Deep Cyclonic Circulation in the Gulf of Mexico

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(Manuscript received 20 December 2002, in final form 10 March 2005)

ABSTRACT

The anticyclonic Loop Current dominates the upper-layer flow in the eastern Gulf of Mexico, with a weaker mean anticyclonic pattern in the western gulf. There are reasons, however, to suspect that the deep mean flow should actually be cyclonic. Topographic wave rectification and vortex stretching contribute to this cyclonic tendency, as will the supply of cold incoming deep water at the edges of the basin. The authors find that the deep mean flow is cyclonic both in the eastern and western gulf, with speeds on the order of 1–2 cm s^{-1} at 2000 m. Historical current-meter mooring data, as well as profiling autonomous Lagrangian circulation explorer (PALACE) floats (at 900 m), suggest that vertical geostrophic shear relative to 1000 m gives a surprisingly accurate result in the interior of the basin. The temperature around the edges of the basin at 2000 m is coldest near the Yucatan Channel, where Caribbean Sea water is colder by ~0.1°C. The temperature increases steadily with distance in the counterclockwise direction from the Yucatan, consistent with a deep mean cyclonic boundary flow.

1. Introduction

The Loop Current and the warm-core rings that detach from it dominate the upper-layer flow in the Gulf of Mexico. The well-known primary flow enters the Caribbean Sea from the open Atlantic Ocean, flows into the Gulf of Mexico through the Yucatan Channel, the constriction between Mexico and Cuba, and then leaves the gulf to form the Florida Current and Gulf Stream (see, e.g., Johns et al. 2002; Schmitz and Richardson 1991; Niiler and Richardson 1973). There have been several attempts to summarize the flow in the gulf using the historical data (see Hoffmann and Worley 1986; Molinari et al. 1978). The upper-layer flow has been observed fairly well in recent years by satellite observations of temperature and sea-surface height. The main flow is restricted to approximately 850 m because the Loop Current exits through the Straits of Florida.

The deep flow has been described by Hamilton (1990). A magnificent new set of mooring data in the Yucatan Channel has been reported recently (Bunge et al. 2002; Ochoa et al. 2001; Sheinbaum et al. 2002).

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The mean upper-layer flow is well described by almost any temperature or density surface above the main thermocline. Figure 1 shows the mean temperature at 400 m. The mean depth of the 27.0 σ_θ density surface shows a similar structure. The dots that appear to be data points are local means at essentially random concentrations of ~5–10 hydrostations; these were selected, using data from many years, from the full historical National Oceanographic Data Center (NODC) database (Conkright et al. 2000) in an attempt to suppress the very great time variability of warm-core rings propagating to the west.

Two individual warm-core features are evident in Fig. 1. In the east, although the Loop Current position is quite variable, a clear mean emerges. In the central and western gulf the anticyclonic pattern is maintained both by the mean wind field and by the passage of large warm-core rings that have separated from the Loop Current. The relative importance of these two forcing mechanisms remains an open question.

Though we know the upper-layer flow fairly well from observations, we do not know the deep flow well. Several numerical models include the Gulf of Mexico, but good observations at depth are scarce. The intriguing fact, however, is that the temperature signature of the upper-layer flow, so obviously anticyclonic, appears to extend to depths of ~1500 to 2000 m in the data. Because we know that the flow in the Florida Current

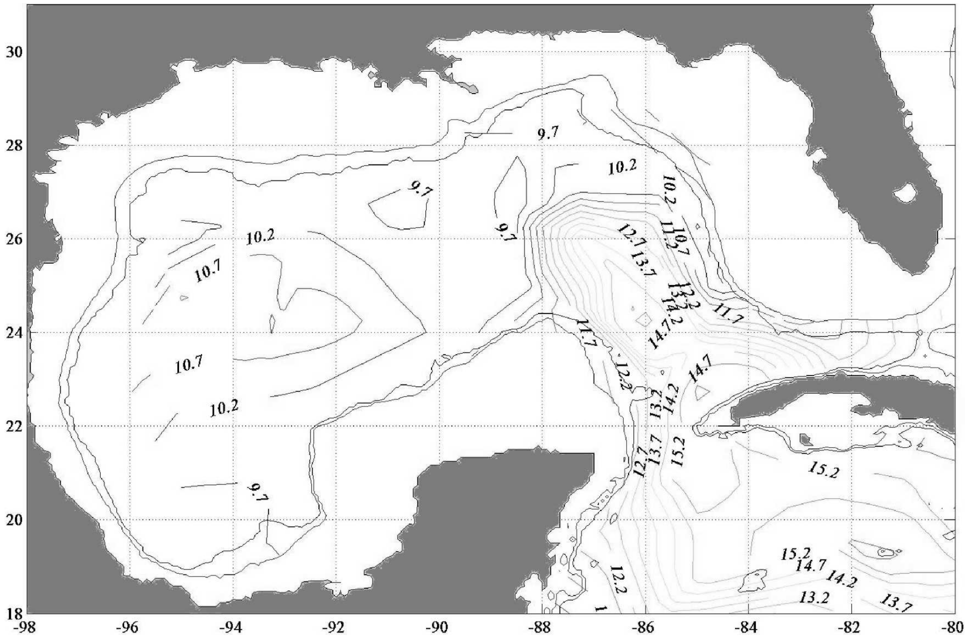

FIG. 1. The time-mean temperature distribution at 400 m, based on the full historical database from NODC. The individual dots that appear to be data points are local means computed from ~5–10 hydrostations concentrated near that point. The 1000-m isobath is shown. The flow between Cuba and Florida is not resolved in this dataset.

penetrates only to ~850 m, we might expect the warm-core structure of Fig. 1 to extend only to that depth—whereas, in fact, it goes much deeper.

