A Puzzling Disagreement between Observations and Numerical Models in the Central Gulf of Mexico

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ABSTRACT

Two large, independent sets of direct observations in the central Gulf of Mexico show a mean near-surface flow of ~10 cm s^{-1} to the west, concentrated in the northern and southern Gulf. Numerical models that the authors have examined do not produce this mean westward flow. The observed speeds appear to be almost an order of magnitude larger than the estimated errors; this paper studies the observations to estimate carefully the possible errors involved and compares the observations with model results. The flow to the west in the southern Gulf is presumably wind driven on the shallow parts of the shelf, and, in slightly deeper water at the outer edges of the shelf, is possibly the result of southward Sverdrup interior flow driven by the negative curl of the wind stress. In another possibly related issue, long-term deep current-meter observations in the northern Gulf at ~1000 m and below find flow to the west, whereas some models find flow to the east. The flow proposed here assumes a mean flow to the west above roughly 300 m, with a required return flow in deep water. The difference between the deep observations and the models will produce a slope of pressure surfaces of the opposite sign below 1000 m, reversing the direction of upper-layer geostrophic flow in the models.

1. Introduction

When a set of observations gives a surprising result, the gold standard in science is to be able to reproduce the observations independently to see whether the result can be duplicated. A second method of checking the reliability of unexpected results is to perform careful, independent analyses of the errors of the observations. When both of these checks suggest that the original result is valid, we assume that the answer can be trusted.

Two different, independent long-term sets of observations in the central Gulf of Mexico suggest that there is a mean surface flow to the west on the order of 10 cm s^{-1}. Most of this westward transport occurs on the continental shelves rather than in the central deep basin [although the large variability in the central Gulf resulting from Loop Current (LC) rings makes the uncertainty of a mean calculation problematic]. Because the western Gulf is closed, there is an obvious need to understand the implications of this mean flow and the required deep return flow. We were motivated to do this study because, although these seemingly trustworthy observations had been reported several years ago, we were puzzled that a casual look at recent numerical model results suggested that models did not find this mean flow.

We focus here on a 1/25° assimilated configuration of the Gulf of Mexico from the Hybrid Coordinate Ocean Model (HYCOM), and we have communicated informally with several colleagues who run different versions of modern numerical models. The observed near-surface mean flow is typically not found in the models. The details, of course, differ from model to model, but the lack of mean flow is common.

The paper first discusses the two major sources of near-surface observations, paying careful attention to the effects of possible errors in both sets of observations. We then compare these observations with model results. In the final section, we examine the mean flow below ~1000 m in the northern Gulf to explore a possible explanation for the discrepancies at the sea surface.

Two problematic issues should be mentioned. First, the Loop Current in the Gulf of Mexico sheds large anticyclonic rings in an irregular manner every ~6–14 months. These rings drift through the region studied

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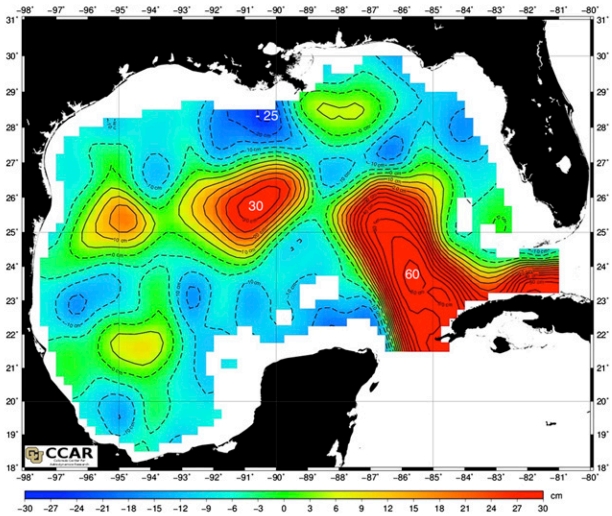

FIG. 1. A snapshot of SSH in the Gulf of Mexico from the work of R. Leben. The dashed contours show negative contours, beginning with zero; contour interval is 5 cm. Several local max and min values are indicated. SSH is not shown at depths less than 30 m.

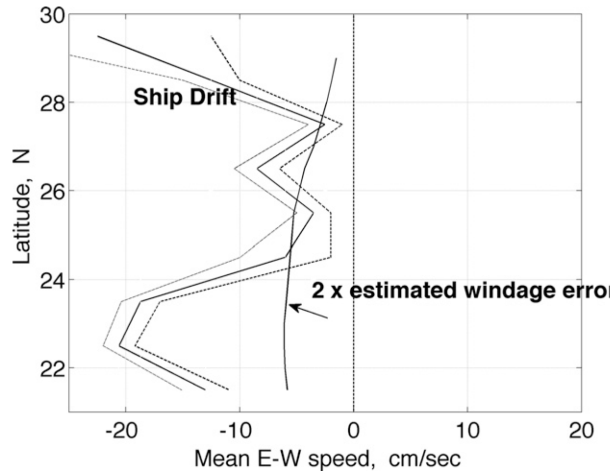

FIG. 2. Observed ship drift in the central Gulf of Mexico, with an error estimate (doubled) based on mean wind speed. The means are in 1° boxes from data along 90°, 91°, and 92°W. The dashed curves show the variability among the three longitudes.

here and create a significant problem in the signal-to-noise ratio. We address this issue with the observations. Second, the datasets and the model times are not simultaneous. The datasets are long, however, compared with the time scale imposed by ring separations, so we assume that the mean values are indeed comparable.

Because the believability of the observations is a central issue here, we discuss the observations and their errors in more detail than what might be considered normal.

