
<tr>
<td>Ekman pumping</td>
<td>5</td>
<td>2.5 – 7.5</td>
</tr>
<tr>
<td>Western Bndry Current</td>
<td>4.5</td>
<td>2.5 - 9</td>
</tr>
</tbody>
</table>

1 From DeHaan 1998
2 From Sturges 1993
Table 2. Depths of potential density surfaces at the northern and southern limits of the central Gulf of Mexico interpolated at longitudes ~ 90° – 93°W.

<table>
<thead>
<tr>
<th>Potential Density</th>
<th>Depth, N</th>
<th>Depth, S</th>
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<tbody>
<tr>
<td>26.8</td>
<td>230 m</td>
<td>250 m</td>
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<tr>
<td>26.9</td>
<td>270</td>
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<td>32.27</td>
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<tr>
<td>36.8</td>
<td>1240</td>
<td>1275</td>
</tr>
</tbody>
</table>
Figure Captions  Sturges and Kenyon Mean Flow in the Gulf of Mexico

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