Figure 2 shows a map of potential temperature at 1250 m. The warm-core patterns are still evident at this depth. They remain evident, if not so clearly, as deep as 2000 m. The horizontal temperature difference at 1500 m is ~0.06°C in the east between the central warm

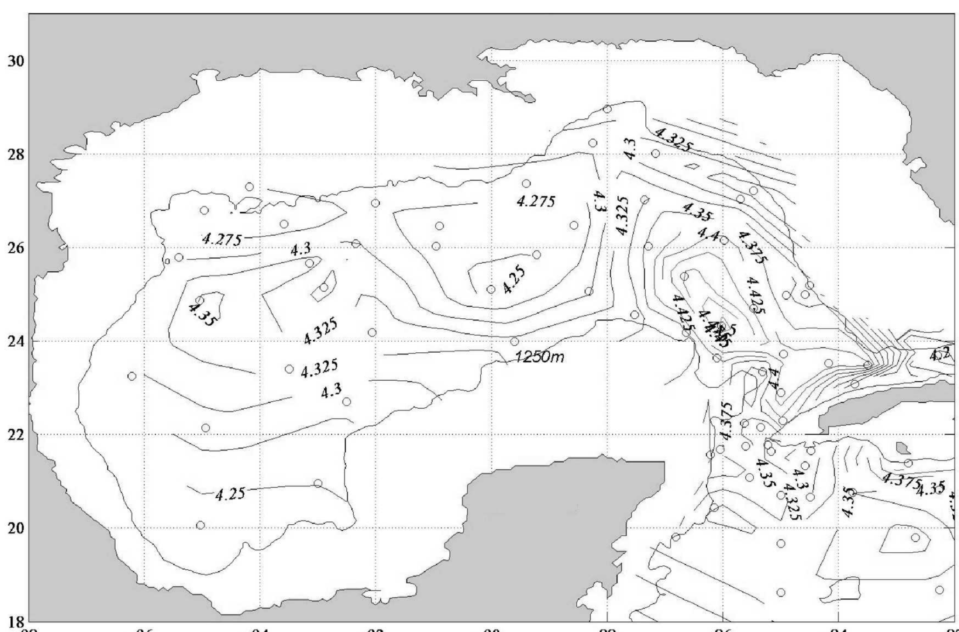

FIG. 2. The time-mean potential temperature distribution at 1250 m. The dataset is the same as for Fig. 1. The 1250-m isobath is shown.

region and the eastern edge; in the central region of the gulf the difference has been eroded to ~0.04°C between the central warm region and the northern edge. The signal is still evident at depths of 2000 m. We have included, in the appendix, figures that show the mean potential temperature distributions in 1° boxes at 1500 and 2000 m. Table 1 shows the essential information for our purpose here.

In general, a "warm-core ring" in the upper ocean suggests anticyclonic flow, but in deep water this signal clearly tells us only about vertical shear. For reasons put forth in the next section, it seems reasonable to expect that the deep mean flow should be *cyclonic*. Thus we ask, is there evidence for a reversal of the mean flow between ~800 and 1500 m?

Some readers may raise the additional question: Why do we care about the direction of deep mean flow? (We might reply rhetorically—why do we care about the direction of the deep western boundary current off the U.S. East Coast?) The Caribbean Sea is the only source of deep water for the Gulf of Mexico, for no deep water is formed locally. The Caribbean source of cold deep water (e.g., Bunge et al. 2002) supplies oxygen and nutrients. The return flow flushes the deep gulf. It is unlikely that we can understand the magnitude and effects of these transports if we do not even know the *direction* of the deep mean flow.

2. Why we expect the deep flow to be cyclonic

There are three mechanisms that could cause the deep flow to be cyclonic. First, it is well known that there is a substantial amount of eddylike activity at depth, composed mostly of topographic Rossby waves (Hamilton 1990; Oey and Lee 2002). Topographic rectification of these waves would contribute to cyclonic mean flow. Second, there is the introduction of cold deep water from the Caribbean, through the Yucatan Channel. At 2000 m, the Caribbean is ~0.1°C cooler than the gulf. Recent observations (e.g., Bunge et al. 2002) show that the deep exchange is as large as 5–10 Sv (Sv $\equiv 10^6\ m^3\ s^{-1}$) for intervals of many months as the

TABLE 1. Mean potential temperature (°C) in the centers and edges of the eastern and central Gulf of Mexico. The first value is potential temperature, and the second is the number of samples.

	Central warm region	Outer cool region	Difference
Eastern gulf, 1500 m	4.185 (44)	4.130 (123)	0.045
Central gulf, 1500 m	4.142 (32)	4.105 (45)	0.037
Eastern gulf, 2000 m	4.050 (19)	4.016 (34)	0.034
Central gulf, 2000 m	4.053 (21)	4.024 (19)	0.029

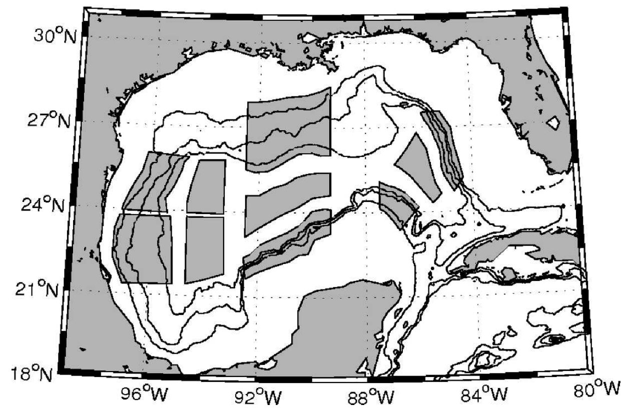

FIG. 3. Location of the regions used to form data means for constructing shear profiles.

Loop Current goes through a ring-shedding cycle. Parcels of water would of course gradually lose their temperature deficit as they mix downstream from the Yucatan. However, we expect cold incoming parcels to hug the right-hand slope even after some mixing and sinking. This supply of cold dense water is similar to, even if much weaker than, the supply of North Atlantic Deep Water to the deep western boundary current along the east coast of the United States, where cold water overflows a sill and flows with the boundary to the right. This effect also introduces colder water around the periphery, enhancing the existing "warm core" shear structure.

A third mechanism operates only in the eastern gulf. During one phase of the Loop Current cycle, as deep Caribbean waters flow over the sill into the gulf, the bottom falls away. Water that enters at near 2000 m flows into a region where the bottom drops abruptly to ~3500 m. Vortex stretching then leads to a cyclonic spin of the entering fluid, consistent with similar findings of Spall and Price (1998). By contrast, when water leaves the gulf during the reverse phase of the Loop Current cycle, the ambient stratification in the gulf, even though it is weak, tends to restrict outgoing fluid to that originating from depths above the sill. So we expect much less vortex compression on the other half of the deep flow cycle.

3. Deep temperature structure and vertical shear

a. Eastern gulf

We have computed the vertical shear associated with these cyclonic patterns that we expect to see in the deep water. Figure 3 shows several regions in which we computed the mean hydrographic conditions based on the available NODC database.

