Deep Currents in the Bay of Campeche

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ABSTRACT

Data from five moorings deployed in the Bay of Campeche during November 2007–July 2008 are used to analyze subinertial motions of waters below 1000-m depth. To the authors’ knowledge, this is the first time such a comprehensive observational program of direct deep-current measurements has been carried out in the region. The mean currents are in agreement with a cyclonic circulation at 1000-m depth; however, this cyclonic pattern is not so clearly defined at deeper levels. Only at the deepest mooring, located at 3500-m depth, are the mean currents uniform all the way to the bottom. Over the Bay of Campeche’s smooth western slope, currents show features compatible with topographic Rossby waves having vertical trapping scales thicker than 700 m, periods between 5 and 60 days, and horizontal wavelengths of 90–140 km. In contrast, the eastern slopes are characterized by rough topography, and motions with periods longer than 28 days decrease toward the bottom, suggesting a substantial reduction in the low-frequency topographic Rossby wave signal. Velocities from one of the two neighboring moorings located over the eastern rough slope have a strong 3-day period signal, which increases toward the bottom and has a vertical trapping scale of about 350 m. These higher frequency motions are interpreted in terms of edge waves.

1. Introduction

The Bay of Campeche (BOC) is the southwestern part of the Gulf of Mexico (GM), where the isobaths follow approximately a U shape with a smooth slope on the western side and a rough slope to the east (Fig. 1). The BOC ends in the west with a narrow shelf near the coast, its central and northern parts are connected with deep GM waters, and its south and east limits are characterized by shallow ample shelves less than 50 m deep. In this study, an unprecedented set of deep current measurements (described in the following section) is used to analyze the mean and subinertial current fluctuations below 1000-m depth.

A mean cyclonic circulation in the upper 1000 m of the BOC was well established by Vazquez de la Cerda et al. (2005). At 2000-m depth, however, there is some evidence that suggests a sluggish (1–2 cm s\(^{-1}\)) cyclonic mean circulation within the entire GM (DeHaan and Sturges 2005).

The subinertial deep currents in the northern GM have been amply observed and documented (Hamilton 1990; Donohue et al. 2008; Hamilton 2009). These studies have revealed topographic Rossby waves (TRWs) with characteristic periods between 25 and 100 days, wavelengths between 150 and 250 km, and propagation along the continental rise (Hamilton 1990, 2009). Near the Sigsbee Escarpment, in the northern central part of the GM, energetic bursts of TRWs trains of shorter periods (~10-day period) and wavelengths from 50 to 150 km have been reported (Hamilton and Lugo-Fernandez 2001; Hamilton 2007).

In the northwestern boundary of the GM, the observations of Hamilton (2009) suggest the interaction between the TRWs activity in the deep layer and an intense anticyclonic eddy detached from the Loop Current. These Loop Current eddies are known to cross the GM westward, where they interact with the western boundary (Smith 1986; Vukovich and Waddell 1991; Oey 1996; Sturges and Leben 2000; Ohlmann et al. 2001). Such eddy–topography interaction is thought to be a potential source of TRWs in the deep layer (Frolov et al. 2004). However, anticyclonic eddies rarely intrude into the BOC (Vukovich 2007); thus, their influence on the deep layers of the BOC is to some extent indirect. For instance,
TRWs generated north of BOC as a result of anticyclonic Loop Current eddies interacting with the slope and shelf are expected to propagate southward into the BOC.

The historical reference theory for TRWs is that of Rhines (1970). He derived the theory for free waves having motions that are unidirectional throughout the fluid, mainly subinertial, of a uniformly stratified ocean over a bottom with a weak uniform slope. These motions have an upper-limit, or cutoff, frequency set by the product of the Brunt–Väisälä frequency with the norm of the gradient of the bottom depth and are trapped in the sense that require a solid bottom from which the amplitudes decay toward the interior. TRWs are waves with periods of a week or longer, suitable over very small slopes, as described in Rhines (1970, case (iii), section 1.1). The historical observations of Thompson (1977) over site D in the North Atlantic western continental slope were probably the first to demonstrate the presence of TRWs. In the Atlantic Ocean at Cape Hatteras, Pickart (1995) showed that a coupling between meanders of the Gulf Stream and the lower layer could generate TRWs of about 40-day period along the continental slope. Similar results were described in numerical experiments of Malanotte-Rizzoli et al. (1995). Over the Grand Banks, Hogg (2000) showed that Gulf Stream displacements and warm-core rings in the neighborhood of the continental rise could also be sources of TRWs. For motions over not so small slopes, Rhines (1970) describes in case (ii), section 1.1, waves in the high, but still subinertial, frequency limit, referred as edge waves.

The BOC has a smooth sloping bottom on the western flank, but in its southeastern flank the bottom differs substantially with a rather uneven and rough continental slope. The study of Rhines (1970) describes in general the linear dynamics of motions over gentle and steep topography, which occur over a broad frequency band including edge waves and TRWs. However, as shown by Rhines and Bretherton (1973) for a homogeneous ocean and McWilliams (1974) for a stratified ocean, current variability over a rough bottom can generate topographic waves (TWs). TWs are oscillations with similar horizontal scale as the topographic roughness, decaying away from the bottom, and interact with oscillatory motions of the same frequency but with a larger trapping scale and horizontal scale comparable with the radius of deformation. Since the interaction occurs among oscillations with the same temporal variability, their distinction arises from their lateral scale and vertical decay. Hogg and Schmitz (1980), using measurements at the Charlie Gibbs Fracture Zone in the North Atlantic Ocean, found highly bottom-trapped subinertial motions (~3-day period) at higher frequencies than those allowed by the TRWs linear theory, a signal which they interpreted as TWs.

In this study we show current meter measurements that provide information on the mean and fluctuating currents over the deep BOC. The following section is a description of the observations and the main tools of the analysis. The third section deals specifically with the mean currents, and the fourth section with the characterization of the subinertial variability. Section 5 shows comparisons between observed and theoretically based approximations of the vertical intensification (or lack of it) as a function of frequency, average Brunt–Väisälä frequency, and bottom conditions (rough versus smooth). The main focus is on the extent that TRWs can explain the observed data. The final, section 6 contains a general discussion and conclusions.
2. Data and method

The data of this study comes from 27 instruments [9 acoustic Doppler current profilers (ADCPs) and 18 Aanderaa RCM11 point current meters] distributed on five moorings deployed from November 2007 to July 2008 (about 250 days, Fig. 1). Four of these moorings were nominally designed for 2000-m depth (M1, M2, M3, and M4), and one for 3500 m (M5). Table 1 gives the positions, instruments used, target and actual depths of the measurements, the temporal resolution of the series, and bin size for the ADCP measurements. There were no data gaps over the observational period.

