Estimating ocean transports with dynamic height moorings: An application in the Atlantic Deep Western Boundary Current at 26°N

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Abstract

Efficient monitoring of large-scale current systems for climate research requires the development of new techniques to estimate ocean transports. Here, a methodology for continuous estimation of dynamic height profiles and geostrophic currents from moored temperature sensors is presented. The technique is applied to moorings deployed in the Atlantic Deep Western Boundary Current at 26.5°N, off Abaco, the Bahamas (WOCE ACM-1 array). Relative geostrophic currents are referenced using bottom pressure sensors and available shipboard direct velocity (lowered-ADCP) sections over the period of the deployment, to obtain a time series of absolute volume transport. Comparison with direct velocity measurements from a complete array of current meters shows good agreement for the mean transport and its variability on time scales longer than 10 days, but larger variability in the current meter derived transport at time scales shorter than 10 days. A rigorous error analysis assesses the contributions of various error sources in the geostrophic as well as direct transport estimates. Low-frequency drift of the bottom pressure sensors is found to be the largest error source in the geostrophic transport estimates and recommendations for improvement of the technique and related measurement technologies are made.

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1. Introduction

The purpose of this paper is to present the methodology and a practical application of techniques for monitoring geostrophic volume transports in the ocean through moored observations of dynamic height and bottom pressure. In
principle the concept of using moored observations of temperature (and if available, salinity) to continuously monitor the dynamic height profile at a given location is straightforward (Stommel, 1947). However, it has been little used in modern oceanographic studies. Zantopp and Leaman (1984) studied the feasibility of using moored temperature measurements to determine dynamic height profiles in the western North Atlantic and found that the technique worked quite well. The only essential requirement for this technique to work is that accuracy and vertical sampling of the temperature and salinity measurements are sufficient to resolve the basic vertical structure of the density field. In the case where only temperature measurements are available, the $T/S$ relation in the region must additionally be stable and tight.

In the application presented here, pairs of such “dynamic height” moorings are used to estimate the spatially averaged geostrophic velocity profile and associated transports over relatively wide mooring separations. This technique contrasts with conventional moored current meter array measurements which require the horizontal, as well as vertical, structure of the current field to be sufficiently well resolved to estimate transports. The dynamic height method requires measurements on both sides of the current field only, and thereby obtains an integral picture of transport itself, making it ideally suited for climate monitoring purposes. Whitworth (1983) used this technique to estimate the baroclinic transport of the Antarctic Circumpolar Current through Drake Passage. Johns et al. (2001) successfully applied it in the Kuroshio off Taiwan and showed that a time series of geostrophic transport derived from dynamic height moorings on each side of the Kuroshio agreed well with direct transport estimates from an array of current meters across the current. Kanzow (2000) evaluated the technique for potential application in monitoring basinwide meridional overturning circulations with promising results.

We describe here the methods used to estimate dynamic height profiles and geostrophic currents from moored temperature records. The resulting geostrophic transport estimates are compared with direct transport estimates from a more spatially dense array of moored current meters. The use of bottom pressure sensors to determine the variability in the near bottom currents is also evaluated, including methods to calibrate the bottom pressure fluctuations for absolute currents. This provides a means to reference the relative geostrophic current profiles and obtain absolute transports. Finally, errors in both the dynamic height and direct transport methods are assessed to determine the relative accuracies and merits of each technique. The results show that the dynamic height approach in many cases should be able to provide transport accuracies comparable to that of moored current meter arrays.

2. Experiment and data

The data described in this paper were collected in the US WOCE western boundary current array for the subtropical Atlantic (ACM-1) at 26.5°N east of Abaco, the Bahamas. The data set is also known as ACCP-3, the third, and most recent, deployment of the ACM-1 array under the auspices of the Atlantic Climate Change Program, and is described in Zantopp et al. (1998). The moored array was deployed from October 1995 to June 1997, and was intended to directly measure the volume and heat transports of important components of the wind driven and thermohaline circulation in the western subtropical Atlantic. These components include the Atlantic Deep Western Boundary Current (DWBC), which flows equatorward at depths greater than 1000 m with a core near 2000 m at this location, and the poleward flowing Antilles Current in the upper ocean. For further information on the currents in this region and previous results from the ACM-1 array, the reader is referred to Lee et al. (1990, 1996) and Fillenbaum et al. (1997).

The ACCP-3 array consisted of five subsurface, taut-wire current meter moorings along 26.5°N with water depths between 1000 and 5300 m (Figs. 1 and 2). The moorings had current meters (EG&G Vector Averaging Current Meters; VACMs) at nominal depths of 100, 400, 800, 1200, 2000, 3000, 4000, and 5000 m (at mooring E only), measuring temperature and currents, and
additional temperature recorders in between (250, 600, 1000, 1600, 2500, 3500, and 4500 m). The easternmost moorings (D and E) reached up to 800 m only. Each of the 4 deep moorings had a bottom-mounted CTD with a high-precision (Paroscientific Digiquartz) pressure sensor. Mooring B (deep mooring closest to the Bahamas escarpment, and closest to the core of the DWBC) lost all flotation and instrumentation at depths above 1000 m after 15 months through the experiment (in late January 1997). We therefore restrict the analysis in this paper to the measurements collected between 1200 m and the bottom

(Fig. 2). Also, due to excessive draw down of mooring B after the loss of its surface flotation, most of the analysis and comparison of methods is limited to the first 15 months of the experiment, from October 1995 to January 1997.

Prior to the calculations described below, all of the data (moored currents, temperatures, and pressures) were low-pass filtered with a 40-h Lanczos filter to eliminate tidal and inertial variations. The estimated uncertainties in measured currents are ±2 cm/s and the temperatures from all sensors have a calibrated accuracy of better than ±0.01 °C. The bottom pressure sensors have a resolution of 0.001 dbar with an absolute accuracy of 0.02% of their full scale pressure rating (7000 dbar), or approximately 1.4 dbar.

3. Methods

We will first describe the methodology used to obtain dynamic height profiles from discrete moored temperature sensors. Once obtained, these profiles are used to estimate spatially averaged, relative geostrophic velocity profiles between moorings. We then go on to describe the methodology for determining spatially averaged near bottom velocities from bottom pressure sensors, including calibration techniques for absolute velocity.

3.1. Dynamic height profiles and relative geostrophic currents

The initial step in estimating dynamic height profiles at the moorings is the determination of the actual depths of all moored sensors as a function of time. Each temperature sensor on the moorings has a nominal depth for near-zero velocity conditions. However, due to frictional drag induced by ocean currents, all mooring components are periodically displaced from their nominal depths. On the ACCP-3 moorings, pressure sensors were installed at two or more levels between the surface and 1200 m to provide a measure of these vertical displacements. The r.m.s. vertical displacements of the moorings at the 1200 m level ranged from 65 m at mooring B.
(before loss of its surface flotation) to less than 3 m at mooring E. Maximum vertical displacements reached up to 200–300 m during extreme events, but as high as 900 m at mooring B after the loss of its surface flotation.

