Basin-wide integrated volume transports in an eddy-filled ocean

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Abstract

The temporal evolution of the strength of the Atlantic Meridional Overturning Circulation (AMOC) in the subtropical North Atlantic is affected by both remotely-forced, basin-scale meridionally coherent, climate-relevant transport anomalies, such as changes in high-latitude deep water formation rates, and locally forced transport anomalies, such as eddies or Rossby waves, possibly associated with small meridional coherence scales, which can be considered as noise. We focus on the extent to which local eddies and Rossby waves when impinging on the western boundary of the Atlantic affect the temporal variability of the AMOC at 26.5°N. Continuous estimates of the AMOC at this latitude have been made since April 2004 by combining Florida Current, Ekman and mid-ocean transports with the latter obtained from continuous density measurements between the coasts of the Bahamas and Morocco, representing the western and eastern boundaries of the Atlantic at this latitude.

Within 100 km of the western boundary there is a threefold decrease in sea surface height variability towards the boundary, observed in both dynamic heights from in situ density measurements and altimetric heights. As a consequence the basin-wide zonally integrated upper mid-ocean transport shallower than 1000 m - as observed continuously between April 2004 and October 2006 - varies by only 3.0 Sv rms. Instead, upper mid-ocean transports integrated from western boundary stations 16, 40 and 500 km offshore to the eastern boundary vary by 3.6, 6.0 and 10.7 Sv rms respectively.
The reduction in eddy energy towards the western boundary is reproduced in a non-linear reduced gravity model suggesting that boundary-trapped waves may account for the observed decline in variability in the coastal zone as they provide a mechanism of fast equatorward export of transport anomalies associated with eddies impinging on the western boundary. An analytical model of linear Rossby waves suggests a simple scaling for the reduction in thermocline thickness variability towards the boundary. Physically the reduction in amplitude is understood as along-boundary pressure gradients accelerating the fluid and rapidly propagating pressure anomalies along the boundary. The results suggests that the local eddy field does not have the potential to dominate upper mid-ocean transport (or AMOC) variability at 26.5°N on inter-annual to decadal time scales.

1. Introduction

Observational studies targeting large-scale circulation changes have relied on “end-point” measurement techniques to provide estimates of the net transport across vertical sections (e.g. Whitworth et al., 1983, McPhaden et al., 2002; Johns et al., 2005; Kanzow et al., 2006). The large-scale flow in the ocean interior is in geostrophic balance to first order (away from the surface and bottom Ekman layers) and end-point vertical density profiles can be used to compute the net geostrophic flow (relative to a level of reference) between them. The absolute geostrophic horizontal velocity $\mathbf{U} = (u,v)$ can be deduced from the horizontal gradient in pressure $P$ in the ocean, according to

$$2 \Omega \times \mathbf{U} = \frac{1}{\rho} \nabla \mu P$$

[1]
with $\Omega$ and $\rho$ denoting the Earth's angular velocity and reference ocean density respectively. On a zonal section the Coriolis parameter $f$ is constant and the horizontal pressure difference between the eastern and western end-points ($P_E, P_W$) is proportional to the zonally integrated northward geostrophic volume transport per-unit-depth $T$,

$$[2] \quad T(t,z) = \left[ P_E(t,z) - P_W(t,z) \right] \rho f $$

with $z$ and $t$ denoting depth and time respectively.

The pressure at $z=-h$ can be written as the displacement of sea level from its time mean $\eta$ and, the density of the water column,

$$[3] \quad P(t,-h) = \rho_0 g \eta(t) - \int_{-h}^{0} \rho(t,z) g dz $$

where $g$ is Earth's gravitational acceleration. It is assumed that an inverted barometer correction has been applied to $\eta$, at a rate of $\Delta P_A(t) = g \rho_0 \Delta \eta(t)$ (e.g., Fu and Pihos, 1994) so dynamically inactive changes $\Delta \eta$ due to deviations in atmospheric pressure from the long-term mean $\Delta P_A$ are removed. The geostrophic balance is a valid approximation even in the strongest (fastest) ocean current systems such as the Antarctic Circumpolar Current (Whitworth, 1983), the Gulf Stream (Sato and Rossby, 1995), the Kuroshio (Johns et al., 2001), the Agulhas Current (Beal and Bryden, 1999) or the Atlantic deep western boundary current (Johns et al., 2005).

Recently, the endpoint measurement technique has been used to infer meridional transport variations across a 6000 km transatlantic section between the Bahamas and the Moroccan coast at 26.5°N in the Atlantic (Fig.1; Cunningham et al., 2007; Kanzow et al., 2007) – hereafter referred to as mid-ocean transport. Cunningham et al. (2007) show that the daily upper mid-ocean transport shallower than approximately 1000 m displays fluctuations of 3.1 Sv rms for the period between
April 2004 and April 2005. As a consequence of westward energy transfer effected by Rossby waves and eddies, and of instabilities in the boundary current systems significantly elevated levels of surface eddy kinetic energy are commonly found near the western boundaries of the ocean (see Fig. 9b of Wunsch and Stammer, 1998). Using mooring based observations of upper-ocean transports across 26.5°N we investigate how eddy variability within 500 km of the western boundary affects basin-wide integrated transports. This question is important for the interpretation of the variability of both the Atlantic Meridional overturning circulation (AMOC) and subtropical gyre at this latitude. Important currents in the western boundary region here include the southward-flowing deep western boundary current (DWBC) below 1000 m and the northward flowing Antilles Current in the upper 1000 m (Lee et al., 1990; 1996, Bryden et al. 2005, Johns et al., 2008).