All ADCPs were looking downward. For the five LR 75-kHz ADCPs (LR75), moored about 700 m below the surface, the bin nearest to the 1000-m depth was used for M1, M2, and M5, and the bins nearest to 1000-m- and 1150-m depths were used for M3 and M4. For the four near-bottom WH 600-kHz ADCPs (WH600), only 8.5–9.5 m (17–19 bins, depending on the mooring) of the profile have good data, and there are no near-bottom current measurements at M2. The side lobes contaminated the bins at 11 m below the bottom-looking ADCPs where a Microcat instrument was located, with intermittent signal losses also occurring for bins below these Microcats. The releases were located 3 m from the bottom, and, in the profiled data, there was no significant vertical shear, producing minor differences in the direction of the principal axis of standard deviation ellipses of about 0.5°. Therefore, the average of the 17 or 19 best quality bins, which span either 8.5 or 9.5 m of good profiled data, produced the time series used in the analysis of near-bottom currents. They are representative of currents at 10 m above the bottom. All moorings with the exception of M2 provided these time series (see Table 1).

The data series were filtered with a low-pass Lanczos filter with cutoff frequency of 0.5 cpd (or periods of 2 days, named 2-DLP time series). Thus, inertial (35-h
periods at 20°N) and main tidal frequencies are removed. The kinetic energy spectra were computed with the Thomson multitaper method (Percival and Walden 1993). The empirical orthogonal function analysis in the frequency domain (Wallace and Dickinson 1972; Denbo and Allen 1984; Hamilton 1990; Donohue et al. 2008; Hamilton 2009) allows the detection of phase lags within a mode (i.e., between the different vertical levels). Therefore, the distribution of phases in the vertical is explicitly determined. Instead of the usual covariance matrix for the EOFs in the time domain, the frequency domain EOF analysis is performed with the cross-spectrum matrix averaged within a user-selected frequency band of interest. The Fourier coefficients of the horizontal velocity components (\( U \) for the eastward and \( V \) for the northward) are first calculated. Then, the cross-spectrum matrix is constructed by averaging cross products of Fourier coefficients in the chosen frequency band. The Hermitian matrix thus produced has real eigenvalues and complex eigenvectors, giving the amplitude and phase of each time series included in the EOF synthesis. With the amplitude and phase for \( U \) and \( V \), in each frequency band, it is straightforward to compute the main features of the current variability ellipse (Emery and Thomson 2001).

The global ‘1’ grid bathymetry V12.1 provided by Smith and Sandwell (1997) was used in this study. Since the bathymetry is rough over some areas, a smoothed version is computed by averaging within a rectangular window of \( \frac{1}{2} \)° (about 52 km) per side. The smoothed topography is used to define the local alongslope and upslope directions and in general the topography gradient. The width of this window corresponds to the scale of the typical first baroclinic Rossby radius of deformation \( R = NH/f_0 \), with \( N \) the typical GM deep layer stratification (\( N = 0.0011 \) s\(^{-1}\), section 5a), \( f_0 \) the Coriolis parameter at the mean latitude of the GM (about 25°), and \( H \) the depth. With depths between 2000 and 3500 m, the deformation radii are between 36 and 62 km.

### 3. Mean currents

Figure 2 shows the mean velocities for 27 current time series between November 2007 and July 2008. Tables 2 and 3 provide the statistics of the 2-DLP time series. The effective degrees of freedom (EDoF, Table 3) are estimated, following Emery and Thomson (2001) and Hamilton (2009), using the velocity time series at each depth of measurement and calculating the autocorrelation time scale. The standard error of the mean is

\[
S_E = \sigma_e / \sqrt{\text{EDoF}},
\]

where \( \sigma_e \) is the standard deviation along the principal major axis of the ellipse.

The mean currents below 1600 m at M1 have a northward along-isobath direction. They intensify toward the bottom (Fig. 2 and Table 2). These measurements show a mean current of 5.8 cm s\(^{-1}\) at the deepest level (i.e., at about 10 m above the bottom) and 2.6 cm s\(^{-1}\) at 1801-m depth (Table 2). Above 1400 m, the deep mean currents reverse direction toward the south.

At M2, at depths from 1000 to 1800 m, the mean currents are southeastward along the bathymetry. Below 1500 m the mean currents at M2 differ over 110° in direction from those at M1. Here the speed of the mean current weakens toward the bottom from 5.8 cm s\(^{-1}\) at 1011 m to 2.4 cm s\(^{-1}\) at 1792 m (Table 2 and Fig. 2).

At M5, the mean currents from 1541 to 3048 m below the surface are nearly uniform in magnitude (between 3.5 and 3.9 cm s\(^{-1}\)) with a northeastward direction. A small clockwise rotation (\(~6°\) rotation) and intensification follows toward the bottom (4.6 cm s\(^{-1}\) at 3520 m below the surface and 10 m above the bottom). The topography gradient at this location is very weak; nonetheless, the velocity vector follows the isobaths (Fig. 2).

At M4 and M3, the mean velocity vectors are weak and decrease toward the bottom (less than 2 cm s\(^{-1}\) beneath 1300-m, Table 2). Note that the mean currents at M3 are insignificant at 1653- and 2108-m depth and at M4 at 1369-m and 1571-m depth (Tables 2, 3). However, at 1906 m at M3 and 2030 m at M4 (Table 2), the means are greater than the standard error with a southwestward orientation (Table 3), whereas they are northward and stronger near 1000-m depth (4.5 cm s\(^{-1}\) at M3).
At 1000 m below the surface the pattern of the mean currents from all five moorings is clearly cyclonic and along the bathymetry. At M5, the circulation below 1000 m is nearly vertically uniform and in agreement with the deep cyclonic circulation suggested by DeHaan and Sturges (2005).