As a first approximation we assume the moorings tilt over with a nearly uniform inclination. The mean angle of mooring inclination \( \theta(t) \) is then determined as follows: the pressure record at 1200 m, \( p_{1200}(t) \), is transformed into a depth time series (Saunders and Fofonoff, 1976), and then into a time series of height above the ocean bottom \( H(t) \) relative to the known bottom depth. Thus, it follows:

\[
\cos \theta(t) = \frac{H(t)}{H_{\text{nom}}},
\]

where \( H_{\text{nom}} \) represents the nominal height above bottom of the 1200 m pressure sensor (i.e. without inclination). The depth of each temperature sensor \( i \) at a nominal height above the sea floor \( h_{\text{nom}}^i \) is then given by

\[
h^i(t) = h_{\text{nom}}^i \cos \theta(t).
\]

To obtain greater accuracy in the corrections it is necessary to deploy additional pressure sensors on the mooring line or use a static mooring performance program to better estimate the sensor depths. Appendix A assesses the effect of residual uncertainties in sensor depths on dynamic height and geostrophic velocity calculations, with the result that the residual errors from the first-order correction are largely negligible compared to other errors in the method.

The second step in the method is to create continuous temperature and salinity profiles at each time step from which dynamic height profiles can be derived. The moored temperature sensors provide us with temperature time series at discrete depth levels, in this case at nominal depths of 1200, 1600, 2000, 2500, 3000, 3500, 4000, 4500, 5000 m (at mooring E) and at the bottom. A useful dynamic height profile, however, requires a much higher vertical resolution which we achieve by integrating the temperature gradient \( \partial T/\partial p(T) \) between adjacent instruments. This method follows the approach used previously by Fillenbaum et al. (1997) and Johns et al. (2001), described below.

First, climatological CTD data from the region are used to create an empirical function \( \partial T/\partial p(T) \), which specifies the mean temperature gradient as a function of temperature. This function is combined with the actual temperatures by integrating upward and downward from adjacent measurement points on the mooring and forming a weighted average of these estimates, namely,

\[
T(p) = \sum_{i=1}^{2} w_i \left[ T(p_i) + \int_{p_i}^p \frac{\partial T}{\partial p}(T) \, dp \right],
\]

\[
w_i = 1 - \frac{|p - p_i|}{p_2 - p_1},
\]

where \( i = 1, 2 \) are adjacent measurement levels, and where the weights \( w_i \) are inversely proportional to the vertical distance from the respective measurement depths. The actual depths (pressures) of the instruments derived from the mooring motion correction scheme are used in these calculations. All integrations are done in pressure coordinates, as indicated in the respective equations.

This procedure forces the temperature profile through the measured \( T(p) \) points on the mooring while otherwise being consistent with the local mean stratification. Upward integration is terminated at 1200 dbar due to the instrument losses at mooring B above that level. The \( \partial T/\partial p(T) \) climatology (Fig. 3) is derived from an ensemble
of CTD stations collected on 27 cruises between 1984 and 1997 along the section between mooring locations B and D, and represents the spatial average of $\partial T/\partial p(T)$ over this region (no significant spatial variability is evident). The $\partial T/\partial p(T)$ climatology is seasonal (varying monthly) above temperatures of 18°C but constant at the annual mean below 18°C (see Fillenbaum et al., 1997, for further details). Only the latter is relevant to this application where the analysis is confined to depths below 1200 m ($T < 6^\circ C$).

Once the continuous temperature profiles are obtained, corresponding salinity values (if not measured, as in our application) need to be assigned. A climatological $T/S$ relationship ($S = S(T)$) derived from the same set of CTD casts is applied (Fig. 4) to generate the salinity profiles. For the deep waters below 1200 m (~6°C) the $T/S$ relation is monotonic and highly stable. (Note: in general, it may be necessary to use a theta–$S$ relation in deep waters to account for compressibility effects and avoid the possibility of the $T–S$ relation becoming double valued. However, for the depths and stratification considered here the in situ $T–S$ relation is unique and can be used without loss of information.) Both of the above steps have uncertainties that affect the calculation of baroclinic transports. Appendix B provides an estimate of the related transport errors.

Finally, using the continuous temperature and salinity profiles, profiles of specific volume anomaly $\delta(p, T, S)$ and thus of geopotential anomaly $\Delta \Phi = \int \delta dp$ are generated to obtain the horizontally averaged geostrophic velocity between two moorings relative to a specified pressure level $p_{ref}$:

$$v_{ref}(p, t) = \frac{1}{f(x_2 - x_1)} \int_{p_{ref}}^{p} \left[ \delta(p, t)^{x_2} - \delta(p, t)^{x_1} \right] dp,$$

where $x_1$ and $x_2$ are the mooring locations and $t$ and $f$ denote time and Coriolis parameter, respectively. In our application, $p_{ref}$ is taken at the greatest common pressure (hereafter GCP) of the mooring pair under consideration. For the mooring pair B–D, which is the main focus of our analysis, the GCP is 4800 dbar.

The time varying, spatially averaged geostrophic flow between moorings B and D, relative to 4800 dbar, calculated using the above methodology, is shown in Fig. 5. The time series reveals alternating periods of northward velocity (red) and southward velocity (blue) relative to the bottom. The vertical shear signals are relatively small over the deep water column below 1200 dbar, of order $\pm 10$ cm/s, with very weak shears near the bottom. Further discussion of these signals in comparison with the current meter data will follow in a later section.

### 3.2. Reference velocity fluctuations from bottom pressure sensors

The final step in the estimation of absolute geostrophic currents is determination of the time varying velocity at the reference level. Without geodetic leveling of the bottom pressure gauges on each mooring, it is not possible to determine the absolute velocity at the reference level. However, after removal of mean value and long-term drift, the bottom pressure records $p_{bot}(t)$ can be differenced to yield the time-varying part of the reference velocity:

$$\vec{v}_{ref}(t) = \frac{1}{\rho f(x_2 - x_1)} [p_{bot}^{-x_2}(t) - p_{bot}^{-x_1}(t)].$$

In Fig. 6a, the raw bottom pressure records from moorings B and D are shown, along with the drift.
curves that were removed from them according to the exponential-linear formula of Watts and Kontoyiannis (1990):

$$p_{\text{drift}} = A[1 - e^{Bt}] + Ct + D,$$

where $A$, $B$, $C$ and $D$ are the parameters determined from a least-squares fit. Fig. 6b shows the pressure fluctuations at the two sites after drift removal.

The absolute velocity at the reference level $v_{\text{ref}}(t)$ is then given by

$$v_{\text{ref}}(t) = \tilde{v}_{\text{ref}}(t) + \bar{v},$$

where the remaining offset $\bar{v}$ is constant in time. This offset can in principle be determined by one-time direct velocity measurements at the reference level at any time $t_i$ during the deployment. These measurements must resolve the horizontal velocity field well enough to yield a reliable mean $v_{\text{ref}}(t_i)$. If such measurements are available, we then have:

$$\bar{v} = v_{\text{ref}}(t_i) - \tilde{v}_{\text{ref}}(t_i).$$

In our application, the determination of $v_{\text{ref}}(t_i)$ was provided by lowered-ADCP (LADCP; e.g., Fischer and Visbeck, 1993) sections between moorings B and D. There were two such sections available during the ACCP-3 deployment to determine $\bar{v}$, one in March 1996 and one in July 1996. Each section included 5–6 stations over the 80 km separation between moorings B and D that were designed to resolve the horizontal structure of the flow between the moorings. Theoretically, the constant offset $\bar{v}$ derived from both sections should be the same.