Because the variability is so great, we wished to use

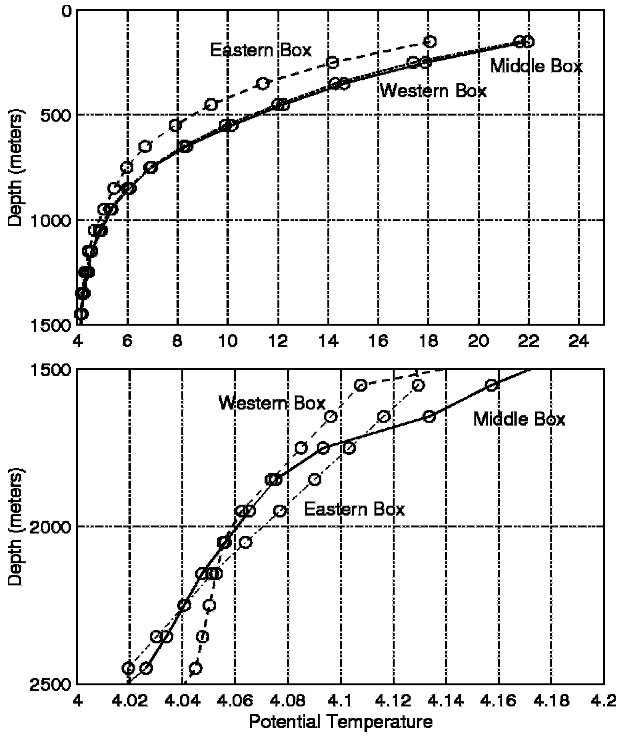

FIG. 4. Mean vertical potential temperature distributions in the eastern Gulf of Mexico; an expanded scale is shown in the lower box. The regions used are shown in Fig. 3.

observations over as great a time span as possible and over as great a horizontal extent as seemed appropriate. From a construction of composite temperature–salinity (T–S) curves, we concluded that observations of temperature have much less apparent error, or scatter, than salinities. Because the salinity gradients are so small in the deep gulf, we have chosen to compute density from a mean temperature–salinity curve, using the temperature data alone for gradient information.

In the eastern gulf, it turns out that the shear profile on the eastern side has a much better signal-to-noise ratio than that on the western side. The western side has sparse observations, contributing to the poor signal-to-noise ratio. The position of the Loop Current is constrained by the Florida shelf on the eastern side, but not on the western side, which may also contribute to the greater variability on the western side.

Figure 4 shows the resulting mean potential temperature signal. Below 1500 m, the standard deviation of the potential temperature variability is ~0.02°C in the central and eastern boxes; Fig. 4 shows that the potential temperature differences are on the same order. (The standard error of the means, of course, is smaller.)

We computed an error estimate in the following way. Because the standard deviations of the potential temperature data values are as large as the signal, we estimated a standard error of the mean by computing the composite mean values between 1000 and 2000 m. The potential temperature difference on the eastern side was 0.13°C, with a standard error of the mean of 0.03°C (based on 106 observations in the east and 163 in the center). We are not able to compute the details of the vertical shear with great accuracy, although the *mean difference* is reliably greater than the standard error. It is true, however, that the biggest variability in these data is in time. Multiple data points on a single hydrocast are not independent in the way we would like. The vertical averaging, however, will reduce problems from instrumental errors, internal waves, and other such sources. Therefore, the significance level is not nearly as high as one would like. Nevertheless, the same result holds true for all the individual calculations: the signals are small but consistently of the sign to support the idea of cyclonic flow.

The flow at 2000 m, in Fig. 5 (relative to 1000 m), is ~1 cm s^{-1} to the north, consistent with the expectation of cyclonic mean flow. The problem remains to determine an appropriate "level of no motion." This issue is treated in a later section.

The signal-to-noise ratio is barely adequate on the

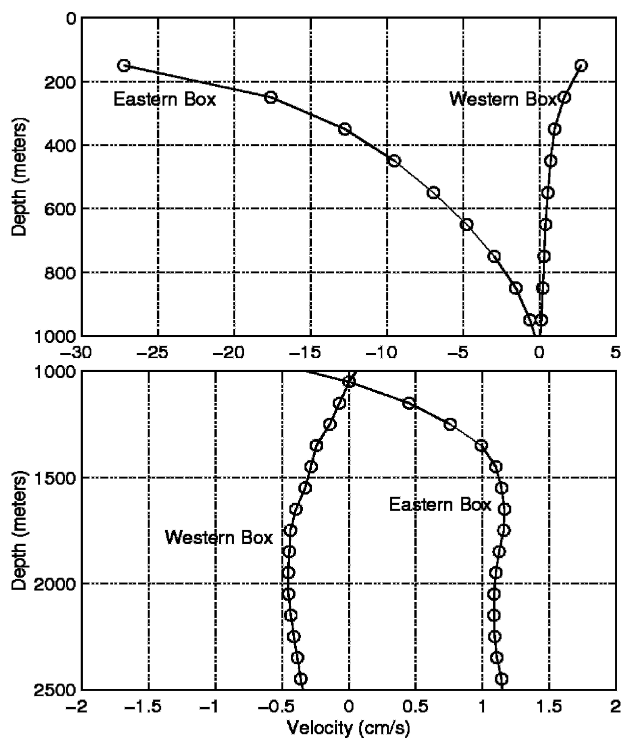

FIG. 5. The mean north–south geostrophic speeds relative to 1000 dbar computed from the potential temperature profiles of Fig. 4 and using a mean T–S curve.

east side of our calculation in the eastern gulf, but is worse on the western side. The mean potential temperature difference, only 0.05°C, has a standard error of 0.04°C. There are a number of reasons why the computed signal could be so small, but we are unable to offer any definitive explanation other than the fact that the data are quite sparse on the western side; further speculation seems pointless. The distribution of the number of stations in the eastern region is shown in Fig. 6. (This is the region with the most data.)

b. *Central gulf*

In a fashion similar to that in the eastern gulf, we have computed the mean geostrophic vertical shear in the central gulf using the regions shown in Fig. 3. Figure 7 shows the velocity structure relative to 1000 m.

The velocity is shown for only the northern side of the gulf because the signal-to-noise ratio is below the noise level on the southern side. The flow at 2000 m, relative to 1000 m, is to the west, suggesting a cyclonic flow pattern. (Again, the issue of justifying the choice of reference level is postponed to the next section.) The velocity at depth is only a *few tenths* of a centimeter per second. This value seems almost ridiculously small, but we note that, first, it *is* to the west and, second, the shear profile below 1000 m is monotonic to at least ~2300 m (in Fig. 7). These factors all support the cyclonic hypothesis that we have put forward.