In the BOC interior proper, south of M5, the vertical distribution from 1000 m below the surface to the bottom of the time-mean currents show an irregular pattern. In the western side (i.e., at M1 and M2), this circulation is more intense than at the eastern moorings (M3 and M4). At M1 the northward mean current increases toward the bottom, while at M2 it is less sheared and southeastward throughout the water column. The northward bottom currents at M1 and the southward weak or null bottom currents at M4 and M3 are in contradiction with a cyclonic deep circulation.

Furthermore, at both western moorings (M1 and M2), the mean current near the bottom remains relatively strong in comparison with the eastern moorings (M4 and M3). These last two moorings are located where the topography is very rough (Fig. 1), suggesting a near-bottom drag enhancement.

4. Variability of deep currents

a. Horizontal distribution of kinetic energy

Figure 3 shows the standard deviation ellipses of variability and mean currents, for the vertically averaged current, measured by the three Aanderaas at about

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**Table 2.** Basic statistics of 2-DLP data. For Reynolds stress (last column), the \( U \) and \( V \) components are in the rotated frame: \( U \) along bathymetry and \( V \) positive upslope. Asterisks indicate depth of the time series used for layer averaging. All the angles are measured positive from the zonal axis. Vel indicates the absolute value of velocity.

<table>
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<th>( \langle \text{Velocity} \rangle ) (cm s(^{-1}))</th>
<th>Orientation ( \langle \text{Vel} \rangle ) (°)</th>
<th>( \langle \text{Vel}^2 \rangle ) (cm(^2) s(^{-2}))</th>
<th>Principal axis orientation (°)</th>
<th>( \langle U V \rangle ) (cm(^2) s(^{-2}))</th>
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At 1000 m below the surface the pattern of the mean currents from all five moorings is clearly cyclonic and along the bathymetry. At M5, the circulation below 1000 m is nearly vertically uniform and in agreement with the deep cyclonic circulation suggested by DeHaan and Sturges (2005).

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Furthermore, at both western moorings (M1 and M2), the mean current near the bottom remains relatively strong in comparison with the eastern moorings (M4 and M3). These last two moorings are located where the topography is very rough (Fig. 1), suggesting a near-bottom drag enhancement.

4. Variability of deep currents

a. Horizontal distribution of kinetic energy

Figure 3 shows the standard deviation ellipses of variability and mean currents, for the vertically averaged current, measured by the three Aanderaas at about
1800 m, 1500 m, and 1300 m at M1, M2, M3, and M4 and with the whole series below 1000 m at M5 (the series used in the average are indicated by an asterisk in Table 2). The major axes of the ellipses are broadly aligned with the isobaths of the smoothed topography. The horizontal Reynolds stresses, in the rotated frame, are also calculated for each series (Table 2; i.e., with the $U$ component along the local isobaths and the $V$ component in the direction 90° to the left of the isobaths and upslope). The moorings located over the smoother western slope of BOC (M1 and M2) show the largest eddy kinetic energy signal. The momentum flux, which equals the Reynolds stresses, is negative at M1. It is also negative at M2 but with less significance (Table 2). As observed by Hamilton (1984) and according to the TRW theory (Rhines 1970), a negative momentum flux in the rotated frame indicates that the energy source is downslope from the location (i.e., an energy flux toward shallower waters), whereas positive values indicate an energy source situated upslope.

The velocity measurements of the deep currents in the eastern shelf of the BOC (M3 and M4) also show fluctuations aligned with the principal direction of the isobaths, but with less kinetic energy than at M1 and M2. At M5 the ellipse has the smallest eccentricity of the five locations—hence a larger dispersion in the velocity direction, although the principal direction of the current variability is along the bathymetry.

It is noticeable that the kinetic energy of the deepest 1000-m-thick layer is higher in the western BOC (M1 and M2) than in the eastern part (M3 and M4). The variability of the subinertial deep-current layer is clearly constrained by the bathymetry (principal directions of ellipses along the bathymetry). The variability at M5 shows weak low frequency oscillations in the 1000-m bottom layer, with sudden bursts that could be associated with the signature of a barotropic or deep eddy crossing by the mooring, as described by Hurlburt and Thompson (1982) and Welsh and Inoue (2000).

### Table 3. Time scales and effective degrees of freedom (EDoF) for velocity statistics.

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<tr>
<th>Mooring</th>
<th>Depth (m)</th>
<th>No. of data</th>
<th>Autocorrelation time scale (days)</th>
<th>EDoF</th>
<th>$S_e$ (cm s$^{-1}$)</th>
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$FIG. 3$. Standard deviation ellipses and mean of averaged currents (cm s$^{-1}$) in the layer between the 1000-m depth and the bottom (see text) computed from the 2-DLP time series.
The energy increase toward the bottom is also obvious in most frequencies. Energy peaks with energy enhancement near the bottom are also observed in the frequency bands corresponding to periods of 9–14 and 5–9 days. Shorter periods than 5 days also show a spectral peak with higher energy at 2000 m relative to other depths.

Figure 5a shows the deep currents, rotated in the bathymetry frame, at different levels in M2. The mean currents and fluctuations at those levels are oriented southeastward. The records show variability with time scales longer than a month with amplitudes close to 15 cm s\(^{-1}\) and an obvious high coherence between all levels. The currents remain rather uniform in direction and amplitude with depth. Unfortunately, owing to instrumental failure and in contrast with the other moorings, the very near-bottom currents were not observed at M2 (Table 1). At M2 a kinetic energy spectral peak common to all depth levels (Fig. 5b) shows dominant low frequency variability in periods from 30 to 60 days. In contrast with mooring M1, at mooring M2 there is a small amount of kinetic energy at periods shorter than 28 days (less than 2 cm\(^2\) s\(^{-2}\) in the variance preserving form of Fig. 5b), but there is relatively more energy at any given depth in the 30–60-day period band. For example, at the M2 1539-m depth the kinetic energy peak in the 30–60-day band is about 12 cm\(^2\) s\(^{-2}\), whereas, for a similar depth (1548 m) at mooring M1, the peak is less than 8 cm\(^2\) s\(^{-2}\). At M2 the kinetic energy spectra show a weak intensification of the currents toward the bottom (Fig. 5b).