There are three possible ways in which the LADCP data can be used to estimate the spatially averaged reference velocity $\bar{v}$. The first, and simplest, method is to simply integrate the direct LADCP velocity measurements across the section at the reference level. The disadvantages of this method are that it uses only a fraction of the LADCP data (only at the reference level), and that the LADCP velocities very close to the bottom are not always reliable (Fischer and Visbeck, 1993).

The other two methods involve use of the full water column LADCP profiles in combination with either the baroclinic velocities calculated from the moorings at the time of the section occupation, $v_{\text{moor}}(p, t_i)$, or the corresponding baroclinic velocities derived from CTD stations at mooring sites B and D, $v_{\text{CTD}}(p, t_i)$, which were collected simultaneously with the LADCP sections. In these methods, one simply adds to the baroclinic velocity at time $t_i$ the time-varying part of the reference velocity derived from the pressure gauges and calculates the vertical mean. The latter is then subtracted from the vertical and horizontal mean of the LADCP velocities between the moorings to yield $\bar{v}$, i.e.:

$$\bar{v} = \frac{1}{\Delta p} \left[ \int_{p_{\text{ref}}}^{p_{\text{top}}} v(p, t_i) \, dp - \int_{p_{\text{ref}}}^{p_{\text{top}}} [\tilde{v}_{\text{ref}}(t_i) + v_{\text{rel}}(p, t_i)] \, dp \right]$$

with $p_{\text{top}} = 1200$ dbar and $\Delta p = p_{\text{ref}} - p_{\text{top}}$. 

Fig. 5. Time series of baroclinic velocity (m/s) between B and D relative to 4800 dbar. The dashed line, around day 460, indicates the time when the top flotation at site B was lost, leading to increased uncertainty in the velocity estimates thereafter.
Owing to the sometimes large extent of velocity finestructure in the LADCP profiles and the fact that these profiles are not always reliable near the bottom, the first approach of obtaining $\bar{v}$ (that is, only using the LADCP velocities at the bottom reference level) is expected to result in larger errors. A vertical integration should reduce these errors, and thus the second and third approach should be more statistically reliable. However, the inclusion of baroclinic velocity estimates in the latter methods will introduce additional errors.

The results obtained for $\bar{v}$ from these three approaches (LADCP, LADCP + mooring dynamic height, LADCP + CTD dynamic height) are shown in Table 1. For the first approach, we used the LADCP velocities at 100 m off the bottom (which varies from 40 to 100 m above the reference level along the section) to avoid using

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Fig. 6. (a) 40 h low-pass filtered bottom pressure fluctuations at mooring sites B and D, with arbitrary offsets. The superimposed exponential-linear drift curves (dashed lines) were estimated using a least-squares fit. (b) Same pressure time series with drift subtracted. (c) Time series of absolute velocity at the ocean bottom (dark) from above bottom pressures at B and D referenced with LADCP measurements (circles). As a check of consistency, the average current velocities derived from the current meters at 4000 m on moorings B-D are shown in the lighter curve.
LADCP data very near the bottom. Judging from the consistency of the estimates of \( \bar{v} \), the third method, combining the CTD-derived baroclinic velocities with the full-depth LADCP data, appears to yield the most reliable estimates. The two estimates of \( \bar{v} \) from this method are nearly identical. The other two methods show larger differences in the \( \bar{v} \) estimates from the two sections, of 1.8 cm/s for the first method and 1 cm/s for the second method. Overall the procedure of combining full depth LADCP data with the full depth baroclinic velocity between the moorings (determined either from the moorings themselves or from colocated CTD stations) appears to be superior to simple integration of the LADCP velocities at the reference level.

For the final choice of \( \bar{v} \) we used the mean value \((-7.5\, \text{cm/s})\) of the two independent estimates from the third method, \( \bar{v}^{\text{CTD}} \). Adding this offset to the velocity fluctuation record (see Eq. (7)) yields the absolute velocity displayed in Fig. 6c. Also shown in Fig. 6c is an estimate of the averaged velocity between moorings B and D at the 4000 m level derived from the direct current meter measurements on moorings B–D. This time series represents a simple trapezoidal integration of the 4000 m currents at the three sites. The 4000 m level is the deepest level of current measurements available on these moorings and does not correspond to the same depth (~4800 m) as the bottom pressure derived velocity time series. Nevertheless, the two time series show a high correlation on time scales longer than 10 days. On shorter time scales, more energetic fluctuations are found in the current meter time series than in the bottom pressure endpoint technique. This may indicate that for fluctuations on time scales of less than 10 days the horizontal mooring separation of 40 km is too wide to sufficiently resolve the spatial scales of the flow (see Appendix B). A spectral comparison of the two time series is displayed in Fig. 7. Again, good agreement is found for time scales longer than 10 days, but there is an increased energy level for the current meter time series at periods shorter than 10 days.

### 4. Results

In this section the horizontally averaged velocity and transport estimates from the dynamic height method are compared to transport results derived from direct integration of the current meter mooring data.

The absolute velocity profiles obtained from the B–D dynamic height mooring pair, including the bottom reference velocity, are shown in Fig. 8a. The corresponding horizontally averaged velocity profile between moorings B and D derived from the direct current meter observations is shown in Fig. 8b. Vertical integration of these profiles, from each method, provides an independent estimate of the total transport between moorings B and D below 1200 dbar.

The current meter velocity profiles are obtained from a two step process involving: (i) cubic spline interpolation in the vertical at each mooring site

<table>
<thead>
<tr>
<th>Cruise 3/1996</th>
<th>( \bar{v} )</th>
<th>( \bar{v}^{\text{moor}} )</th>
<th>( \bar{v}^{\text{CTD}} )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cruise 7/1996</td>
<td>(-0.051)</td>
<td>(-0.090)</td>
<td>(-0.076)</td>
</tr>
<tr>
<td>Average</td>
<td>(-0.060)</td>
<td>(-0.085)</td>
<td>(-0.075)</td>
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using a shape preserving cubic spine (Akima, 1970), with extension of the vertical shear between the deepest two levels (3000–4000 m) to the sea floor, followed by (ii) linear interpolation in the horizontal of the derived current profiles at moorings B–D and integration over the interval from B to D. This is the standard procedure applied to ACM-1 arrays for estimating direct transports from moored current meter data (e.g., Lee et al., 1996). For the dynamic height method, the transport in the small area between the sea floor and the reference level is computed by applying the reference level velocity to this area, i.e., assuming no shear below the reference level.

The profiles of horizontally averaged velocity between moorings B and D (Fig. 8) computed from these two independent methods show a generally good agreement. Over most of the record both time series show a dominant southward flow with maximum speeds of $O(20 \text{cm/s})$, with occasional reversals to northward flows of about 10 cm/s. A noticeable difference is the relative “smoothness” of the dynamic height (hereafter DH) velocities compared to the current meter (hereafter CM) velocities, which exhibit considerably more high-frequency variability. As mentioned above, this is probably due to the discrete horizontal sampling of the current meter moorings and related spatial aliasing effects. Ageostrophic flows measured by the current meters could also contribute to these differences; however, these effects should be small since all of the data were low-pass filtered with a 40-h filter prior to the computations to eliminate inertial/internal wave contamination.