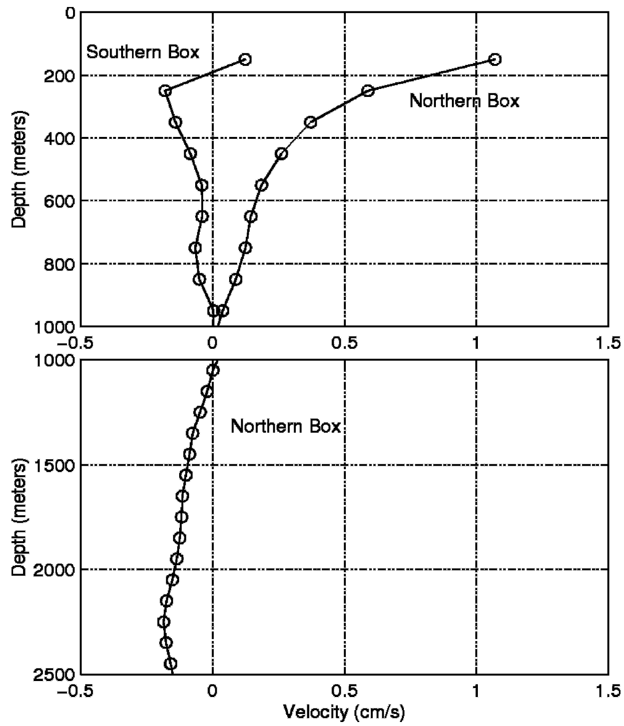

FIG. 7. The mean east–west speeds, relative to 1000 dbar, computed from the temperature signal in the central gulf. The result is shown for the deeper section only for the northern half, where the signal-to-noise ratio allows a significant result.

c. *Western gulf*

The averaging areas, or boxes, shown in Fig. 3 also show areas in the western gulf where we made calculations similar to those shown previously. In Fig. 8 the velocity profile in the southwestern gulf shows speeds (relative to 1000 m) of order 0.5 cm s^{-1}. The flow is to the south. Again, this is consistent with the assumption of cyclonic flow. The signal does not emerge above the noise for the northern region in the west.

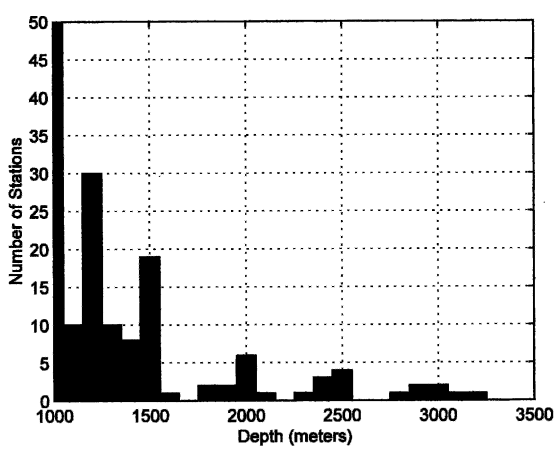

FIG. 6. A histogram of the number of observations at each depth for the eastern region (shown in Fig. 3) of the eastern gulf for these calculations.

d. *Variability of the geostrophic shear*

The geostrophic results presented in Figs. 5, 7, and 8 show the mean values. To what extent, one wonders, is it possible to estimate the variability of the geostrophic velocity?

Figure 9 shows a collection of 46 station pairs from which individual velocity shear patterns were computed. The stations in each pair are from a single cruise. (The data used here are from the same dataset as in the previous section but are treated differently.) We had expected that using pairs of stations from individual cruises might improve the accuracy of the individual calculations because calibration issues would be minimized.

As would be expected, the means from these calculations are essentially the same as from the previous calculations. There are many reversals in sign, but similar means emerge. In an attempt to construct the most

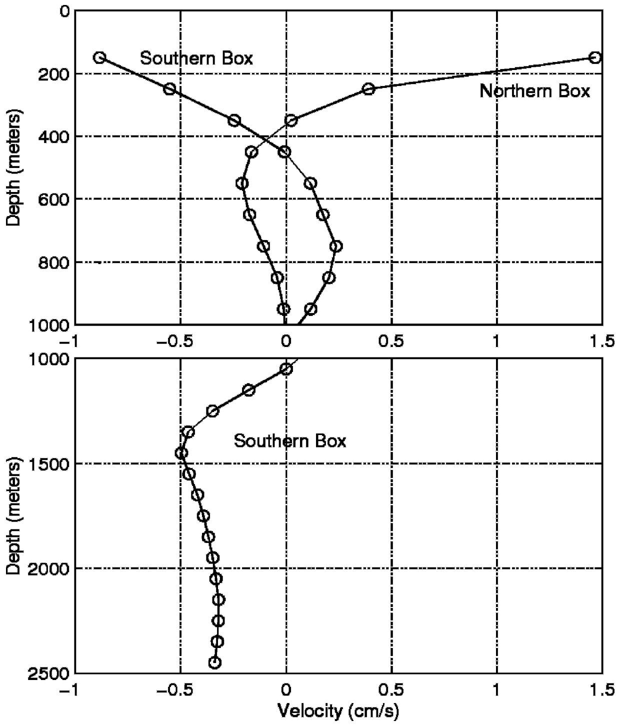

FIG. 8. The mean north–south geostrophic speeds in the western Gulf of Mexico.

accurate values, we have computed an absolute mean at each level from which to compute the standard deviations. These are shown in Table 2. In this case only, the values are computed relative to 2000 m (to allow consistent comparisons at varying depths). It is remarkable that the standard deviations of the values are so similar to the speeds themselves (because the means are so near zero). Because these are determined from 46 individual calculations, the standard errors of the mean

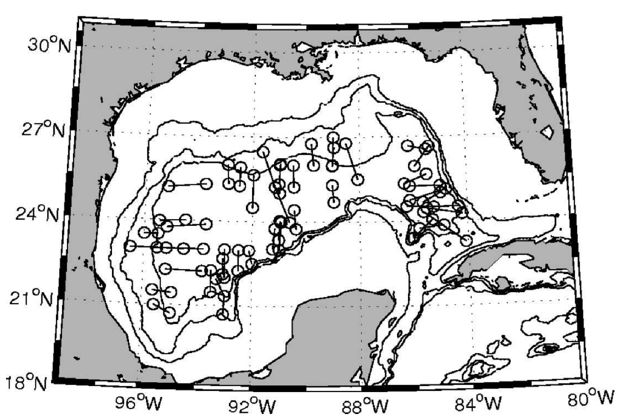

FIG. 9. Locations of station pairs (from single cruises) from which geostrophic velocities can be computed. Results are shown in Table 1 (see text).