2. Observations from ship drift

Sturges and Kenyon (2008) examined the mean flow in the central Gulf of Mexico. One of the main issues in that paper was to determine whether the effects of waves striking against ships' hulls could be responsible for errors in the archive of ship-drift observations. The effect of Stokes' drift and of the reflection of waves from hulls was investigated thoroughly and found to be much smaller than the observed mean speeds determined from ship drift.

Figure 1 shows an example of sea surface height (SSH) of the Gulf, showing a Loop Current ring shortly after it has separated. The snapshot is at an essentially arbitrary time but chosen to show the typical position as a ring drifts to the west.

Figure 2 shows the observations from ship drift averaged from three 1°-latitude bands, 90°–92°W, with a doubled estimate of the maximum leeway effect of winds (blowing normal to the ships' hulls) based on the work of Richardson (1997). (The data are reported at 90.5°, 91.5°, etc. and our notation here drops the half-degree for brevity.) The dashed curves show the range of results from the three longitudes. Richardson did a clever study of the windage effect and found that the maximum error, or leeway, was 0.6% of the wind speed when the winds were directly normal to the ships' tracks. The error estimate shown in Fig. 2 is twice his estimate, and as will be shown below, is probably an overestimate by a factor of ~4. The observed speeds are ~10 cm s^{-1} larger than the errors estimated from winds except in the central region where Loop Current rings drift through. The bulk of the flow to the west is observed in the northern and southern parts of the basin.

At most latitudes in the central Gulf of Mexico, there are thousands of individual ship-drift observations in each 1° box. In the 22°–24°N region there are only a few hundred per 1° square, but at each of three longitudes, leading to nearly a thousand observations per 1° box in the mean. The individual observations are based on ships' positions over the course of either 12 or 24 h and are reported to 0.1 knot (kt; 1 kt = 51 cm s^{-1}) or 5 cm s^{-1}. The archived values are given at 1° spacing. If a ship reports its ''ship-drift speeds'' over 12 h, a 10-kt (older) ship will have traveled ~120 miles, so the reported value is an average over ~2° N–S. Under the assumption that the mean positions are approximately random, there will be as many tracks in the westward-flowing half of the ring as in the eastward return flow. The westward drift of rings is known to be ~5 cm s^{-1}, which is consistent with the results in the central region in Fig. 2. The actual tracks of the rings wobble to the north and to the south, so the mean values do not find the high speeds observed in a single ring.

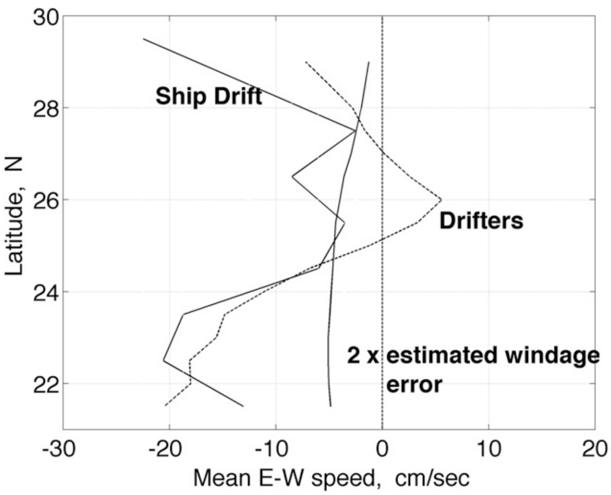

FIG. 3. Observed mean westward speeds from 50-m drifters (after DiMarco et al. 2005), showing the mean from 90° to 92°W. Data points are every half degree, N–S. The ship-drift observations are repeated from Fig. 2.

The accuracy of the ship-drift speeds is uncertain; if they are in error by as much as 0.5 kt (an extreme value) the standard error of the mean, from order 900 observations, is reduced by the square root of 900, or 0.017 kt; less than 1 cm s^{-1}. Thus the accuracy is not in question unless there is some nonrandom error, such as a wind bias, and that now seems unlikely.

In recent years, oceanographers have tended to assign little value to the older ship-drift data, possibly as a result of the higher resolution and accuracy of modern observations as well as the uncertainty of the windage error. In both the northern and the southern Gulf, however, the ship-drift speeds are so much larger than the error estimates that neglecting these observations would be ignoring a valuable data source.

One final point: the great majority of the ship tracks in the central Gulf are from merchant ships traveling through the Yucatan Channel to ports on the Texas coast. There are approximately as many tracks from south to north as from north to south, eliminating one possible source of bias (e.g., from wind being ahead or astern) in the data. Most important, however, the ships' tracks tend to be at ~45° to the prevailing winds out of the east. Richardson (1997) found that for winds blowing at random angles other than normal to the ship, the errors are less by about one-half. Thus, the error estimate shown in Fig. 2 may be an overestimate by a factor of as large as four.

3. Newer observations from drifters

A newer set of observations, based on satellite-tracked near-surface drifters drogued nominally at ~50 m, has been analyzed by DiMarco et al. (2005). Their results for the means in the central Gulf are shown in Fig. 3, with the results from Fig. 2 repeated.

The drifter observations have from ~100 to 500 degrees of freedom for the means at each of the three longitudes, using data over 10 years. DiMarco et al. based the degrees of freedom on the number of observations divided by 32, the time scale of the drifter variability, for 6-h fixes. Most of the drifters were launched by Horizon Marine, Inc. (Marion, Massachusetts) using the "Far Horizon Drifter." The 6-h positions of drifter data were spline fit, first, and the inertial motions were removed; the velocity was obtained from three successive (filtered) positions. The N–S resolution of the archived data is 0.5° and are plotted at each half-degree in the N–S.