Time series of the flows at mooring M4 on the eastern Campeche Bank continental rise are shown in Fig. 6a. The measurements at depths between 1000 and 1571 m (i.e., the top four series) show the highest variability with oscillations of \(~5\) cm s\(^{-1}\) in amplitude. The southwestward velocities of 5 cm s\(^{-1}\) amplitude occurring during...
May 2008 appear to be coherent in the whole water column and could be related to a deep eddy crossing the site. In general, there is no clear coherence between the top four levels and the deeper levels. What is noticeable at the two deepest levels, 1824 and 2030 m, is an energy increase in the high frequency band toward the bottom. The decrease (increase) of kinetic energy toward the bottom in the lowest (highest) frequencies is clearly seen in the kinetic energy spectra of Fig. 6b.

At M3 for periods longer than 17 days, the vertical distribution of deep currents is very similar to that of M4 (Fig. 7), which is closely located 57 km to the northeast (Fig. 1). Nonetheless, at M3 there is no clear current intensification toward the bottom at high frequencies, as occurs in mooring M4.

c. Vertical coherence in terms of EOFs

As described in section 2, the statistics of a set of horizontal velocity series in terms of EOFs in the frequency domain has a suitable representation in terms of ellipses of variability at each position. Following Hamilton (1990, 2007, 2009), this kind of EOF has been used as an effective tool for extracting TRW characteristics from current meter measurements vertically distributed in a mooring. Figure 8 shows the ellipses of the first EOFs for specific frequency bands computed from the velocity time series at moorings M1 (Fig. 4) and M2 (Fig. 5). The relative phase and direction of rotation in the ellipse is indicated by the arrow. For elongated ellipses the direction of rotation is quite uncertain, but also somewhat irrelevant.

The EOFs at mooring M1 in Fig. 8 show the columnar, coherent, and in-phase motions that are bottom intensified. The first EOF for the oscillations with 30–60-day periods explains 94.3% of the variance, with the elongated ellipses aligned with the isobaths. The first EOF for the 18–30-day periods explains 81.7% of the variance; it is also composed of elongated ellipses, but now with small clockwise and anticlockwise rotations of the semimajor axes.
relative to the local isobath. For higher frequency bands the motions are less rectilinear. In the frequency bands corresponding to the 9–14-day and 5–9-day periods, the first EOF explains 83.9% and 51.1% of the signal variance through the 1000-m deepest layer. For the frequency band of the 30–19-day and 5–9-day periods, the ellipses have anticlockwise deviations at 1000 m and at the deepest level—a discrepancy with respect to the TRWs unidirectional behavior. This is less clear in the case of the 30–60-day and 19–14-day period bands, which show the more coherent columnar behavior. Note that at depths between 1346 and 1801 m and for the higher frequencies, an obvious clockwise deviation of the ellipses major axis from the direction of the bathymetry is observed.

Figure 8 shows also the ellipses of the first EOF mode for the M2 mooring in the frequency band of 29–60-day period, which explains 95.1% of the variance. The first EOF shows the almost columnar, coherent, and near in-phase motion through the water column. The phases between the four depth levels of this first EOF differ by less than 34°. At this mooring, velocity intensification toward the bottom is small. Unfortunately, in this mooring the deepest measurement is 200 m above the bottom, unlike the others that have measurements 10 m above the bottom.

At moorings M3 and M4, so as to increase the vertical resolution in the upper 1000-m bottom layer, measurements from the 75-kHz ADCP located at about 1150 m below the surface have been added (Table 2). In the 28–60-day period band of the current fluctuations, their first EOFs are columnar and, in contrast with the EOFs from moorings M1 and M2, do not show intensification of currents toward the bottom. Maximum amplitudes are at about 1200-m and 1400-m depths for M3 and M4 and clearly decrease toward the bottom (Fig. 9). These first EOFs explain 67.2% of the variance at M3 and 81.7% at M4.

In the 2–5-day period band at M4, the variability of the currents, as shown in Fig. 10, is intensified very near the bottom and along the local isobath. The explained variance of the first mode in this frequency band is 50.1% at M4 and is nearly rectilinear, coherent, and in phase along the vertical.

5. Interpretation of the observed fluctuations

a. Background of topographic Rossby waves

The linear theory of Rhines (1970) is for motions fulfilling potential vorticity conservation for a uniformly stratified rotating fluid on a slope. This is the theory of TRWs that are the low frequency, and hence quasigeostrophic, limit of edge- or bottom-trapped subinertial waves. Some fundamental elements of the TRWs are shown below to be reproduced in the available data. The Rhines model has a rigid horizontal lid at $z = 0$ and a sloping bottom at $z = -H$. The Coriolis parameter is $f = 2\Omega \sin\phi$, where $\Omega$ is the earth’s angular velocity and $\phi$ the latitude. In the Rhines model, the Brunt–Väisälä frequency $N$ is taken as constant. Reid and Wang (2004) have extended the theory to the case of an exponential
profile of \( N \), but in this study only the case of constant \( N \) has been considered. Two full column CTD profiles, near 23\(^\circ\)N, 91\(^\circ\)W, where the depth is about 3500 m, provide the data for estimating \( N \) (Fig. 11a). The average between 1000-m and 2000-m depths, \( \bar{N} = (1.1 \pm 0.6) \times 10^{-3} \text{ s}^{-1} \), is used in the following estimations of wavelength and cutoff frequency. This value is assumed to be representative of the mean stratification between 1000 and 2000 m in the GM. The pressure perturbation vertical structure of TRWs in the \( f \)-plane model in the absence of latitudinal variation of \( f \) and for a constant \( N \) is given by

\[
P(z) = P_0 \cosh(\lambda z),
\]

where

\[
\lambda = NK/f
\]

and

\[
K = \sqrt{k^2 + l^2}
\]

is the magnitude of the wavenumber [i.e., of the vector \( \mathbf{K} = (k, l) \), attached to the dynamical variables of the linear equations of motion in the form \( \exp(\mathbf{K}(z_t - y_{w0})) \), where \( \omega \) is the frequency]. The dispersion relation is

\[
\omega = \frac{N^2(kh_x - lh_y)}{f \lambda \tanh(\lambda H)},
\]

where \((h_x, h_y)\) is the gradient of the bottom depth. It is worth pointing out that wavelength is not a strong function of frequency for TRWs (Oey and Lee 2002). When \( \lambda H \gtrsim O(2) \) and \( \coth(\lambda H) \sim 1 \), the dispersion relation can be approximated by

\[
\omega \approx N|Vh| \sin(\theta),
\]