TABLE 2. The mean velocity shear and standard deviation relative to 2000 m. The values were calculated by finding the magnitude of all the individual station pairs and taking the mean and standard deviation.

Depth (m)	Mean velocity magnitude (cm s^{-1})	Standard deviation of velocity (cm s^{-1})
100	16.7	15.0
500	5.7	5.1
1000	2.1	2.1
1500	0.5	0.5

would be reduced by ~6.7 (i.e., $\sqrt{45}$), suggesting that the mean values are indeed significantly different from zero.

There is an important point that perhaps should be emphasized. The mean speeds are very small, but the individual velocity values are certainly strong. The fact that the individual velocity values are so large (i.e., the variance is large) is why the signal-to-noise-ratio is so small.

4. Resolving the reference-level issue: Observations of deep flow

a. Eastern gulf

Long-term current-meter moorings that are well suited to provide a check on absolute deep velocities are scarce. The best observations of deep velocity that we have found to be appropriate as a reference for these geostrophic calculations were made in a Minerals Management Service (MMS)–sponsored program in the early 1980s (Hamilton 1990).

Figure 10 shows the locations of two deep moorings; plots of the north–south (along isobath) velocity components from the easternmost mooring are in Fig. 11. The means are shown in Fig. 12. The variability is similar at the other mooring, and the mean speeds at depth are also to the north, but the values are only barely greater than zero. Mooring A is in 1700 m of water, with the deepest current meter at 1600 m. Mooring G is deeper, in 3200 m, with the deepest instrument at 3175 m.

The spectrum of the north–south velocity component at the deepest instrument shows that the energy peaks at periods of 20–30 days. There are ~500 days of total record at the lower instrument, capturing ~20 "periods." Because the correlation coefficient falls to zero at one-fourth of a period for a narrowband signal, we estimate that there are approximately 80 independent observations, leading to an estimated uncertainty (standard error of the mean) of ~0.5 cm s^{-1}. Thus we conclude that the mean value at 1600 m (Fig. 12) of ~4 cm s^{-1} is to the north and is significantly different from zero.

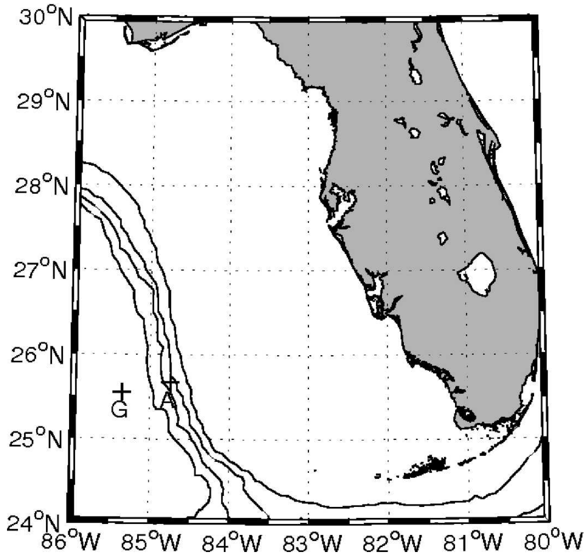

FIG. 10. The location of current-meter moorings A and G off the west Florida escarpment.

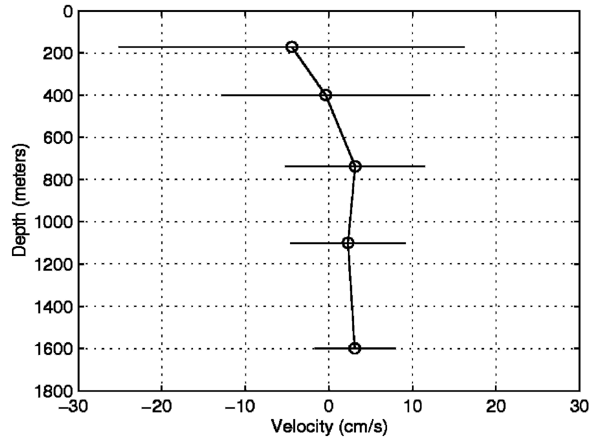

FIG. 12. The vertical distribution of mean longshore component of velocity at the inshore mooring (A) of Fig. 10. The error bars show one standard distribution of the velocity signal, not the standard error.

Perhaps we should emphasize the main point here: the observed mean speed over nearly 3 years is several centimeters per second to the north, and is significantly different from zero, during this time interval. Considering the large variability of such signals on decadal time scales we would not assume that this mean value or error estimate is valid for all time. It does seem plausible, however, that our choice of 1000 m as a reference level is an effective choice for the purpose at hand. Our estimated mean speed at 1600 m, relative to 1000 m, was found to be ~1.1 cm s^{-1} (Fig. 5). To this we may add ~3 cm s^{-1} (Fig. 12), from the bottom current meter, suggesting a deep cyclonic speed in the eastern gulf O(3–4 cm s^{-1}). It may seem sensible to choose a reference level shallower than 1000 m, given that the mooring shows cyclonic flow above 1000 m. We choose 1000 m because it is unlikely that there is one single reference level for the entire gulf. Figure 2 clearly shows that the surface signature penetrates to at least 1250 m in the central gulf. By choosing 1000 m, we hoped also to stay beneath most of the influence of Loop Current rings and the Loop Current itself. From our plots of the geostrophic shear, it is clear that there is very little shear in the deep water and so moving the reference level has little effect, especially given the high variability of the flow. We are not claiming that 1000 m is the correct reference level for the entire gulf, but that given the uncertainties it is reasonable, and less likely to be contaminated by the surface flow.