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Fig. 8. (a) Time series of absolute geostrophic velocity profile (m/s) averaged between moorings B and D, obtained by adding the baroclinic velocities derived from dynamic height profiles at moorings B and D to the reference velocities derived from the bottom pressure gauges. (b) Average velocity profiles between moorings B and D derived from the current meters on moorings B–D. (Dashed line see Fig. 5).
All of the major flow events are reflected in both records, except near the beginning and end of the record where the DH series tends to show a weaker northward flow and/or greater southward flow through the water column (e.g., during days 0–40, and after day 420). As explained in more detail later, we believe these differences are caused mainly by a southward bias in the reference velocity during these periods, resulting from uncertainties in drift removal of the bottom pressure gauges (see Appendix A). These errors have the potential to be largest near the beginning and end of the record. Another source of error in the latter part of the record is the large vertical excursions that occurred at mooring B after the loss of its surface flotation around day 460. This necessitated upward extrapolation of both the temperature and velocity profiles at mooring B over considerable vertical distances (of up to 900 m at times), which adds greatly to the measurement error of both techniques. For this reason our statistical comparison of the methods is limited to the first 450 days of the record. Another interesting difference between the two time series is the relative persistence of a southward velocity maximum near 2000 m in the CM-derived velocities, which is less evident in the DH-derived velocities. The two time series are compared further in Fig. 9 where they have been lowpass filtered with a 15-day cutoff filter to better illustrate the low-frequency variability. Again, except for the above noted differences, the visual agreement between the two sets of profiles is good.

The mean velocity profiles from both methods over the first 450 days of the deployment are compared in Fig. 10. The mean flow is southward at all depths below 1200 m with a maximum near 2000 m in both cases. However, the southward flow maximum derived from the CMs, representing the core depth of the DWBC at this location (Lee et al., 1996), is more intense, and located slightly above the southward flow core determined

![Fig. 9. As in Fig. 8, but for 15 day low-pass filtered data.](image)
by the DH method. Below the DWBC core, the CM-derived profile exhibits greater shear than the DH-derived profile, such that difference between the two methods reverses and below 2500 m the DH profile is the stronger of the two. Over the entire water column between 1200 m and the bottom, the mean velocities of the CM- and DH-derived profiles are 9.18 and 9.04 cm/s, respectively, or less than 2% different. It is tempting to conclude that the DH method somehow underestimates the northward shear in the layers both immediately above and below the DWBC core layer; however, it must be borne in mind that both techniques have errors (see Appendix B), and due to resolution issues the CM profile cannot necessarily be considered as being the more accurate of the two. Further, the overall vertical shear signals are extremely weak, varying by only 3–5 cm/s peak to peak between the bottom and the 2000 m DWBC core layer.

Total transports from each method from 1200 dbar to the bottom, for the 80 km cross-section between sites B and D, are compared in Fig. 11. The transport contributions to the DH method from the baroclinic shear relative to 4800 dbar (hereafter referred to as the “barotropic” transport) are shown separately in Fig. 11a. It is clear from this figure that the dominant contributions to the transport variability are from the barotropic component.

The agreement between the total transport estimates from both methods (Fig. 11b) is good over most of the record except for a clear period of disagreement at the beginning of the record and again near the end of the record, which, as mentioned above, we believe is related to errors in the de-trending of the bottom pressure records. As explained in Appendix A, there is always some uncertainty in the bottom pressure trend that is removed from pressure gauges due to contamination of the least-squares fit by (real) geophysical signals, even if the functional form chosen for the drift curve exactly follows the real drift response of the sensors. This error directly impacts the value of the time varying reference velocity associated with the “barotropic” component of transport, and is by far the most significant source of error in our estimates of the total transport based on the dynamic height method. It contributes an average uncertainty of almost 5 Sv to the total transport estimate between moorings B and D. The distribution of this error is expected to be nonuniform and largest near the ends of the record (see Appendix A). By contrast, the errors in the baroclinic component of the total transport are much smaller, of order 1.5 Sv.

A statistical summary of the two transport methods is shown in Table 2. The mean transport from the CM method is −25.0 Sv, versus a mean transport of −25.5 Sv from the DH-method. Both series show a similar range of total transport variability, which is very large (≈60 Sv peak to peak for the CM transports and ≈75 Sv for the DH transports). The spectra of these transport time series show the same behavior noted above for the reference velocities, that is, a greater high-frequency variability in the CM-derived transports than in the DH-derived geostrophic transports (Fig. 12). The amplitude of variability is similar for periodicities longer than 10 days but for time scales less than 10 days the variance of the CM-derived transports is larger by a factor of 2–5 than that of the geostrophic transports. Again we feel
that the most likely explanation for this difference is overestimation of the transport variability by the CMs at higher frequencies due to insufficient spatial resolution (see Appendix B), whereas the geostrophic estimates are more naturally integrative of the spatial variability between moorings.

The same methodology can be applied separately to the mooring pairs BC and CD (Fig. 13), which add up to the total transport between B and D. It can be seen in Fig. 13 that the largest differences between the CM and DH methods occurs for the BC pair, especially near the beginning and end of the record. This suggests that the main source of the transport discrepancy between moorings B and D for the two methods lies in the drift removal from the bottom pressure record at mooring B. If the actual drift curve at B had been flatter at the beginning and ends of the

Table 2

<table>
<thead>
<tr>
<th>Moor. pair</th>
<th>CM Mean</th>
<th>Std</th>
<th>Min</th>
<th>Max</th>
<th>DH Mean</th>
<th>Std</th>
<th>Min</th>
<th>Max</th>
<th>CM–DH Mean</th>
<th>r.m.s</th>
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<tr>
<td>BD</td>
<td>-25.0</td>
<td>11.7</td>
<td>-46.0</td>
<td>+14.1</td>
<td>-25.5</td>
<td>14.0</td>
<td>-54.0</td>
<td>+22.6</td>
<td>+0.5</td>
<td>8.9</td>
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<tr>
<td>BC</td>
<td>-15.0</td>
<td>9.2</td>
<td>-35.3</td>
<td>+15.5</td>
<td>-15.9</td>
<td>9.8</td>
<td>-40.0</td>
<td>+8.9</td>
<td>+0.9</td>
<td>8.9</td>
</tr>
<tr>
<td>CD</td>
<td>-10.1</td>
<td>7.9</td>
<td>-31.6</td>
<td>+19.6</td>
<td>-9.7</td>
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<td>-34.7</td>
<td>+29.3</td>
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</table>
Fig. 12. Spectra from geostrophic and current meter derived transport (below 1200 dbar averaged between B and D).

Fig. 13. Comparison between geostrophic and current meter derived transports below 1200 dbar averaged between B and C (top) and C and D (bottom).
record than the curve shown in Fig. 6a, then the bottom pressure difference between D and B would be smaller, and the resulting (southward) reference velocities and barotropic transports for the DH method would also be smaller during these periods.

The r.m.s. difference between the CM- and DH-derived transport for the area between B and D is 8.9 Sv over days 30–450 days of the record (Table 2). For the BC and CD pairs, respectively, the r.m.s. differences are 8.9 and 6.1 Sv. These r.m.s. differences can be compared to the expected differences between the transport values arising from errors in each of the transport techniques. Appendices A and B contain analyses of the expected errors, and our methods for estimating them, for the DH and CM methods, respectively. For the DH-method, the cumulative transport error between 1200 dbar and the bottom is expected to be about 5.5 Sv (Fig. 14). (Note the errors are independent of the mooring separation and therefore should be identical for each of the mooring pairs.) For the CM-method the errors for each mooring pair are different since they depend on the current variance at the individual mooring sites and the number of moorings involved in the spatial integration (see Appendix B). For the BD mooring separation of 80 km, the cumulative error from 1200 dbar to the bottom from the CM method is comparable to the uncertainty produced by the DH method (about 5 Sv, Fig. 14a), whereas for the BC and BD mooring separations of 40 km the errors are smaller (~4 and 3 Sv, respectively, Figs. 14b and c). The smaller uncertainty for the CD pair arises due to the lower current variance at sites C and D relative to that at site B. It must be borne in mind that these errors represent the “instantaneous” errors of each method and do not include any time averaging of the data that would tend to reduce the statistical uncertainty of each method. The typical integral (i.e., decorrelation) time scales are approximately 10 days in this region; therefore, the random statistical error of the mean transport from, say, a 1-year record, would be reduced by a factor of about 6 relative to the instantaneous error, or to \( O(1 \text{ Sv}) \).