\( \theta \) being the clockwise angle that the wavenumber vector makes with the bathymetry gradient (Oey and Lee 2002). In any case, as remarked by Thompson (1977), \( \coth(\lambda H) \equiv 1 \); hence, from (5), the inequality

\[
\omega \geq |N \forall h \sin(\theta)|
\]

is always satisfied. Moreover (5), with the assumption \( \coth(\lambda H) \equiv 1 \), implies that

\[
C_g = \frac{N|Vh| \cos(\theta, -k)}{K^2},
\]

where \( C_g \) is the group velocity. As stated by Oey and Lee (2002), \( C_g \) is directed 90\(^\circ\) clockwise (upslope) with respect to \( \mathbf{K} \) when it points downslope (i.e., \( 0 < \theta < \pi/2 \)) and 90\(^\circ\) anticlockwise (downslope) when \( \mathbf{K} \) points upslope (i.e., \( \pi/2 < \theta < \pi \)). This means that looking upslope the TRWs packets always travel leftward in the Northern Hemisphere.
The topography has been taken from Smith and Sandwell (1997) (global 19; V12.1). The bathymetry has been smoothed with a running-mean window of ½ in longitude by ½ in latitude (52 km per side), which is a scale comparable with the first internal Rossby radius of deformation (i.e., NH/f). The bathymetric gradient has then been computed (Fig. 11b) with the smoothed topography and the cutoff period plotted with \( N = 1.1 \times 10^{-3} \text{ s}^{-1} \) [see (6) and Fig. 11c]. Along the western continental rise of the BOC, at the locations of M1 and M2, the cutoff period is relatively short, between 3 and 6 days (with a large uncertainty of ±5 days, owing to the mean stratification uncertainty, which is ±0.6 × 10^{-4} \text{ s}^{-1}). Figure 11 is in agreement with Fig. 7 of Oey and Lee (2002). For the locations of M3 and M4, the topographic gradient is weaker and the cutoff period increases to 6–9 days. Between the western and eastern rises, within a southward-oriented tongue reaching 20°N, the cutoff frequency decreases substantially, preventing the existence of TRWs with periods shorter than 12 days.

b. Comparison with theory and other studies

The trapping scale can be found by least squares fitting the semimajor axis amplitudes as a function of depth using (2). Fits of this kind are shown in Fig. 12a. Hamilton (1990) found the wavelengths and directions of presumable TRWs in the northern GM by computing phase differences between separate moorings in his data. Unfortunately, because of the large separations in our moorings, direct estimation of direction and wavelength, as in Hamilton, is not feasible. However, knowing the trapping scale, the wavelength of TRWs can be estimated from (3), and the wavenumber direction from (6).

1) VERTICAL STRUCTURE

At M1 the characteristics of the first EOF are compatible with theoretical TRWs. Figure 12a shows the ellipse semimajor axes vertical distributions from the first EOFs of four frequency bands. Fitting the observed vertical structure according to (2) implies decaying scales between 1083 and 724 m \( (lH ; 2 \text{ or larger}) \). Wavelengths, according to (3) are 97 and 145 km for the shortest and longest trapping scales, respectively (Fig. 12a). The uncertainty in \( N \) leads to a large uncertainty in wavelengths of ±53 and ±79 km for the shortest and longest trapping scales. From Fig. 12 no trend can be observed linking the trapping scale with frequency, suggesting that wavelengths are not in any way ordered by frequency. This independence is compatible with the dispersion relation (7) and with the results of Oey and Lee (2002).

Comparing our results with those from the study of Hamilton (1990) in the northern GM and focusing on the lowest frequency bands corresponding to periods of 30–60 days and 17–30 days, the motions at M1 are more trapped toward the bottom or equivalently [see Eqs. (1) and (2)] have shorter wavelengths. Hamilton found wavelengths on the order of 121–297 km for fluctuations with 38–300-day periods and wavelengths of 110–220 km for the 18–37-day period band. At M1 the analysis indicates wavelengths of 97–139 km for fluctuations with 5–70-day periods. But, for oscillations with periods of 5–30 days, near the Sigsbee Escarpment and also in the northern GM, Hamilton (2007) estimated TRWs wavelengths between 75 and 150 km, and these values are similar to those found at mooring M1. Recent studies (Donohue et al.
2008; Hamilton 2009) in the northwestern corner of the GM, just north of the BOC, have reported TRWs with
wavelengths between 75 and 135 km for fluctuations with periods from 23 to 66 days, which are comparable
with the motions at M1.

The deepest measurements at M1 of each frequency
band signal show a slight discrepancy with the expected
TRWs vertically uniform directionality. At M1, the prin-
cipal direction of this ellipse is rotated anticlockwise in
comparison with the shallower levels (Fig. 8). Like in
Donohue et al. (2008), this can be interpreted as evidence
of a bottom Ekman layer in which the bottom friction
deviates the current to the left.

2) THE RELATIONSHIP BETWEEN VELOCITY
DIRECTION AND FREQUENCY

The quasigeostrophic theory requires transverse (i.e.,
velocities perpendicular to wave crests) and rectilinear
motions. This implies a null or 180° phase lag between U
and V components (in any orthogonal reference sys-
tem), a characteristic that is closely, but not completely,
fulfilled as shown by the elongated ellipses. In the limiting
form of the dispersion relation, shown by (6), ω is pro-
portional to sinθ, which implies that for very low fre-
quencies the wavenumber vectors are nearly parallel to
Vh (i.e., wave crests and motions are almost along iso-
baths) and for the maximum or cutoff frequency (N|Vh|)
wavenumber vectors are parallel to isobaths (i.e., motions
in the direction of maximum column stretching or
shrinking). In Fig. 12b, we compare the angles between
isobaths and the principal axis of the first EOFs ellipses at
M1 and M2 and the theoretical angle, given the frequency
[i.e., θ in (6)].