We explored the possible coherence between temperature and velocity at these two moorings. If cold bursts of incoming deep water from the Caribbean Sea retain their temperature deficit this far into the gulf, it would be an interesting finding. Unfortunately, the temperature signals were barely resolved at each instrument, and there was no clear coherence between velocity and temperature. At depths near 1600 m, the deepest instrument at mooring A, the temperature difference between the Caribbean and gulf water is very small, so lack of coherence there is not surprising. Near

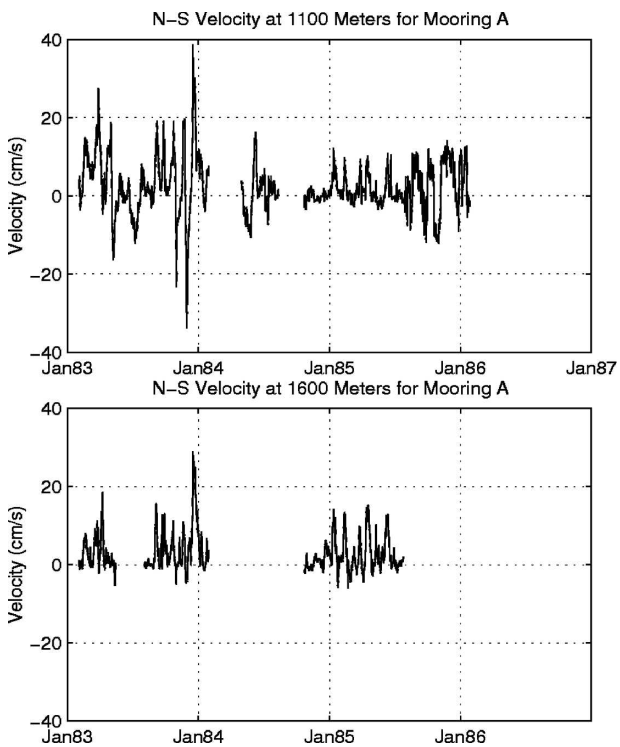

FIG. 11. Alongshore velocity components from the deepest instrument on the two moorings of Fig. 10.

sill depth (2000 m), the Caribbean is roughly 0.1°C cooler than water at the same depth in the Gulf of Mexico. However, at mooring G the nearest instrument was at ~2360 m; temperature record was composed largely of "background temperature," with occasional departures of a few hundredths of a degree. The instrument appeared (possibly) to be deeper than the incoming bursts of cooler water. Because there were so many "background temperature" data points, which appear as "zero" in a calculation of the spectrum, we considered the calculations of cross-spectral coherence to be of questionable validity.

It seems worthwhile to estimate the transports associated with these deep flows. At mooring A, the inshore mooring, the mean barotropic N–S velocity is ~3 cm s^{-1}, over a depth span of at least 1000 m and possibly twice that. If we take the horizontal extent to be only 75 km (the distance to the mooring farther offshore, G) the mean transport is thus estimated to be ~3–6 Sv. This mean value applies to a narrow boundary current.

For the stronger bursts of flow, however, the speeds are easily 25–30 cm s^{-1} at mooring G, and clearly are barotropic to ~3200 m. It is hard to escape the conclusion that the transports in these bursts (to the north) are much larger than our estimate for the boundary flow.

Additional evidence (for cyclonic flow) from deep long-term moorings is found in the work of Molinari and Mayer (1980), who measured the flow at ~1000 m offshore of Tampa Bay, Florida (27.5°N). They show (their Fig. 33) that at the uppermost mooring (at 150 m) there was almost no net along isobath flow for the whole year (June 1978–May 1979). At the deepest mooring (950 m, 100 m above the bottom) there was flow to the northwest (along the isobaths) on the order of 5 cm s^{-1} in 9 out of 11 months. In the other two months (November and December) the flow was essentially zero or in the noise level. The mean flow at the deepest mooring over the full record was ~3.3 cm s^{-1} to the northwest. At 55 m, the strongest flows were aligned with the deeper flow.

Another array of five moorings were put in an array near 28°N, 90°W to measure flow especially near the bottom (as well as full water column) at depth near 2000 m (Hamilton et al. 2003). The mean speeds ranged from ~2 to 4.2 cm s^{-1}, all to the west (along local topography). At mooring I2, which had the longest record, 2 yr, the mean speed was 4.2 cm s^{-1}, with a standard deviation of 16.75 cm s^{-1}. The topographic Rossby wave energy peaks at ~20 days. For relatively narrowband signals, there are essentially four individual observations. In 2 years this would give a standard error of 1.3 cm s^{-1}, or less than one-third of the mean.

b. *Central gulf*

Although there have been many current-meter moorings in the central gulf, the best dataset that we have found for our present purposes is composed of mean velocities from an experiment using profiling autonomous Lagrangian circulation explorer (PALACE) floats. These are a profiling version of the original ALACE floats (Davis et al. 1992; see also the web site of the manufacturer online at http://www.webbresearch.com). The National Ocean Partnership Program sponsored an experiment in the Gulf of Mexico. Approximately four dozen floats were tracked from 1998 to 2002. A potentially serious problem is that wind-induced surface drift contaminates the results. However, several navigation fixes at the surface allowed the deep float velocities to be corrected for surface motion. We are greatly indebted to our colleagues G. Weatherly and N. Wienders for allowing us to show these new results (Wienders et al. 2002; Weatherly et al. 2005). The means and variance ellipses are shown in Fig. 13. The mean velocity vectors were obtained by averaging all of the float velocities in 0.5° bins.

For these results, only bins with more than five float values were used (an admittedly arbitrary choice). While the mean values are small, there is a robust tendency for a weak cyclonic pattern. While some regions are sampled poorly (and some not at all), the tendency for cyclonic flow is evident. Figure 13 shows that the flow has become cyclonic at 900 m. Our geostrophic shear calculations suggest that the speeds increase, but only by the order of no more than ~1–2 cm s^{-1} down to ~2000 m (see Table 2).

5. The path of cold renewal water

At depths of 2000 m, the Caribbean Sea is colder by ~0.1°C than the gulf. Cold Caribbean water flows in at the Yucatan Channel during every Loop Current cycle, so, if there is a deep cyclonic boundary flow, we expect that the temperature around the edges of the gulf should increase with distance from Yucatan. To examine the possibility of such an effect, Fig. 14 shows the temperature at 2000 m, averaged in 1° boxes around the edge, plotted as a function of distance from Yucatan. Because the path is irregular, the x axis is only roughly a measure of the distance, in hundreds of kilometers, from the source of cold water entering at Yucatan. The individual data are shown, with a smoothed curve through the scattered data points. The increase in temperature with distance from Yucatan is strikingly clear. The sharpest gradient is nearest the entrance at