DiMarco et al. (2005) show variance ellipses associated with the means in their Fig. 5. The variances are on the order of 100 (cm s^{-1})^2 and are slightly larger in the N–S direction than E–W. As before, given the total number of observations at the three longitudes, the standard error of the mean will be, at worst, on the order of 0.3 cm s^{-1} [i.e., 10 cm s^{-1} (30)$^{-1}$] and usually less.

Windage errors of near-surface buoys have been studied extensively: for example, Pérez-Brunius et al. (2012), and Poulain et al. (2009). The errors from wind blowing on the surface component of modern drogued drifters are small compared with the speeds shown in Fig. 3. Specifically, the slip error U in the downwind direction is found to be modeled well by the relation

$$U = aW/R, \tag{1}$$

where W is wind speed, R is the ratio of drag area of the drogue to the surface buoy, and a is determined experimentally. For winds less than 10 m s^{-1}, a is found to be 0.05 ± 0.01. The "Far Horizon Drifters" have a ratio of drogue to surface buoy area of ~11 when the parachute is fully open—the usual case under windy conditions. Using these values in (1), the error of the drifter speeds in Fig. 3 is 0.005 times the wind speed; in other words, essentially the same or less than for ship-drift errors. In fact Poulain et al. (2009) found that even when the underwater drogue was lost from the drifters in the Surface Velocity Program, the downwind drift was only 1% of the wind speed. Thus, we conclude that a windage error of ~0.5% of the wind speed is a valid error estimate for the conditions here.

Pérez-Brunius et al. (2012) used the same type of drifters as reported by DiMarco et al. (2005). Their extensive analysis found that the "50-m drogues" tended to be at a depth of ~35 m as a result of tension on the cable. An extremely important result from their work is that the motions of their drifters were correlated at only 0.2 with the winds. Thus, the drifter motions were dominated by currents and gave little evidence of being influenced directly by the wind.

One possible objection to the drifter results is that they could be biased by being launched preferentially inside Loop Current rings. The reason we conclude that there is no significant bias, for the application here, is that the positions of the drifters are known with high precision. They measure the ring speeds when they are in rings, but they are ejected and continue to have long durations providing data outside the rings. The data in the rings are well distinguished from those farther from ring paths. In the central Gulf, the higher resolution of the drifters allows them to show remnants of the higher ring speeds as Loop Current rings drift through. Figure 3 shows that the bulk of the mean flow to the west is confined to the far north and far south of the Gulf, over shallow water, with the flow in the central region (~25°–27°N) smaller than errors and uncertainty.

Although the two curves of Fig. 3 seem quite different in the central region where rings pass through, the differences can result from differences in the sampling techniques and may not be significant.

The N–S mean of ship drift is 12 cm s^{-1} to the west; the error from wind is typically ~2–5 cm s^{-1}, depending on how we apply the Richardson (1997) results. Although there are nine different latitudes in the mean, the wind effect could be a systematic bias and so we choose not to reduce the estimated standard error of the mean by the square root of the number of latitude bands. The N–S mean for the drifters is 7 cm s^{-1}s to the west. Because there is evidence that the drifter speeds are not well correlated with local winds (e.g., Pérez-Brunius et al. 2012) the error assigned to the drifter mean may be smaller than for the ship-drift results. For typical winds of 5 m s^{-1}, 0.5% error would be only 2.5 cm s^{-1}, and to the extent that they are uncorrelated, perhaps even less. If we combine the two means and uncertainties from both sets of observations, the result is a westward mean speed of ~9 ± 2.5 cm s^{-1}.

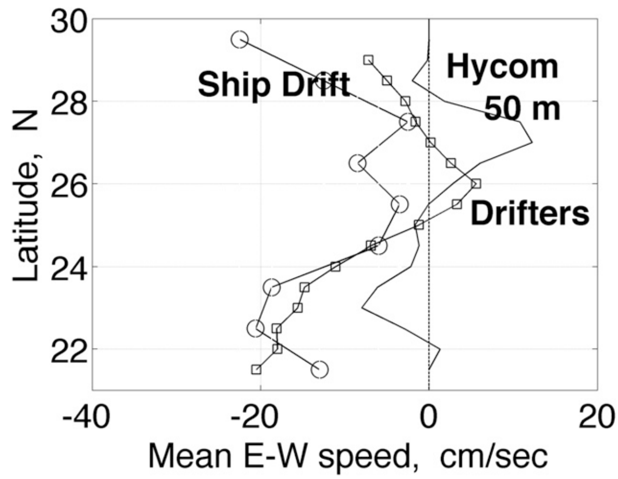

FIG. 4. Results from Figs. 1 and 2 with the mean upper-layer flow from 90°–92°W in HYCOM reinterpolated on a 0.5° grid.

4. Model results and comparisons

a. HYCOM

The model data are extracted from a 1/25° (~5 km) HYCOM (http://www.hycom.org)-assimilated configuration of the Gulf of Mexico running from January 2003 to June 2010. The vertical discretization of the 20 layers of this configuration combines pressure coordinates at the surface, isopycnic coordinates in the stratified open ocean (sigma 0), and terrain-following coordinates over shallow coastal regions (Bleck 2002; Chassignet et al. 2003; Halliwell 2004). The model is forced by 0.5° Navy Operational Global Atmospheric Prediction System (NOGAPS) atmospheric fields. The Navy Coupled Ocean Data Assimilation (NCODA; Cummings 2005) system is used to assimilate satellite and in situ sea surface temperature (SST), SSH, available in situ vertical profiles of temperature and salinity from XBTs, ARGO floats, and moored buoys.