The discontinuous trace in Fig. 12b depicts the angle
as a function of frequency using (6) and |Vh| = 0.014 and
N = 1.1 × 10^{-3} s^{-1}, which are the values for M1. The
circles represent the angles for the EOF ellipses of all
depths with the exception of the deepest (about 10 m
above the bottom) with an asterisk. These last show the
possible deviations in the flow direction due to Ekman
frictional effects. The middle frequency of the corre-
sponding band is used to pair frequency–angle values.
These values fit well the theoretical curve and are

\[ T = \frac{2\pi}{N|Vh|} \]

computed from topographic gradient with
N = 1.1 × 10^{-3} s^{-1}.
compatible with the inequality given by (7), although with a large dispersion of the values, particularly for the 30–17-day band. The major axis of the deepest ellipses (asterisks in Fig. 12b) show a consistently counterclockwise deviation of the angle, but it is small and does not alter the results in a significant way. Positive angles are clockwise between the local isobath with the shallower side on its right and the nearest major axis of the ellipse. The negative (positive) angle implies a downslope (up-slope) group velocity; see (8). This is in agreement with the calculation of the momentum flux (Table 2); negative momentum fluxes in the rotated frame indicate that the energy source is downslope.

Figure 12b also shows the angle versus frequency comparison at mooring M2, although here $|Vh| = 0.025$ and $N$ is the same as for M1. The squares represent the angles for the M2 ellipses and are labeled following the depth of the measurements. The presence of only one energy peak in the frequency spectra and the large dispersion of angles at M2 imply a poor fit to (6). Furthermore the sign of the angle varies with depth: positive at 1539 m, nearly null at 1792 m, and negative above 1337 m. The large clockwise deviation at 1011 m relative to deeper levels could be related to upper-layer activity. Hence, the direction of group velocity whether up or downslope at M2 is quite uncertain.

The motions in several low frequency bands at M1 are compatible with TRWs dynamics. At M2, the compatibility is only qualitative (i.e., intensification toward the bottom and motion nearly along bathymetry). Among other factors, friction, mean currents, and vertical shears can make the measured value deviate from the linear theoretical value. Even with the large uncertainty in the calculations and the lack of a direct estimation of wave-numbers, the deep current fluctuations in this part of the GM show, perhaps surprisingly, substantial compatibility with linear TRWs, especially at the M1 site.

3) ENERGY PROPAGATION AND PATH

Some evidence has been presented in the last section suggesting the presence of TRWs at moorings M1 and M2. Because TRW energy flux must have a southward component along the western continental rise of the BOC, some propagation occurring from M1 to M2 is expected. To check this, lagged correlation calculations have been computed among the along-bathymetry components of the 2-DLP times series for depths between 1000 and 1800 m at the two moorings. The time series of along-bathymetry velocity components at M1 and M2 have a maximum correlation of $0.71$, with M2 lagging M1 by 5.3 days (Fig. 13b). Figure 13a shows filtered series at about 1500 m for both, where the series at M1 has been lagged 5.3 days. The same computation has been done between the velocity at approximately 1000-m, 1300-m, and 1800-m depths (figures not shown), revealing similar maximum correlations (from 0.65 to 0.74), all of which are statistically significant and at comparable lags (M1 lagging behind M2 from 3.6 to 5.5 days). Note that the quasiperiodic shape of the lagged correlation function shown in Fig. 13b indicates that the major contributions are the oscillations of about 1–2-month periods.

Since the correlations are significant at each depth level, coherence and phase difference (Fig. 13c) have been computed by averaging the four cross-spectra of the along-bathymetry velocity components at M1 and M2. The maximum correlation of $+0.71$, with M2 lagging M1 by 5.3 days (Fig. 13b). Figure 13a shows filtered series at about 1500 m for both, where the series at M1 has been lagged 5.3 days. The same computation has been done between the velocity at approximately 1000-m, 1300-m, and 1800-m depths (figures not shown), revealing similar maximum correlations (from 0.65 to 0.74), all of which are statistically significant and at comparable lags (M1 lagging behind M2 from 3.6 to 5.5 days). Note that the quasiperiodic shape of the lagged correlation function shown in Fig. 13b indicates that the major contributions are the oscillations of about 1–2-month periods.
coherence magnitude is +0.78 at the 40-day period, and
the second maximum is at a 16-day period with +0.66.
The phase difference is −47° at 40 days, which translates
into a 5.2-day lag, whereas phase differences of −80° at
16 days are found, which translates into a 3.5-day lag.

These correlations and the coherence are compatible
with the expected propagation of TRWs of about 40-day
periods from mooring M1 to M2. Using a distance between
M1 and M2 of 175 km, it is straightforward to estimate the
group velocity of the 40-day wave packet as being about
34 km day −1, which is a lower bound on the group ve-
cocity. From (5) and for fluctuations nearly aligned with
bathymetry θ ≪ 1 (about 5°, Fig. 8) and k/K ≪ 1, the group
velocity can be estimated at M1 and M2 as being 30 ±
16 and 53 ± 27 km day −1, respectively—velocities com-
patible with our direct estimation from correlation and co-
herence analysis. The 10–20-day period band that shows
significant coherence is much less energetic at M2 (Fig. 5b).

Figure 14 shows the forward (in red) and backward (in
black) energy path computed using the linear dispersion
relation (5), group velocity (8), and estimate of K at M1.
The method has been often used in the literature (Pickart
1995; Oey and Lee 2002; Donohue et al. 2008; Hamilton
2007, 2009) for calculating ray paths of TRWs in the Mid-
Atlantic Bight and northern GM. Under the WKB ap-
proximation, the equations governing the evolution of
a wave packet and wavenumber are (LeBlond and Mysack
1978)

\[
D_t x = \frac{\partial \omega}{\partial k} = C_g
\]

and

\[
D_t k = \sum - \frac{\partial \omega}{\partial \gamma_i} \nu_i
\]
The forward path (in red) indicates that the energy follows the continental slope up to M2 in a little more than 5.5 days (more than five wavenumber vectors), a value that is compatible with the lag found in the coherence analysis above. Furthermore, the path remains confined at the base of the prominent escarpment (in lighter gray) and accelerates (larger spacing between arrows) between M1 and M2. According to the WKB approximation, to conserve frequency the group velocity increases its alignment with isobaths as the slope increases. Oey et al. (2009) suggests focusing of energy flux coming from the deep ocean below the zone of maximum slope over the continental slope. In this case, the TRW energy may be concentrated along the 2000-m isobaths, as suggested by the higher kinetic energy content at M2 relative to M1 within the 30–60-day period band (cf. the spectra at 1550-m depth in Figs. 4b, 5b).