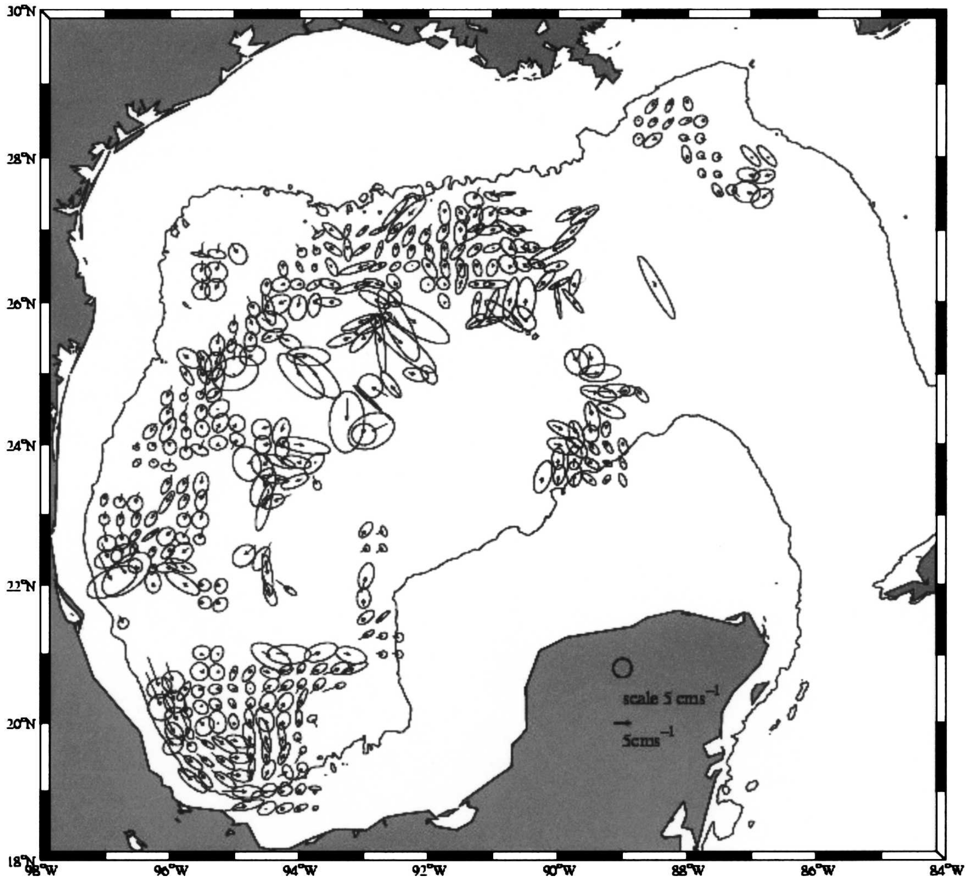

FIG. 13. The mean velocity and variance ellipses of drifters at 900 m over the course of several years. The floats surfaced every 7 days; velocity values have been corrected for surface drift effects. These are averaged in ½° bins (courtesy of N. Wienders and G. L. Weatherly, unpublished figure). No values are shown for bins with fewer than five values.

Yucatan, which is consistent with strong initial mixing. Apart from this strong gradient in the first half-dozen data points, a linear fit through the other data would be adequate. The total increase in temperature is ~0.07°C, which is consistent with the net difference in temperature between the two basins. This effect is a clear indication of the cyclonic boundary flow. However, it is entirely possible that the supply of cold water at Yucatan is a possible mechanism for *causing* the boundary flow, in a manner similar to that of the deep western boundary current along the U.S. East Coast. The historical dataset is adequate at 2000 m to see the change in Fig. 14, but at greater depths is sampled adequately only at 2500 and 3000 m. A better understanding of the mechanism and details of the deep-water renewal remain for future work.

6. Conclusions

We find that there is a substantial amount of evidence for cyclonic deep flow in the eastern Gulf of Mexico beneath the Loop Current, as well as in the western gulf. The evidence suggests that the *mean* deep flow is weak, but clearly above the noise level. The PALACE float data in the central and western gulf are very convincing. In the eastern gulf our conclusions are based in part on one long series of current-meter moorings; the steady increase in deep temperature with distance from the cold source at Yucatan, however, is unambiguous. We are not aware of *any* long-term dataset that offers conflicting evidence.

The general mechanism of topographic rectification would be expected to lead to deep cyclonic circulation in this basin, given the large amount of eddy motions generated by the Loop Current. It is also likely that the supply of cold, dense water from the Caribbean is a cause of the boundary flow and not merely a tracer. We are not able to distinguish between these two forcing mechanisms given the available data.

One tacit simplification that we have made is that the deep flows in both the eastern and western gulf can be adequately described as simple large gyres. We suspect

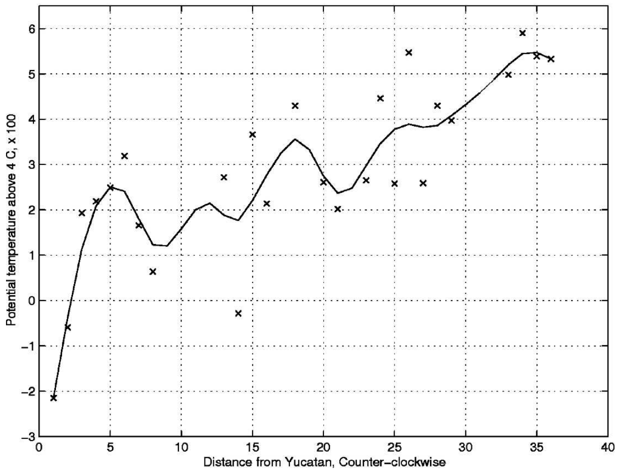

FIG. 14. Time-mean temperature around the edge of the Gulf of Mexico at 2000 m, averaged in 1° boxes, plotted as a function of distance from the sill in the Yucatan Channel. The x axis is (roughly) a measure of distance in hundreds of kilometers. The individual data points are shown (×); if a 1° box had no data, linear interpolation was used, and the record was smoothed slightly.

that better measurements in the future will lead to an awareness of a richer mean field with more detail.

Acknowledgments. We thank Dr. Peter Hamilton, SAIC, Inc., Raleigh, North Carolina, for generously providing us with mooring data on many occasions. We thank our colleagues G. L. Weatherly and N. Wienders for allowing us the use of early results of their corrected float data from the NOPP program. Also, Ms. Jane Jimeian was an invaluable ally with the data analysis. This work was supported by the U.S. Minerals Management Service, Contract 01-99-CT-31027, for which we are most grateful.