If we now assume that the errors in the two techniques are independent, which should be a valid assumption, then we can form an estimate of the expected r.m.s. difference between the two methods, which is simply the square root of the sum of the error variances, black solid lines) and the actual observed r.m.s. differences (gray solid lines) are also shown. The top, middle and lower panels are for the BD, BC and CD mooring pairs, respectively.

Fig. 14. Cumulative error (integrated upward from the bottom) for 40-h lowpassed geostrophic (dashed) and current meter (dotted) transports as derived from the error analyses presented in appendices A and B. The estimated r.m.s. differences between the methods (i.e. square root of the sum of their error variances, black solid lines) and the actual observed r.m.s. differences (gray solid lines) are also shown. The top, middle and lower panels are for the BD, BC and CD mooring pairs, respectively.
(light solid lines in Fig. 15) it can be seen that the estimated error is in good agreement with the actual error for the CD pair, but smaller by about 20–30% than the actual r.m.s. differences observed for the BC and BD pairs. This suggests that we have underestimated the overall error of one or both methods for the pairs involving mooring B. Insofar as the DH method is concerned, the main reason for this is probably the increased error associated with drift removal from the bottom pressure gauge at B, as noted above. For the CM method, it is also possible that we are underestimating the transport errors at site B relative to sites farther offshore because of reduced spatial decorrelation scales of the flow close to the Bahamas escarpment. As described in Appendix B, to estimate the CM method errors we used a single lateral correlation function derived from all of the available current meter data from the ACM-1 arrays, irrespective of distance offshore, in order to obtain a robust correlation function. However, there are indications in the data of reduced spatial scales near the western boundary that would tend to result in larger errors for the CM method close to the boundary than farther offshore.

Despite the above differences, we find a generally good agreement in the expected versus observed r.m.s. errors of the two techniques, suggesting that the error analysis presented in the appendices is valid. The overall conclusion from this comparison is that the DH method has an accuracy comparable to that obtainable from conventional current meter arrays, as typified by the ACM-1 moored array configuration. In a parallel investigation, Meinen et al. (2004) used moored inverted echo sounders (IESs) deployed near mooring sites A, B, and D, together with our moored bottom pressure sensors, to estimate total transports using a “GEMS” technique (Meinen and Watts, 2000). In general their results agree very well with the CM-derived transports and our DH method transports, although they find that the IES-GEMS technique is not fully capable of resolving deep shear signals associated with the DWBC.

Fig. 15. (a) Baroclinic (thin) transport (rel. to 4800 dbar) and barotropic (bold) transport fluctuations below 1200 dbar between sites D and E. (b) Geostrophic total transport fluctuations below 1200 dbar between sites D and E (black) and between B and D (gray).
5. Discussion and conclusions

The main strength of the dynamic height method over traditional current meter arrays is its ability to spatially integrate over large distances without having to resolve the flow in the intervening region. For many applications, particularly those involving climate monitoring, it may be sufficient to know only the volume transport and its vertical distribution rather than its detailed horizontal structure. In cases where such an estimate is required over a horizontal scale that is much greater than the typical decorrelation scale of the flow, the dynamic height method becomes an attractive alternative to the use of “coherent” current meter arrays. In practice, this means transport estimates over a distance of order 100 km or greater. Further, the cumulative error in current meter derived transports continues to grow with increasing length of the mooring section, at a rate that is dependent on mooring separation, while the errors in the geostrophically derived transport are theoretically fixed.

As an example of application of this technique over a much wider mooring separation, we show in Fig. 15a the transports derived from the DH method for the mooring pair DE, which spans a separation of 360 km. In this case only the fluctuating part of the “barotropic” transport is shown, due to a lack of sufficient LADCP measurements between sites D and E to provide an absolute reference for the bottom geostrophic velocity. Therefore the barotropic transport for the DE section is uncorrected and is shown relative to a zero mean. The variability of the total transport is again dominated by the barotropic transport as found for the BD region.

The transports in the offshore (D–E) region are shown in comparison with the transports in the western boundary (B–D) region in Fig. 15b, where the mean transport has been removed from each series to compare only the variability. It is evident from this comparison that the transport fluctuations in these two regions are typically anti-correlated and tend to compensate for the major transport events occurring in the other region. This may be associated with, for example, meandering of the DWBC to a location offshore of mooring D or strong barotropic eddies passing from one region into the other. Alternatively, it could indicate the spin up and spin down of a semi-permanent deep recirculation gyre located just offshore of the DWBC (Leaman and Vertes, 1996; Johns et al., 1997; Lee et al., 1990) that would periodically cause an intensification of the DWBC and an associated northward return flow in the offshore region between D–E. While our purpose in this paper is not to provide a detailed analysis of the physical processes underlying these transport fluctuations, we note that at least two of the major events (around day 50 and day 520) were clearly related to meandering of the DWBC to a position offshore of mooring D, while at other times the other above processes or a combination of them are indicated by the data. The main point of this comparison is to show that the ability to spatially average over large distances of energetic and complicated eddy variability (for example, across the entire width of a western boundary current system) can help to reduce unwanted noise in transport estimates.

The most serious difficulty in implementation of the dynamic height method is the long term drift of the bottom pressure sensors and the resulting errors that can occur in the estimates of spatially integrated near-bottom velocity between moorings. Removal of an empirical drift curve from the pressure gauges can be compromised by the fact that low-frequency geophysical variability can be interpreted as sensor drift and corrupt the drift removal process. This is true even if the empirical form of the drift response of the sensors is well known, which is not entirely true for these sensors (Watts and Kontoyiannis, 1990; Eble and Gonzalez, 1991). As stated previously, the sensitivity of the drift removal error to low-frequency variability is greatest at the beginning and end of the record. The potential for error is most severe when the effective time scale of the drift curve (the e-folding scale) is similar to the dominant time scale of the background mesoscale variability. This occurs especially at the beginning of the record, where the drift curve is steepest, with an e-folding scale of order 1 month, superimposed on a dominant variability with time scales of 1–3 months. For this reason it is probably unrealistic to expect that
a sufficiently accurate drift can be removed from the first month or so of such records to yield acceptable estimates of barotropic transport. The same problem can occur near the end of the records although it is less severe.