The mode structure at M1 for the ~40 day period band clearly suggests the southward propagation of TRWs toward M2, as depicted by the ray calculation. However, questions on the origin and fate of such TRWs in the BOC remain to be addressed, particularly the dramatic difference of the vertical structure of the ~40 day period oscillations over the rough eastern slope in comparison with the smooth western slope.

c. Topographic waves over a rough bottom

We will now focus on the high frequency 3-day period oscillations observed in the eastern rough part of the BOC, where lower frequency motions decrease in kinetic energy toward the bottom. Figure 15a shows the velocities at M4 filtered using a bandpass filter centered on a 3-day period for the two deepest time series at 1824-m and 2030-m depths (bottom) and the 2-DLP time series at 1011 m (top). The figure shows bursts of 3-day period oscillations, mainly rotating counterclockwise, at the deepest levels with quasi-monthly modulations and amplitudes of ~3 cm s⁻¹. These trains of waves decay with a scale of about 10 days. They are particularly intense between March and June 2008 and occur simultaneously with larger variability of the currents at 1011 m. In May 2008 the intensification occurred over the whole water column, as reported in section 4b and Fig. 6.

Figure 15b shows the vertical structure of the first EOF for the 2–5-day period band, which corresponds to a peak in kinetic energy spectra (Fig. 6). These oscillations, near inertial and vertically coherent, are observed to be strongly bottom trapped only at M4. The trapping scale is 342 m, shorter than the scales associated with TRWs in Fig. 12a, and indicative of a shorter horizontal wavelength.

Rhines and Bretherton (1973) and McWilliams (1974) discuss how a corrugated bottom (with or without a mean slope) produces a coupling of highly bottom-trapped
oscillating motions having the lateral extent of the bottom corrugations with larger-scale quasigeostrophic motions of the same frequency. These highly trapped oscillations are referred to as topographic waves (TWs) and are confined to the bottom layer (McWilliams 1974; Hogg and Schmitz 1980).

The smallest vertical trapping scale for TWs forced by bottom corrugation is given by the ratio of ocean depth ($H = 2000$ m) with the Burger number. The Burger number is $B = (2\pi NH)/(fL_T)$ in which $L_T$ is the lateral scale of the corrugations. We assume $L_T = 20$ km (see Fig. 10) and $N = 5 \times 10^{-4}$ s$^{-1}$ at this location, which is the mean value of the Brunt–Väisälä frequency 400 m above the bottom. This value has been chosen because the motion occurs obviously in the 400-m near-bottom layer (trapping scale of 342 m). These parameter values yield $B \approx 6.3$ and a trapping scale of $H/B = 317$ m. This scale is consistent with the result of first EOF at M4 shown in Fig. 15b (342 m). However, consistency is not complete since we only observed the small-scale motion described in the theory without the presence of the large-scale quasigeostrophic motions of the same frequency. Furthermore, the trapping scale is of the same order as the vertical scale of the roughness elements (about 250 m; Fig. 10). These latter scales are not small enough for the theory to be applicable with confidence.

A more plausible explanation is that these near but still subinertial oscillations are edge waves within a bowl-shaped depression with large topography gradients, thus allowing motion at higher frequencies close to inertial ($\omega \sim f$). At M4, we have locally $\omega h > 0.05 \sim f/N$. As described in Rhines [1970; his case ii, see section 1.1], there are edge waves that are degenerate cases of TRWs in the limit of steep topography. These have near-inertial periods, and the wavelengths can be estimated with the vertical trapping scale in the same way as the TRWs [i.e., via Eq. (3); Rhines 1970]. The reason why edge waves are better suited to fit the observations at M4 is that there is no evidence of the large horizontal scale motions required for TWs. All of the data from the other four moorings (M1, M2, M3, and M5) show no motions alike. In any case, highly bottom-trapped motions exist at one location above the rough bottom over the continental slope of the Campeche Bank. The effect of rough bathymetry on the deep circulation remains to be fully addressed.

6. Discussion and conclusions

In this study we analyze the mean and subinertial currents of a layer straddling from 1000 m below the surface to the bottom in the Bay of Campeche. The observations come from the initial recovery of five moorings that were deployed for 9 months (November 2007–July 2008) in the region. Although for the processes studied here the spatial separation between moorings is large and the temporal length of the measurements is a limiting factor, the good vertical resolution of the measurement allows detection of basic features and scales of the variability.

At 1000-m depth a cyclonic circulation throughout the basin is observed. Below 1000 m, the mean currents do not show a well-defined circulation pattern. Only at mooring M5 (see Fig. 2) are the mean currents in the same direction and without significant shear. The rest of the moorings show the existence of relatively strong shear in the mean currents, with weakening toward the bottom at moorings M3 and M4 and even reversal at M1. The deep mean circulation agrees with the results of DeHaan and Sturges (2005) at M5, but not so clearly in the interior of the BOC.
DeHaan and Sturges suggest that topographic rectification (Huthnance 1981) can occur near the bottom; that is, TRWs and enhanced bottom friction can produce a net cyclonic mean mass transport in the bottom layer in the GM (below 1000 m). Mizuta and Hogg (2004) have investigated the up-slope propagation of TRWs onto an increasing steep slope. Their calculations show that a rectification process occurs and a mean flow—nearly vertically uniform—develops over the slope with deeper water on its left side. The rectification process is created by the divergence of wave Reynolds stresses in the bottom boundary layer. A mean current with (plausibly) similar conditions is observed at mooring M2 (see Fig. 2). Intense vertical shear and northward reversal of the mean flow in the vicinity of the bottom at M1 remain unexplained. The mechanisms that drive mean currents in the BOC deserve more investigations and observations.

EOFs in the frequency domain and for bands where spectral energy peaks are observed indicate fluctuations that are vertically coherent and unidirectional (see Figs. 8 and 9), a structure consistent with low-frequency ocean waves. The method eliminates the incoherent fluctuations and provides phases and amplitude estimates as a function of depth (Wallace and Dickinson 1972; Denbo and Allen 1984). Only the first EOF exceeding 50% of explained variance has been discussed in this study. This statistical method does not warrant that the resulting structures or modes represent a physical process, and several physical processes can also be mixed in one EOF (Wallace and Dickinson 1972). For example, vertically coherent currents due to waves and eddies traveling across the mooring can contribute to the EOF structure complicating their interpretation as a well defined wave pattern (Donohue et al. 2008). This issue may be problematic in our case, owing to the shortness of the available time series and it could explain, for example, the lack of agreement between the first EOF with the theoretical TRW mode at M1 in the 18–30-day-period band or at M2 for the 1/40 cpd oscillations [see section 5b(2) and Fig. 12b].