APPENDIX

Details of Deep Temperature Structure

Figure A1 shows the distribution of mean potential temperature (relative to sea surface pressure) in every 1° horizontal bin in the Gulf of Mexico at 1500 m. The printed numbers give (top to bottom) the number of

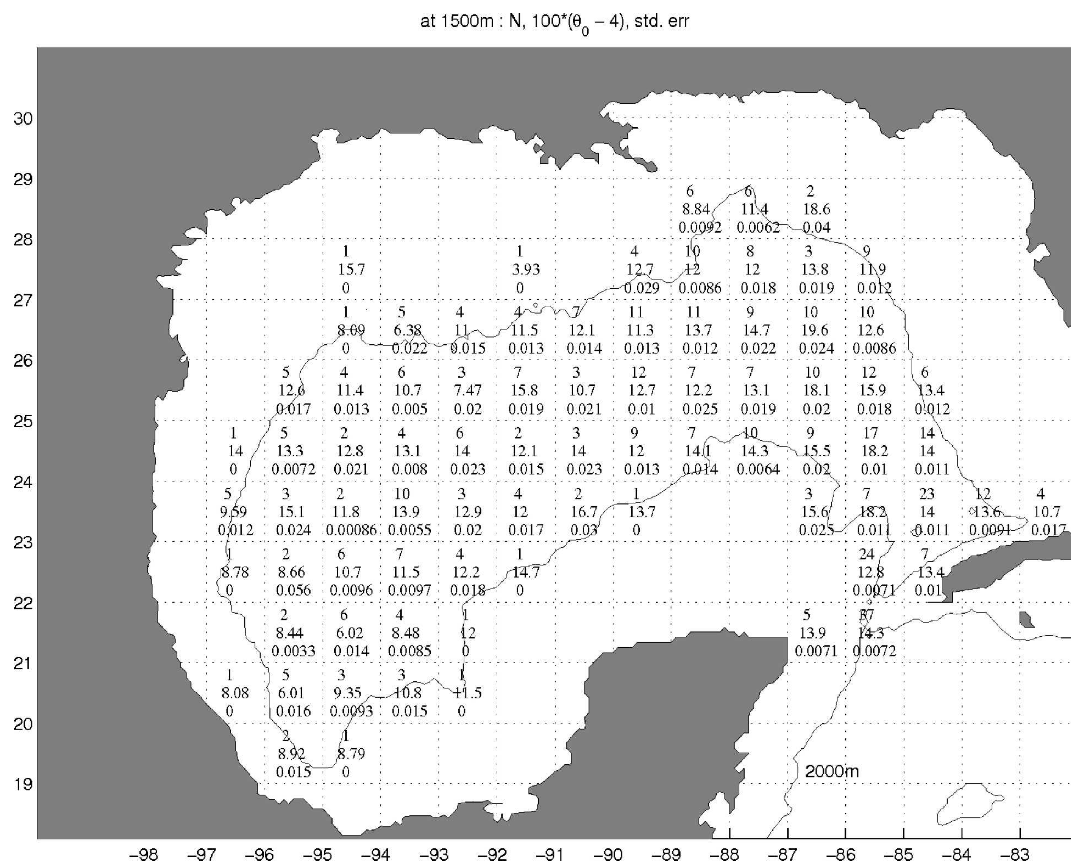

FIG. A1. Details of the potential temperature at 1500 m. In each 1° box all available historical data were averaged. The number of samples, the mean potential temperature, and the standard error are shown.

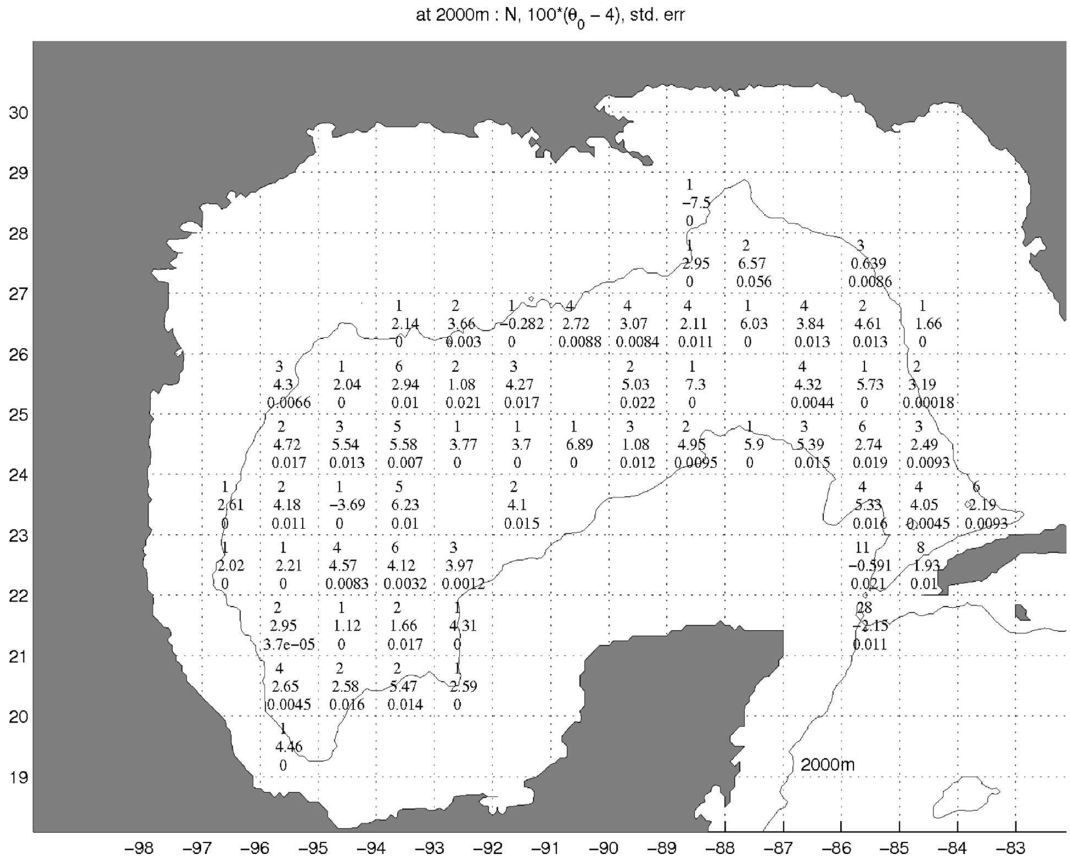

FIG. A2. As in Fig. A1 but for 2000 m.

samples, the mean temperature (in excess of 4°C and multiplied by 100; i.e., 4.123°C would show as 12.3) and the standard error of the mean. These values include all samples within 50 m above and below 1500 m. These figures show the data that were used to compile the data in Table 1. Figure A2 shows the equivalent result for 2000 m. Even at these depths, there is a warm central region in the east and in the west. There also is a cooler region along the west coast of Florida and along the deep U.S. coast from Louisiana to Texas. The data are so sparse that constructing temperature contours seems to require rather more imagination than science. Yet, as compiled in Table 1, the warm center and cool edges seem to be a real feature in the data.

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