The best way to increase the reliability of the drift removal process would be to acquire calibration sections with LADCP or some other direct velocity observations near the beginning and end of the records. Especially near the end of the records, such data would be very powerful in constraining the drift curve. This was in fact our intent in the present experiment, but an equipment failure prevented us from acquiring an LADCP section across the array at the end of the deployment. Ideally, four such calibration sections would be acquired during a given deployment, one at (or near) the beginning and ends of the records, and two more distributed in the central portion of the record. To the extent that differences are evident in the estimated reference velocity constants near the ends of the record relative to the ones in the middle, a fitting procedure could be used to remove these drift biases. In principle this could be applied iteratively in the original drift removal process. At a minimum, three calibration sections should be acquired during a deployment, one near the beginning, one near the end, and one in the middle of the record. Additionally, as can be concluded from the exponential shape of the drift curves, the drift rate decreases with time. Thus, the deployment duration should be as long as possible.

At the time of this writing, a new generation of bottom pressure recorders (combined with an inverted echosounder and acoustic data telemetry option) have become available (R. Watts, pers. comm.). Such devices could be left on the sea floor continuously for several years while the data could be transmitted to a vessel at desired intervals. Also, new developments in lowered-ADCP data acquisition and processing may provide a more accurate means for referencing bottom pressure records, through the use of “post-processed” bottom tracking to improve the accuracy of near-bottom velocities (M. Visbeck, A. Thurnherr, personal communication).

The estimation of baroclinic velocities from dynamic height moorings appears to work very well, and the expected errors can be easily evaluated using available hydrographic (CTD) profiles in the region of study, according to the procedures described in Appendix A. The largest error source in the baroclinic transports in the present study came from the assignment of salinity values using a climatological $T/S$ relation (approximately $1.8\text{ Sv}$ for the total surface to bottom transport), with the total error combining all error sources being about $2.5\text{ Sv}$. Incorporation of salinity sensors on the moorings would reduce this error. However, it should be noted that the utility of salinity sensors may be limited by their long term accuracy, especially in deep waters where deviations from climatology may be comparable to the presently available instrument accuracies. The use of pressure sensors on all instruments would also reduce the errors by providing continuous measurements of the depths of each sensor, thereby avoiding having to make any assumptions about mooring behavior. In general, vertical interpolation methods were found to have very little impact on the results, for the reason that any biases introduced by the interpolation method are largely eliminated when adjacent profiles are differenced to create a baroclinic shear profile (see Appendix A). That is, even very basic vertical interpolation schemes (e.g., linear interpolation) yield similar results to more sophisticated interpolation schemes, even though they do not reproduce the actual vertical structure of the dynamic height profiles as well.

In the region of this study off Abaco, Bahamas, the transport fluctuations in the depth interval of interest ($1200–4800\text{ m}$) are dominated by barotropic variability, and the contributions by baroclinic shear were typically much smaller. However, this will not be the case generally and in many other settings it may be expected that the baroclinic shears would be greater and the variability would be dominated instead by the baroclinic transport (the transport relative to zero reference at the bottom). The errors in the dynamic height technique might then be even more favorable compared to estimates derived directly from current meters. In certain applications, such as basinwide monitoring of the meridional overturning circulation (Hirschi et al., 2003), the barotropic
flow is irrelevant and only the baroclinic shear field needs to be measured. In this case the difficulties associated with drift removal and leveling of the bottom pressure gauges are not critical issues. However, for future general applications of this technique, it will be important to improve both the technology of available bottom pressure sensors (e.g., reduction of sensor drift and predictability of drift characteristics), and to refine techniques for absolute geodetic leveling of the bottom pressure sensors.

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Appendix A

A.1. Errors in geostrophic transports derived from dynamic height moorings

This appendix treats geostrophic transport errors from pairs of dynamic height moorings, considering different sources of error which affect the baroclinic (derived from dynamic height) and the barotropic (derived from bottom pressure) contributions.

A.2. Errors in baroclinic transport

We begin by evaluating errors in baroclinic transport between moorings derived from dynamic height, or geopotential, profiles generated at the mooring sites. For this purpose, a dataset consisting of 52 high resolution CTD profiles collected on repeat cruises to the region between 1985 and 1997 was assembled, made up of 26 station pairs collected near mooring sites B and D. Each pair of stations was collected within a few days of each other. These CTD station pairs are used to simulate various errors in the baroclinic transport that can arise through the process of generating continuous vertical profiles of dynamic height from the discrete-depth sensors on the moorings. These errors include: (i) measurement errors in the temperatures recorded by the sensors, (ii) measurement errors in the pressure (depth) values recorded by the sensors or in the depths assigned to any sensors without pressure measurements, (iii) errors due to vertical interpolation, and (iv) errors associated with the assumption of a fixed $T/S$ relationship.

First we examine the transport errors induced by the use of an assumed $T/S$ relation, in the case where no salinity measurements are available on the moorings (as is the case in this study). To do this, a set of relative geostrophic transport profiles is first generated from the original 26 CTD station pairs, which are taken as “truth”. These profiles are then compared to a second set of geostrophic profiles generated by discarding the original salinities and inserting salinity profiles derived from the temperature profiles via the mean $T–S$ relationship. The baroclinic transport errors are obtained by subtracting the corresponding transport profiles from each other. The r.m.s. error in the baroclinic transport incurred by this procedure ($E_s$) is shown in Fig. 16 and reaches a value of about 1.8 Sv at the surface relative to 4600 dbar. This is the largest single source of error in the moored baroclinic transport estimates.

A second error source is the interpolation procedure used to generate a continuous vertical temperature profile from the moored temperature sensors, such as the $\partial T/\partial p$ technique from Eq. (3). The error simulation is carried out in a similar fashion as above. The temperature profiles from the same set of CTD station pairs (near B and D) are first subsampled at the nominal sensor pressures $p_i$. Then the $\partial T/\partial p$ climatology is applied to yield continuous temperatures ($T_b^{B}, T_b^{D}$) and the
corresponding salinities \( S_{\text{sim}}^B, S_{\text{sim}}^D \) are estimated from the local \( T/S \) relationship. The resulting error in the geostrophic transport therefore includes both the errors due to vertical interpolation and the assignment of salinities from the mean \( T/S \) relation. Owing to the fact that the salinity profile is dependent on the interpolated temperature profile, it is not possible to purely isolate the vertical interpolation error in such a calculation. However, the additional error resulting from vertical interpolation can be determined by comparing the results of this calculation to the errors arising from the \( T/S \) assignment alone. The combined interpolation and \( T-S \) lookup error \( (E_s) \) is shown in Fig. 16 and reaches a value of about 2.3 Sv at the surface relative to 4600 dbar. If we assume that the errors due to vertical interpolation and \( T/S \) lookup are uncorrelated and random for the 26 station pairs, then the error due to vertical interpolation can be approximated as \( [E_s^2 + E^2_{\text{si}}]^{1/2} \), which reaches about 1.4 Sv at the surface or about 75% the size of \( E_s \). In actuality, the two errors are not independent because the salinity error depends on the temperature error via the \( T/S \) relationship. Over most of the ocean depth it turns out that this helps to reduce the resulting dynamic height error, because the slope of the \( T/S \) relation is positive and therefore temperature errors of one sign lead to salinity errors of the same sign and partly compensate each other in terms of density. For example, in about 4000 m depth the temperature induced density error is about 25% reduced by the corresponding salinity “error”.