At mooring M1, over the western side of the BOC, several spectral peaks were found in the low frequency band with the largest around the ~40-day period. EOF calculations on frequency bands determined by the presence of four energetic spectral peaks show vertically coherent structures captured in the corresponding first EOF and the intensification toward the bottom of currents with similar orientation and phase at different depths, in fair agreement with the theory of TRWs (see sections 4 and 5, Figs. 8 and 9). The wavelengths, estimated via their theoretical relation with the vertical scale of observed intensification, are comparable with the 40–250-km wavelengths of similar motions found in the northern Gulf of Mexico (Hamilton 1990, 2007, 2009; Donohue et al. 2008).

The hydrographic profile for defining N was measured during August 2004 near 23°N, 91°W (Fig. 11). The expected difference with values in the area and time of interest is negligibly small because our focus is on the deep layers. A discrepancy in the theoretical modes, in particular their increase toward the bottom, might arise from the use of a uniform or constant value for N rather than a realistic profile. The averages of N from specific depths were used in the analysis. As shown by Reid and Wang (2004), an exponential decay of N, modeled from a profile very much in agreement with the one shown in Fig. 11a, modifies the vertical intensification of TRWs. We also do not take into account in our estimates the effect of a mean current or its vertical shear, which are factors to consider in a more comprehensive comparison with linear theory (Rhines 1970; Oey and Lee 2002). These processes might explain the weaker intensification toward the bottom at mooring M2 relative to that at M1 for the 29–60-day period band and the lack of intensification in other frequency bands at M2. Our decision to use the simplest theory available to fit the observations had precisely the purpose of highlighting the locations and processes where such a simple theory breaks down.

The distance between moorings M1 and M2 is 174 km, and the currents along the 2000-m isobath at each mooring show significant coherence in the 40- and 16-day-period bands. The lagged correlations among these series reach a maximum with M2 lagging 5.3 days behind M1 (Fig. 13b). The theoretical energy path for 40-day period TRWs, starting at M1 with the values suggested by theoretical considerations, goes toward the southern corner of BOC, approaching the M2 site and the ray reaching the closest point of M2 approximately 5.5 days later. The mean deep current, horizontal and vertical variations of N (not considered here), and the noticeable large curvature of isobaths of the BOC may seriously limit the applicability of the WKB hypothesis. Therefore, one should bear in mind that, although the southern propagation of energy at M1 appears realistic, paths built in BOC via ray tracing with the WKB hypothesis might be a misleading interpretation of reality (Oey and Lee 2002).

Donohue et al. (2008) have traced the ray path of the low-frequency TRWs (66 days) from the northwestern corner of the GM into the BOC (their Figs. 4.4–5), suggesting a possible pathway for TRWs. These results suggest that TRWs can follow the continental shelf from the northern Gulf of Mexico, where eddy activity (Vukovich 2007) is supposed to excite free TRWs (Hamilton 2009). Sutyrin et al. (2003) has shown in a two-layer model that the interaction between baroclinic surface-intensified anticyclonic eddies and a western continental slope and
shelf is able to induce vorticity perturbations that disperse in the form of TRWs in the deepest layer. The generation mechanism of the observed 40-day periods TRWs at moorings M1 and M2 clearly deserves further investigation.

There is a notorious weakening of observed kinetic energy from moorings M1 and M2 to moorings M3 and M4 (Fig. 3). The EOFs at M3 and M4 show in-phase fluctuations, decreasing toward the bottom with almost a node at the bottom at periods longer than one month. It appears that the main difference between the eastern and western BOC slopes is the roughness of the bottom. On the other hand, even over a rough sloped bottom, TRWs could dominate the low-frequency motion, but the roughness of the bottom may dramatically increase the bottom drag and may thicken the bottom boundary layer, leading to a modification of the vertical structure of TRWs.

The motions at M4 in the spectral peak with periodicities from 2 to 5 days are intensified toward the bottom, in close vicinity, and are not observed at any other of the moorings. At M4, the allowed TRWs must have periodicities longer than 6 days (Fig. 11), an estimate obtained using a smooth topography proper for motions whose lateral scale is about or larger than the first baroclinic radius of deformation. Similar observations over rough topography have been discussed by Hogg and Schmitz (1980) and interpreted as a different kind of topographic waves (TWs), but the concomitant larger scale oscillations required in the theory are not observed at M4. We interpret these high frequency oscillations as transient edge waves over locally steep topography, as described in Rhines (1970). Our lack of a high-resolution realistic topography inhibits further analysis to check agreement with this theory.

To conclude, within the deep southern BOC a kinetic energy spectral peak of periodicities between 30 and 60 days decreases eastward in amplitude (Figs. 4–7). A large fraction of these motions at moorings M1 and M2 is unidirectional, vertically coherent in phase, and orderly intensive toward the bottom, hence interpreted as TRWs. A similar structure occurs in fluctuating currents at M1 with periodicities between 5 to 60 days and, following theoretical guidelines, we estimate that these fluctuations have horizontal wavelengths between 90 and 140 km. In contrast, the measurements over the rough topography, at moorings M3 and M4, show a clear decline of kinetic energy toward the bottom. A striking exception, with kinetic energy increasing toward the bottom occurs at mooring M4 for motions with spectral peak in the 2 to 5 day band. These oscillations are beyond the high frequency cutoff for TRWs and are interpreted as edge waves supported by a steep local bathymetry.

Several issues require further analysis, but three clearly stand up: (i) the eastward change of vertical structure across the BOC from bottom intensified motion to upward intensified motion, (ii) the energy gap for periods shorter than 28 days at the southern and eastern moorings (M2, M3, M4), and (iii) the generation mechanism of the TRWs that propagate into the western BOC. Fortunately, more data will come in the future from this observational program, which hopefully will shed light on the dynamics of this interesting region.

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REFERENCES