Figure 4 shows both the observational results together with the mean E–W speeds from 90° to 92°W from the HYCOM results averaged over the upper 50 m. The standard deviation of the individual values among the three longitude lines in HYCOM is typically ~5 cm s^{-1}. An estimate of the uncertainty of the model values, based on the standard deviations divided by the number of uncorrelated values, from ~7 years of output, is on the order of 1 cm s^{-1}. The N–S mean from HYCOM is essentially zero. Because the observed values from 25° to 27°N are probably within some reasonable noise level, we focus here on the results south and north of the central region.

b. Other models

The mean E–W speeds from the Princeton Ocean Model (POM) at 90°W, using the model configuration and outputs described in Chang and Oey (2012), Xu and Oey (2011), and Chang et al. (2011), are shown in Fig. 5. This configuration uses the National Centers for Environmental Prediction (NCEP) blended wind product at 0.25° wind forcing. These plots are for a single longitude, not for the mean of 90°–92°W. The POM results are compared with the HYCOM output at the same longitude but not for exactly the same time span. Despite

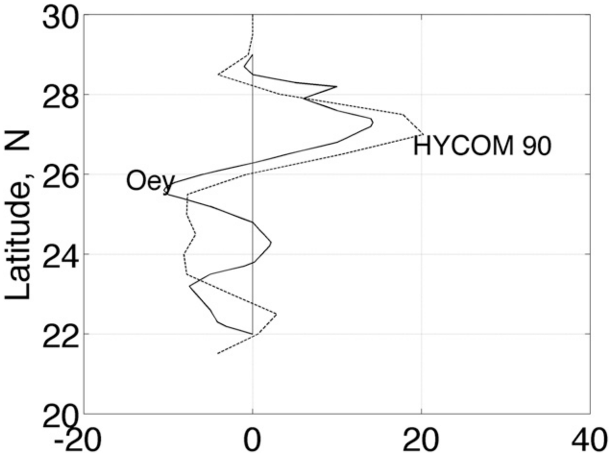

FIG. 5. Mean near-surface speeds (cm s^{-1}) at 90°W from an 8-yr run of POM (courtesy Leo Oey) compared with the HYCOM mean over the upper 50 m.

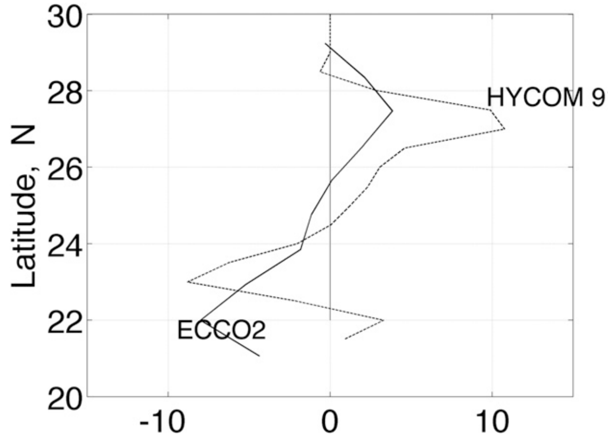

FIG. 6. Upper 50 m–mean E–W flow (cm s^{-1}) at 91°W from a 16-yr run of ECCO2 compared with HYCOM.

some discrepancies in the small scale, the general shape of the N–S currents is similar in both models; the N–S means both are negligibly small. Chang and Oey (2012) have shown that POM has near-surface flow to the west in the southern part of the basin, as in Figs. 2–4 here, but not in the north. Their flow to the west is only about one-half of the speed of the drifters at 23°N, which is similar (for speed) to the flow to the east found in HYCOM, although at different latitudes.

A 16-yr N–S mean of the upper 50-m speeds from the Estimating the Circulation and Climate of the Ocean, phase 2 (ECCO2) model (Menemenlis et al. 2008) is compared with the HYCOM results at 91°W in Fig. 6. As with POM, ECCO2 and HYCOM speeds at 91°W present negligibly small N–S means. The ECCO2 model results shown here are from a full-ocean calculation at 1° in an eddy-permitting version. There is a small region of flow to the west in the far south, with speeds similar to HYCOM but almost no flow in the north as in POM. The N–S mean flow over the full extent of the basin, however, is essentially zero.

Because these independent model results both disagree so similarly with the observations, it makes one reluctant to conclude immediately that the models must be wrong and the observations correct.

5. So what are the believable differences?

The difference between the ship-drift results and the 50-m drifters at ~26°N can be attributed to several factors: to differences in horizontal resolution of the measurements, as well as to differences in the times over which the measurements were made; and, no doubt, to actual error. The scatter between the two sets of observations at 29°N is larger than one would like. In the northern Gulf at 90°–92°W, however, there is a substantial variation N–S position of the boundary of the ocean, so the scatter in the ship-drift data can possibly be attributed to real differences in exactly where the flow was observed. Ohlmann and Niiler (2005), using near-surface Surface Current and Lagrangian Drift Program (SCULP) drifters, found a strong coastal flow to the west in this region. The strongest flow to the west on the continental shelf was observed in a single event that occurred during about two years of drifter observations. We assume (without proof) that the means over ten years (drifters) and many decades (ship drift) suppress the extremes that arise from short period fluctuations.

The largest differences between the model and the observations are near 23° and 27°N; the two peaks of flow in HYCOM differ by >10 cm s^{-1} from the observations in both regions and would seem to be well above the observational error. In both regions the model has stronger flow to the east.

6. Sverdrup transport

The strong flow to the west near 23°N in both sets of observations (Fig. 4) and in the POM appears to be a real feature of the flow. On the inshore segment of the Mexican shelf, the alongshore winds will force flows to the west. For the outer half of a wide shelf, however, it is often found that local alongshore winds are not the primary forcing mechanism. We therefore offer a suggestion for a possible alternative mechanism.