The impact of different choices for the vertical interpolation method is investigated in Fig. 17 which shows the errors in geopotential anomaly

\[
\Delta \Phi_p^0 = \int_{P_0}^P \left[ \delta(S_{\text{sim}}, T_{\text{sim}}, P) - \delta(S, T, P) \right] \, dp
\]

using three different interpolation schemes. Besides the \( \delta T/\delta p \) method, described in the text, also cubic spline and linear interpolation were tested as possible interpolation techniques between the temperature sensors. The total r.m.s. geopotential error is divided in Fig. 17 into a mean bias error (that is, the mean error common to all of the 26

Fig. 16. Gray lines: r.m.s. dynamic height transport errors due to interpolation technique as well as temperature and pressure uncertainty. Black lines: Composite errors.

Fig. 17. Left: r.m.s. \( \Delta \Phi \) error due to use of an interpolation technique between the subsampled temperature data followed by applying a local \( T/S \) relationship to estimate the corresponding salinities. Middle: corresponding mean \( \Delta \Phi \) error. Right: corresponding std \( \Delta \Phi \) error.
station pairs) and a standard deviation (random error) about the mean bias.

The $\partial T/\partial p$ and spline case show nearly equal errors whereas the linear interpolation results in a considerably larger bias error especially above 1500 dbar. All techniques exhibit nearly the same random error throughout the water column. The associated baroclinic transport errors are shown in Fig. 18.

The error profile for each of the techniques is similar, with each of them reaching about 2.3–2.5 Sv at the surface relative to 4600 m. While at first it seems surprising that the linear interpolation technique does not result in significantly larger r.m.s. transport errors than the other two methods, the reason for this is that the large systematic bias in $\Delta \Phi$ only enters the transport error to a small extent because those biases are subtracted from each other in the transport calculation. The transport bias for the linear interpolation is slightly larger than for the other two methods; however, in all three cases the bias is negligible compared to the random errors as is apparent from the r.m.s. values. Nevertheless it is important that the bias error be made as small as possible, since the random errors will average out over time while the bias error will remain. The $\partial T/\partial p$ method has the smallest bias through most of the water column and therefore was chosen as the best interpolation method for the calculations reported in the paper.

The next source of error considered is the uncertainty in determining the actual depths of the sensors. We first consider the impact of deviation of the mooring from an assumption of rigid inclination when exposed to ocean currents. Using a static mooring motion simulation program, the configuration of mooring B was exposed to a comparably strong current profile taken from the current meter measurements at that site. Assuming that pressure measurements were only available at a single depth, nominally at 800 m, the depths of the other sensors were estimated from the inclination angle $\theta$ (using Eqs. (1) and (2)) in the

![Graph](image_url)

**Fig. 18.** Left: r.m.s. transport error due to interpolation between the subsampled temperature data followed by application of a local $T/S$ relationship to estimate the corresponding salinities. Right: corresponding mean transport error.
paper). Those were compared to the actual sensor depths calculated by the static mooring motion program which are considered as “truth”. The results are displayed in Fig. 19 (left), along with the results for two other cases of more moderate flow conditions. In the “extreme” case the actual depths of the sensors can differ by up to 20 m from the assigned depths, and up to about 10 m for the “strong” case. The associated errors in the baroclinic transport are shown by the thin lines in Fig. 19 (right). The error for the “strong” case reaches a maximum of 1.2 Sv near the surface, which is shown in Fig. 16 in comparison to all the other error sources.

A second pressure induced transport error may arise from a pressure offset $\Delta p$ of the pressure sensor itself. For our data an upper bound error of $\Delta p = 10 \text{ dbar}$ was chosen, indicated as the bold solid line in Fig. 19 (left). Such an offset would have the effect that all sensors on the mooring would be displaced vertically by this amount. For the error simulation this offset was applied to mooring B whereas no offset was assumed at site D. The resulting transport error is represented as solid thick line in Fig. 19 (right). It results in an error of 0.5 Sv at 1000 dbar (rel. to 4600 dbar) and increases rapidly toward the surface.

Finally transport errors resulting from temperature sensor measurement errors are considered. The deployed sensors have a measurement uncertainty of $\Delta T = 0.01 ^\circ \text{C}$. To simulate the associated errors, the CTD temperature profiles are subsampled at the nominal sensor pressures. Then to each of the temperature values a random error is added. This error is introduced by a random number generator that yields normally distributed random numbers with a standard deviation of $\Delta T$. The $\partial T/\partial p$ method and the $T/S$ relationship are then applied as usual to obtain high-resolution vertical profiles of $T$ and $S$. These profiles are

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**Fig. 19.** Left: Thin lines (solid and dashed) represent pressure errors under the assumption of rigid mooring inclination in 3 different current scenarios (see top inset in right panel). The bold line represents a constant pressure offset throughout the entire water column, arising from measurement errors of the pressure sensor. Right: Resulting transport bias due to the assumption of rigid inclination (thin lines) and constant pressure offset (bold line).
compared to profiles that were treated the same way except that no random errors were added after subsampling. Finally, transport differences are calculated, which indicate the influence of temperature uncertainties on the transports. The error due to temperature measurement uncertainty reaches about 1 Sv at the surface and therefore must not be completely neglected compared to the other error sources discussed above and shown in Fig. 16.

If we randomly combine all of the above error sources, the standard error in the baroclinic transport reaches about 2.6 Sv at the surface relative to 4600 dbar, or as large as 4 Sv if the errors are combined additively. For the deep transports reported in the paper, where the upper limit of integration is 1200 dbar, the error of the baroclinic transports is expected to be of order 1.5 Sv. In general, the vertical interpolation technique \( \frac{\partial T}{\partial p} \) and associated \( T/S \) lookup uncertainty is the dominant error source in the estimation of moored dynamic height profiles and relative baroclinic transports.

A.3. Errors in barotropic transport or reference velocity

For the barotropic transport component derived from bottom pressure measurements, we consider three main error sources: (i) the temperature sensitivity of the sensors, (ii) the pressure sensor resolution, and (iii) the de-drifting of the bottom pressure records.

According to Paroscientific’s specification we assume the temperature sensitivity to be 0.0008% of the total pressure per degree Kelvin. The temperature standard deviation—taken as a measure of its typical variability—is around 0.03 K at the bottom of all four mooring sites (B–E). With the total pressure being roughly 5000 dbar, the bottom pressure uncertainty due to temperature fluctuations amounts to about 0.001 dbar (1 mm). This corresponds to a transport uncertainty for the 1200–4800 dbar pressure range of 0.8 Sv.

The resolution of the bottom pressure measurements for the sampling configuration used in the experiment was 0.001 dbar (1 mm), which again corresponds to a 0.8 Sv transport error.

The major error contribution turns out to be from detrending the bottom pressure records. Watts and Kontoyiannis (1990) treat this issue by dividing the pressure records into two segments and then detrending each segment independently. Then, at the break points, the differences between the segments (1st segment, 2nd segment and overall) drift curves are regarded as a measure of the error. While this is virtually the only way to produce an error estimate from the bottom pressure data itself, we found that in some cases this method does not generate very reliable estimates as it is quite sensitive to the low frequency variability of the pressure time series. Instead, we develop an estimate of the de-detrifing error that is based on the differences between the reference velocities determined from the two LADCP sections. The two sections were acquired about 4 months apart from each other and the two resulting reference velocity offsets proved to be consistent within 1 cm/s. If we assume that all of the uncertainty is attributable to the pressure sensors (and not to the LADCP and dynamic height measurement involved in the referencing) we come up with a pressure error \( dp = pfv dx/\sqrt{2} = 0.0036 \text{ dbar for the 4 month period.} \) This pressure trend (which equates to about 0.01 dbar over 1 year) results in a r.m.s. pressure uncertainty of 0.006 dbar or a transport uncertainty of 4.7 Sv. It must be borne in mind that, unlike the other errors noted above, this error affects the mean transport and long-term trends over the record length but not the short-term transport variability. The sensitivity of the de-trending procedure to low-frequency signals in the data is greatest near the ends of the records, where only one part of a fluctuation or eddy signal may be recorded, leaving a poor estimate of the temporal mean value for that segment of the time series. Depending on the phasing of the geophysical signal relative to the real drift behavior of the sensor, this can cause the fitted drift curve to be either too flat or steep. The deviation from the true drift is necessarily smallest in the central portion of the record where the drift curve is locally well constrained by the data. Therefore the errors due to de-detrifing of the pressure sensors, which translate directly into errors in the reference
velocity, are generally expected to be largest near the ends of the record.