Because the mean winds over the Gulf are to the west, the associated Ekman flow is to the north. These winds to the west reach maximum speeds near ~22°S and extend over the full N–S extent of the Gulf. Figure 7 shows the seasonal E–W wind stress. Calculations of the

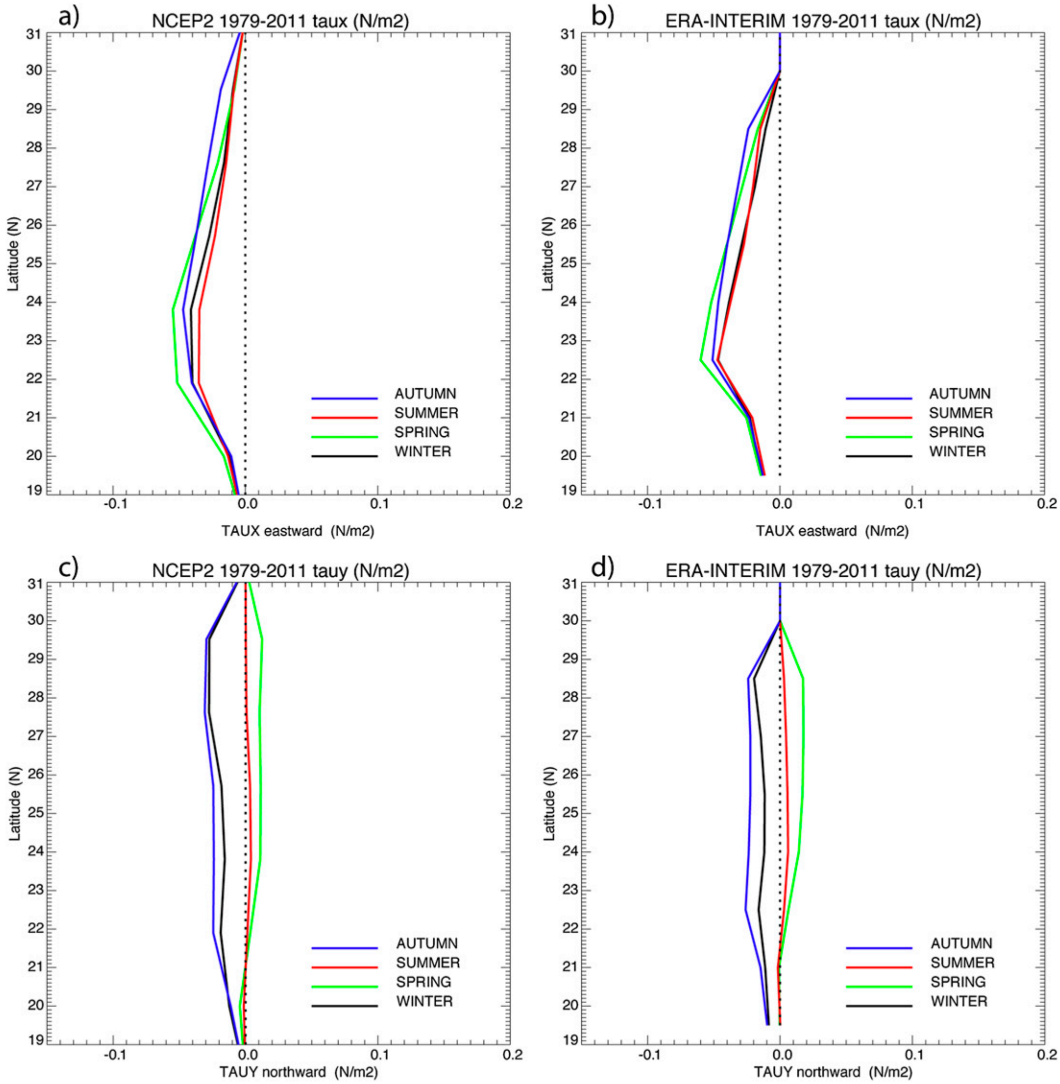

FIG. 7. Seasonal (a),(b) E–W components of mean wind stress and (c),(d) N–S component in the central Gulf of Mexico from (left) National Centers for Environmental Prediction (NCEP)–U.S. Department of Energy (DOE) Atmospheric Model Intercomparison (AMIP)-II reanalysis (NCEP-2) for 1979–2011 reanalysis and (right) European Centre for Medium-Range Weather Forecasts (ECMWF) Interim Re-Analysis (ERA-Interim) for 1979–2011.

wind stress curl and the Sverdrup balance give an interior transport to the south on the order of 5 Sv (1 Sv $\equiv 10^6\,m^3\,s^{-1}$). The contribution to the curl from the N–S wind component is very small. There is a seasonal variability, but it is not crucial for the issues studied here. For our estimate of 5 Sv, we choose a conservative width of 10°E–W influenced by the curl. Note that it is the Sverdrup transport that is the focus here, not the flow in the Ekman layer directly.

7. Comparison with the central North Atlantic

When we think of the wind forcing of the major North Atlantic anticyclonic gyre, we know that the Sverdrup transport toward the south is a major part of the flow pattern. We tend to focus on the center of the gyre, and know that the (negative) curl goes to zero much farther to the south, where the North Equatorial Current carries the Sverdrup transport to the west. In the Gulf of Mexico, however, the curl changes sign at ~22°N, just beyond the Yucatan Peninsula. Thus there is little space in the southern Gulf for the westward transport to carry the Sverdrup transport. We suggest that the observed flow to the west on the outer shelf (~50–200 m) is possibly the result of the interior Sverdrup transport.

Although the observations suggest that the mean flows to the west are concentrated in the north and south parts of the Gulf, one calculation, based simply on the

N–S-mean flow, seems relevant here. The mean observed upper-layer speeds are ~10 cm s^{-1} near the sea surface; we ask what depth is required for a near-surface layer to transport 5 Sv to the west, assuming a linear decay with depth: the result is 100 m. Different assumptions of the vertical shear give different results, but this is not crucial to the discussion. We know from previous work (Sturges and Kenyon 2008, their Table 2) that the density surfaces below ~300-m slope in the correct way to allow a geostrophic shear to permit the necessary slow, mean return flow back to the east. Sturges (2005) found that the potential vorticity surfaces are consistent with a mean flow from the Gulf back to the Caribbean at depths of ~1200 m and deeper. Thus, it is plausible to conclude that there is ample return flow at depth in the central Gulf. In their original paper, Sturges and Kenyon discussed the usual processes that could mix surface water downward in the western Gulf. More recently, Cenedese (2012) has shown that a great deal of downwelling of surface waters is accomplished by mixing in strong flows, which could take place as the Loop Current rings reach the western boundary; see also the additional wind-forced mechanism discussed by Deremble and Dewar (2012).

Although the Sverdrup transport is on the order of 5 Sv, the Ekman transport (to the north) can account for only on the order of 1 Sv of meridional flow; assuming no momentum losses, it will take on the order of ~5 yr of wind pumping to produce the required momentum to account for the Sverdrup transport. This is not new: a similar situation is obtained in the North Atlantic, where there is ample space (N–S) for the Sverdrup flow to the south to occur in a separate region of the ocean from the westward-flowing water that feeds the western boundary current. Apparently in the Gulf of Mexico, because the western Gulf is closed and the N–S extent is so limited, these flows occur in the same area, making for a more confined situation than we normally would expect.

It is tempting to want to compute the slope of density surfaces in the E–W direction in the upper layers, to confirm the idea of a Sverdrup flow to the south. Unfortunately, the great variability resulting from the Loop Current and its rings causes an uncertainly in the E–W slope that is larger than the slope itself.

8. Deep flow in the central Gulf of Mexico

The dominant mechanism controlling the observed deep, near-bottom flow in the northern Gulf is the rectification of topographic Rossby waves. DeHaan and Sturges (2005), following the results of Hamilton (1990, see also 2007a,b and 2009) and Hamilton and Lugo-Fernandez (2001), describe an extensive collection of current-meter observations of flow at these depths, over many years, with mean flows along the northern border consistently to the west. There are clear suggestions from the geostrophic flow (via the density field), showing persistent flow to the west all around the northern rim of the Gulf. The horizontal extent to which these flows extend out into the basin, however, is not well known.

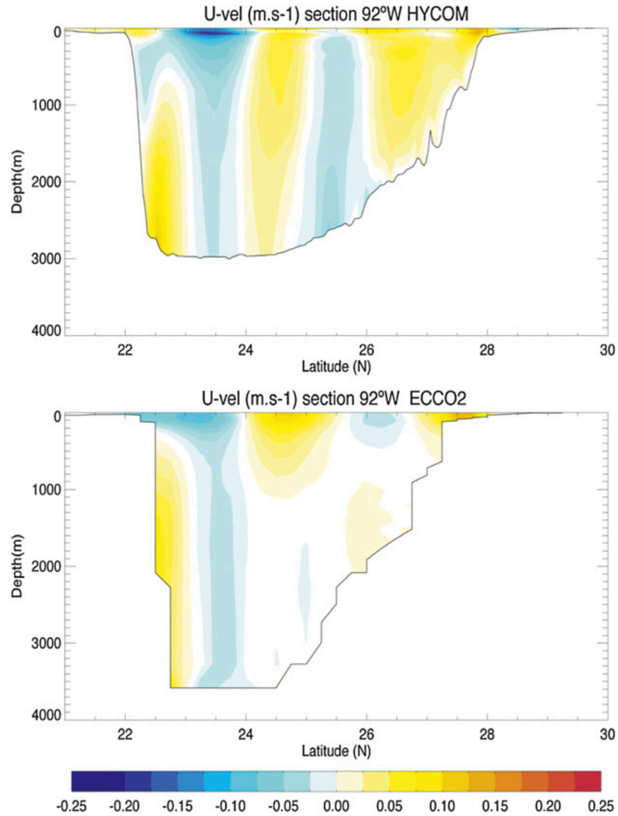

FIG. 8. Mean E–W flow (cm s^{-1}) at 92°W from (top) HYCOM and (bottom) ECCO2 (courtesy C. Wunsch).

Figure 8 shows a vertical cross section of the mean E–W flow in HYCOM and the ECCO2 model. These show the flow at 92°W rather than the mean from three longitudes as in the earlier figures. The flow near the northern boundary is to the east from depths of ~100–2000 m. A narrow sliver of flow to the west along the northern boundary can be found near the bottom below 2000 m, but there is no evidence in either model of a general near-bottom flow to the west as suggested by the extensive current-meter observations. The two models show different details in the northern half of the basin, yet neither shows any substantial deep flow back to the east.

One recent model-based analysis, however, does show large-scale deeper return flow, attributed to the LC–LC eddy-shedding cycle (Chang and Oey 2011). The point to be made is that a substantial mean flow to the west in the near-surface layers, which is a clear result in the data, must have a deeper return.