Adding up the three different contributions above, our best estimate of the total r.m.s. transport error from the dynamic height method is 4.8 Sv. Thus, the barotropic transport uncertainty by far dominates the baroclinic transport uncertainty (1.7 Sv). If the two LADCP sections used for referencing had been acquired at a larger temporal interval from each other (preferably one a short time after the beginning and the other at the end of the mooring deployment period) and they would still have been consistent with each other at the 1 cm/s level, then the bottom pressure dedrifting error estimate could be further reduced. Such a strategy for leveling of the bottom pressure gauges should be employed whenever possible.

Appendix B

B.1. Errors in current meter derived transports

This appendix addresses transport errors resulting from spatial interpolation of measured currents from an array of moored current meters. Since we are using these directly derived transports as a check on the results of the geostrophic method it is important to assess their accuracy. Also it allows for a comparison of the relative accuracies expected for the two techniques.

Transport estimates derived from arrays of moored current meters usually involve some form of interpolation method which may range from simple “box element” summations of the point current measurements to two-dimensional optimal interpolation schemes. For the Abaco observations described here, a two-step approach is used which first consists of vertical interpolation of the point measurements at each mooring into a continuous current profile using a shape-preserving Akima spline (Akima, 1970), and then trapezoidal integration of these profiles over the separation between moorings. This is equivalent to piecewise linear interpolation in the horizontal.

The errors in the current meter derived transport estimates involve errors in both the vertical and horizontal interpolation procedures. The deep currents off Abaco in the 1200–5000 m range are highly coherent between the measurement depths on the moorings (see below) and so the primary error is due to horizontal interpolation between the moorings. The error is dependent on the lateral correlation function and the mooring separation, and it can be estimated using standard objective analysis techniques provided that the correlation function is known or can be suitably approximated (e.g., Bretherton et al., 1976; Fandry and Pillsbury, 1979; Roemmich, 1983).

For the zonal correlation function for the meridional velocity below 1200 m off Abaco we use the general form

$$c(r) = e^{-r/r_c} \cos \left( \frac{\pi r}{2r_w} \right),$$  \hspace{1cm} (B.1)

where $r_c$ and $r_w$ are the Gaussian decay and oscillatory spatial scales, respectively.

Based on the observed lateral correlations between moorings B–D as well as those from earlier moored observations off Abaco at depths greater than 1000 m (Lee et al., 1990, 1996), we find the parameters $r_c = 80$ km and $r_w = 40$ km fit the data quite well (Fig. 20). The pronounced negative lobe in the correlation function is primarily attributable to lateral meandering of

Fig. 20. Observed lateral correlation of meridional velocities off Abaco at depths larger than 1000 m (squares) as a function of mooring separation. The correlation Eq. (B.2) (solid line) with $r_c = 80$ km and $r_w = 40$ km fits the observations well. The resulting percent error in the average velocity between two moorings is given by the dashed line.
the DWBC. The resulting lateral decorrelation scale (defined where the correlation drops to \( e^{-1} \)) is about 30 km in the deep water, which is slightly less than the mooring separation (40 km). This implies that errors in interpolation of the currents between the moorings can be significant, even though a mooring spacing of the same order as the decorrelation scale is arguably an efficient one for transport observations.

The dependence of the correlations on time scale was investigated by computing band-passed correlations for fluctuations with time scales between 2 and 10 days, and for time scales longer than 10 and 30 days, respectively. There is a general decrease in the decorrelation scale with decreasing time scale, with the most significant drop occurring for the 2–10 day time scales. The average correlation in the 2–10 day period band at the shortest spatial separations available (20 km) was only 0.20, compared to correlations of 0.58 and 0.70 for time scales greater than 10 and 30 days, respectively. (The average correlation at 20 km spatial separation is 0.42 including all these time scales; Fig. 20). Therefore, we conclude that the higher “noise level” in CM-derived transports for periods less than 10 days is most probably a result of decreased lateral correlation at those time scales.

Based on the ensemble correlation function (Fig. 20), the error in the estimate of the average current between two moorings, at any given depth, is shown in Fig. 20 as a function of mooring separation. For the 40 km separation between moorings B–D, the error in this average velocity (or the transport per unit depth, when multiplied by the mooring separation) is approximately 25% of the root-mean-square meridional velocity variance at that depth. The mean-square error in the transport between the moorings can then be

![Fig. 21. Left panel: Standard deviation of velocities at mooring sites B–D. Middle panel: \( q(z) \), according to Eq. (B.4), which accounts for the vertical correlation of velocity errors. Right panel: Resulting cumulative transport errors for the BC, CD and BD case.](image-url)
estimated by integrating the error contributions in the vertical (Fandry and Pillsbury, 1979):

\[ \hat{V}^2 = \int_{z_1}^{z_2} \hat{V}^2(z)q(z) \, dz, \quad (B.2) \]

\[ q(z) = \int_{z_1}^{z_2} c(z - z') \, dz'/|z_1 - z_2|, \quad (B.3) \]

where \( \hat{V}^2(z) \) is the mean-square error profile of the transport per unit depth, and the factor \( q(z) \) accounts for the correlation of the errors in the vertical (which is estimated from the vertical correlation between the observation levels—see Fig. 21, middle panel). In the present case the errors are nearly additive in the vertical due to the high vertical correlation of the observed currents. The cumulative transport error (from the bottom upward) is shown in Fig. 21 for the moored array segments B to C, C to D, and for the combined 3-mooring segment from B to D. The errors reach about 5 Sv for the direct transport estimate between B and D, and are larger for the BC segment than for the CD segment owing to the larger meridional current variance at mooring B (see first panel of Fig. 21). These error estimates are relevant for single “snapshots” of the transport that include all subintertial time scales of variability. Temporal averaging over time scales larger than the local temporal decorrelation scale (10 days) can further reduce the transport error.

The above errors are random and do not account for any possible bias that may be present in the mean transport due to undersampling. Such a bias error is, in general, difficult to estimate from the data itself. However, for the Abaco DWBC region, Lee et al. (1996) found from comparing current meter arrays with varying horizontal resolution that the 3-mooring “BCD” configuration used in the present experiment (ACCP-3) tends to underestimate the mean southward DWBC transport, due to underresolution of the DWBC core structure as it meanders in the region between moorings B and D. A statistical correction was applied by Lee et al. (1996) for this sampling bias in their combined analysis of the ACM-1 arrays. We assume that a similar bias may exist in the deep transports estimated from the direct current measurements, however, for the purposes of the methods comparison in this paper no corrections for this possible bias are applied.

References