Such deep flows are normally too weak to be observed directly where the basin is wide, although the mean should be clear (one would think) in a model result. A transport of ~5 Sv back to the east, if spread over the full area of the central Gulf, requires a mean speed of only ~0.2 cm s^{-1}, which may be below the capability of the models as well as of the observations. The long-term Mexican moorings in the more narrow Yucatan Channel, however, found a deep return flow back into the Caribbean of 1.3 Sv (Rivas et al. 2005, see their Fig. 3), consistent with the return flow described in the analysis of Sturges (2005).

9. Discussion

One difficulty comes to mind: if there is a mean upper-layer E–W flow on the order of 10 cm s^{-1}, why do dynamic height maps not show it as an obvious feature? The regions of observed flow to the west, however, are found in the south and in the north at depths less than ~1000 m, so the desired comparisons are not commonly computed [e.g., the Maul and Herman (1985) dynamic height maps relative to 1000 db]. Molinari et al. (1978), in addition to maps relative to 1000 db, computed monthly maps relative to 500 db. The data are skimpy in the south, but a slope (total change in height) on the order of ~15 cm with the correct sign can be found—but only in some months where adequate data were present.

The westward flow in the south continues, presumably, along the northern rim of the permanent cyclonic gyre in the far southwest corner of the Gulf; that feature is well known. The region of interest in the north does not show well on the dynamic height maps.

Figure 9 shows a schematic of the flows in the upper layers consistent with the discussion in this paper. Both the directly wind-driven flow along the southern boundary, as well as the return flow of the Sverdrup transport, turn to the north and form a western boundary current (see, e.g., Sturges 1993). The manner in which the western boundary current extends into the interior is, of course, obscure, both here and in the larger ocean basins. This schematic shows only that part of the upper-layer flow proposed here to rationalize the mean flow reported in the observations. Two well-known major features of the western Gulf are not shown: the permanent cyclonic gyre in the southwest, and the mean anticyclone in the northwest (which is usually obscured by the presence of Loop Current rings).

The primary conclusion from this work is that long-term means of observations from ship drift and from near-surface drifters show N–S mean flow on the order of 10 cm s^{-1} to the west and appear to be well above the possible known errors. The results of these observations are not duplicated in the results of the numerical models we have examined, some briefly by informal communication, but more carefully with the HYCOM results.

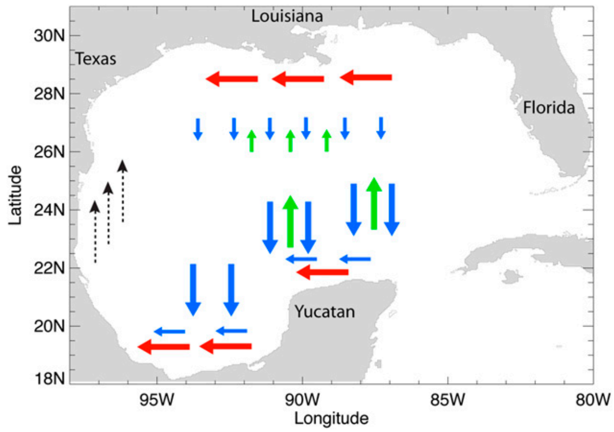

FIG. 9. A schematic of the flow in the upper layer (above 300 m) as proposed in the text. Red arrows show near-surface, wind-driven flow along the inshore shelf. Green arrows show upper Ekman-layer flow. Blue arrows show the Sverdrup return flow, including the segment at the outer edge of the wide continental shelf off northern Mexico. Black dashed arrows show the approximate position of the western boundary current off the coasts of Mexico and Texas. Arrow lengths are not to scale.

Some colleagues have suggested that it might be instructive to compute the difference between the Lagrangian and Eulerian mean values, using model output. While it would appear that the ship-drift and near-surface drifter observations are in the Lagrangian frame, the ship-drift positions are only 1 day apart, and the drifter observations are center differences between fixes only 18 h apart. The time scale of the passage of energetic Loop Current rings, by contrast, is many weeks. It would take much longer analysis periods (in the time steps) to obtain a Lagrangian measure if ring motions are a source of the discrepancy.

One possible candidate for an explanation comes from the deep flow. At the north side in Fig. 8, we see that the mean flow to the east (presumably the result the passage of Loop Current rings) extends all the way to the bottom. The flow at ~1000 m, near the north end of the section, requires that pressure surfaces slope down to the north. The ECCO2 model (C. Wunsch 2012, personal communication), as shown in Fig. 8, also shows flow to the east at 1000–2000 m, similar to that of HYCOM. A similar flow to the east at 1000 m is found in the model results of Kantha, as presented by Nowlin et al. (2001) and confirmed by L. Kantha (2012, personal communication).

One possible conclusion, therefore, is perhaps more of a speculation: that some numerical models develop a deep flow to the east in the northern Gulf of Mexico that is not found in nature. The models' flow to the east requires pressure surfaces to slope downward from south to north; the observations in the deep northern Gulf of Mexico show flow to the west, requiring pressure surfaces to slope upward toward the coast. The difference between these two flow patterns reverses the tilt of the pressure surfaces above the deep flow. The upper-layer flow to the west, discussed here, which appears to be above ~300 m, requires a sea surface slope upward from south to north; this slope would therefore not be found in the model results.

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REFERENCES

Bleck, R., 2002: An oceanic general circulation model framed in hybrid isopycnic–Cartesian coordinates. *Ocean Modell.,* **37,** 55–88.

Cenedese, C., 2012: Downwelling in basins subject to buoyancy loss. *J. Phys. Oceanogr.,* **42,** 1817–1833.

Chang, Y.-L., and L. Y. Oey, 2011: Loop Current cycle: Coupled response of the Loop Current with deep flows. *J. Phys. Oceanogr.,* **41,** 458–471.

——, and L.-Y. Oey, 2012: Why does the Loop Current tend to shed more eddies in summer and winter? *Geophys. Res. Lett.,* **39,** L05605, doi:10.1029/2011GL050773.

Chang, O., X. Lu, F.-H. Xu, H.-F. Lu, and A. Fujisaki, 2011: 2010 Oil spill—Trajectory projections based on ensemble drifter analyses. *Ocean Dyn.,* **61,** 829–839, doi:10.1007/s10236-011-0397-4.

Chassignet, E. P., L. T. Smith, G. T. Halliwell, and R. Bleck, 2003: North Atlantic simulations with the Hybrid Coordinate Ocean Model (HYCOM): Impact of the vertical coordinate choice, reference pressure, and thermobaricity. *J. Phys. Oceanogr.,* **33,** 2504–2526.

Cummings, J. A., 2005: Operational multivariate ocean data assimilation. *Quart. J. Roy. Meteor. Soc.,* **131,** 3583–3604, doi:10.1256/qj.05.105.

DeHaan, C., and W. Sturges, 2005: Deep cyclonic circulation in the Gulf of Mexico. *J. Phys. Oceanogr.,* **35,** 1801–1812.

Deremble, B., and W. K. Dewar, 2012: First-order scaling law for potential vorticity extraction due to wind. *J. Phys. Oceanogr.,* **42,** 1303–1312.

DiMarco, S. F., W. D. Nowlin Jr., and R. O. Reid, 2005: A statistical description of the velocity fields from upper ocean drifters in the Gulf of Mexico. *Circulation in the Gulf of Mexico: Observations and Models, Geophys. Monogr.,* Vol. 161, Amer. Geophys. Union, 101–110.

Halliwell, G. R., 2004: Evaluation of vertical coordinate and vertical mixing algorithms in the Hybrid-Coordinate Ocean Model (HYCOM). *Ocean Modell.,* **7,** 285–322.

Hamilton, P., 1990: Deep currents in the Gulf of Mexico. *J. Phys. Oceanogr.,* **20,** 1087–1104.

——, 2007a: Deep-current variability near the Sigsbee Escarpment in the Gulf of Mexico. *J. Phys. Oceanogr.,* **37,** 708–726.

——, 2007b: Eddy statistics from Lagrangian drifters and hydrography for the northern Gulf of Mexico slope. *J. Geophys. Res.,* **112,** C09002, doi:10.1029/2006JC003988.

——, 2009: Topographic Rossby waves in the Gulf of Mexico. *Prog. Oceanogr.,* **82,** 1–31, doi:10.1016/j.pocean.2009.04.019.

——, and A. Lugo-Fernandez, 2001: Observations of high speed deep currents in the northern Gulf of Mexico. *Geophys. Res. Lett.,* **28,** 2867–2870.

Maul, G. A., and A. Herman, 1985: Mean dynamic topography of the Gulf of Mexico with application to satellite altimetry. *Mar. Geod.,* **9,** 27–44.

Menemenlis, D., J. Campin, P. Heimbach, C. Hill, T. Lee, A. Nguyen, M. Schodlock, and H. Zhang, 2008: ECCO2: High resolution global ocean and sea ice data synthesis. *Mercator Ocean Quarterly Newsletter,* No. 31, Mercator Ocean, Toulouse, France, 13–21.

Molinari, R. L., J. F. Festa, and D. W. Behringer, 1978: The circulation in the Gulf of Mexico derived from estimated dynamic height fields. *J. Phys. Oceanogr.,* **8,** 987–996.

Nowlin, W. D., A. E. Jochens, S. F. DiMarco, R. O. Ried, and M. K. Howard, 2001: Deepwater physical oceanography reanalysis and synthesis of historical data. U. S. BOEM Tech. Rep 2001-064, 166 pp.

Ohlmann, J. C., and P. P. Niiler, 2005: Circulation over the continental shelf in the northern Gulf of Mexico. *Prog. Oceanogr.,* **64,** 45–81.

Pérez-Brunius, P., P. Garcia-Carrillo, J. Dubranna, J. Sheinbaum, and J. Candela, 2012: Direct observations of the upper layer circulation in the southern Gulf of Mexico. *Deep-Sea Res. II,* **85,** 182–194, doi.org/10.1016/j.dsr2.2012.07.020.

Poulain, P.-M., R. Gerin, and E. Mauri, 2009: Wind effects on drogued and undrogued drifters in the eastern Mediterranean. *J. Atmos. Oceanic Technol.,* **26,** 1144–1156.

Richardson, P. L., 1997: Drifting in the wind: Leeway error in shipdrift data. *Deep-Sea Res. I,* **44,** 1877–1903.

Rivas, D., A. Badan, and J. Ochoa, 2005: The ventilation of the deep Gulf of Mexico. *J. Phys. Oceanogr.,* **35,** 1763–1781.

Sturges, W., 1993: The annual cycle of the western boundary current in the Gulf of Mexico. *J. Geophys. Res.,* **98,** 18 053–18 068.

——, 2005: Deep-water exchange between the Atlantic, Caribbean, and Gulf of Mexico. *Circulation in the Gulf of Mexico, Geophys. Monogr.,* Vol. 161, Amer. Geophys. Union, 263–278.

——, and K. E. Kenyon, 2008: Mean flow in the Gulf of Mexico. *J. Phys. Oceanogr.,* **38,** 1501–1514.

Xu, F.-H., and L.-Y. Oey, 2011: The origin of along-shelf pressure gradient in the Middle Atlantic Bight. *J. Phys. Oceanogr.,* **41,** 1720–1740.