Transient eddy-variability is sometimes important for horizontal redistribution of heat, salt and momentum but may be a source of noise in experiments aimed at characterising the temporal evolution of the large-scale ocean circulation using end-point methods (e.g. Kanzow et al., 2008a). An eddy’s rotation-related maximum pressure anomaly is located at its centre so that the rotational transport is largest integrating over its radius. Integrating over the eddy's diameter, the rotational transport cancels. Thus, if the eddy’s horizontal extent is fully contained within the section when passing across it, only the across-section eddy translation but not its rotational flow has an impact on net across-section transports. Conversely eddies passing over one end-point contribute to fluctuations in the net zonal pressure gradient.

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1 Eddy variability: a collective term in this study referring to transient eddies, Rossby waves and recirculations of the western boundary Antilles Current.
– causing noise that has the potential to mask the temporal evolution of pressure
gradients associated with the large-scale flow field.

Altimetry might represent an efficient tool to estimate the time-variable

strength of upper-ocean velocities. Transient features of geostrophic surface flow are
proportional to the sea-surface slope \( \eta \) and have been monitored by satellite altimetry.
Numerous application have been presented for the study of wind-driven gyres,
mesoscale flows and strong ocean currents (e.g. Stammer, 1997; Chelton et al, 2007;
Cunningham and Pavic, 2007). The time-variable geostrophic transport \( T_\eta \) over the
depth range \([-h,0]\] can be estimated by associating the geostrophic surface flow with
one dominant vertical mode of horizontal velocity \( F(z) \) according to,

\[
[4] \quad T_\eta = [\eta_w - \eta_E] \int_{-h}^{0} F(z) dz .
\]

Away from the boundaries the upper-ocean geostrophic flow on time scales
longer than a few weeks exhibits a surface-intensified first mode baroclinic structure
(Wunsch et al., 1997), whose velocity zero-crossing is typically found around 1500 m.
Such motions mainly reflect westward propagating baroclinic eddies and Rossby
waves, as satellite observations reveal (Chelton et al., 2007). However, approximating
the vertical structure of geostrophic flow by linear baroclinic modes can fail near the
sea surface. A sign of failure is that there is no sea surface temperature signal to the
linear mode. Observations from the Mid-Ocean Dynamics Experiment (MODE
Group, 1978) showed that temperature-inferred vertical displacements do not
necessarily decrease towards the sea surface (Richman et al., 1977), with the near-
surface departure from linear mode theory possibly being due to shear modes
(Beckmann, 1988).
Barotropic motions gain in importance in the less stratified deep ocean (Wunsch et al., 1997; Johns et al., 2005; Kanzow et al., 2008a) and from a comparison of in situ observations and altimetry in the tropical North Atlantic Kanzow et al. (2008a) concluded that variations of $\eta$ are not representative of transport changes below 1000 m.

Near boundaries a separation of the barotropic and baroclinic modes is ambiguous because an across-boundary integral of barotropic along-boundary velocities over sloping bathymetry inevitably results in a baroclinic transport profile. Topographically trapped waves may also become important (e.g., Johns and Watts, 1986) such that the relationship between $\eta$ and upper-ocean transports may vary across the boundary. In addition, competing dynamics and their manifestations as vertical modes means a stable relationship between $\eta$ and upper ocean transports might not exist.

Using density and current meter data from the RAPID/MOCHA moorings within 500 km of the western boundary at 26.5°N we will investigate the relationship between transports and $\eta$. In particular, we test whether a robust relationship between $\eta$ at the western boundary (or the boundary-to-boundary difference in $\eta$) and the upper mid-ocean transport across 26.5°N can be established according to [4]. If successful, this would enable us to estimate the strength of the AMOC at 26.5°N continuously since the launch of the TOPEX/POSEIDON altimeter in 1992 (combined with continuous estimates of the Gulf Stream and Ekman transports over this period from the Florida Current telephone cable measurements (Baringer and Larsen, 2001) and wind field reanalysis products (such as NCEP / NCAR)). A stable relationship between the difference in $\eta$ within a few hundred kilometres from western boundary
and \( \eta \) at the eastern boundary may provide a proxy of the strength the wind-driven

subtropical gyre at this latitude.

The paper is structured as follows. In section two we present the underlying
data and methods. Section three describes the mid-ocean zonally integrated transports
between Morocco and four locations up to 500 km east of the Bahamas and these
transports are compared with variability in altimetric observations. In section four we
analyse the effect of eddies impinging on the western boundary on basin-wide
integrated flows by means of numerical model experiments. A theoretical explanation
for the model results is presented in section five. The effect of eddies close to and at
the western boundary on end-point measurements and the implications for measuring
and interpreting basin-wide integrated volume transports such as the AMOC are
discussed in section six, followed by a brief summary (section 7).

2. Data and Methods

a. Data

To study the influence of eddy-variability on basin-wide integrated transports
we derive a set of estimates of the meridional geostrophic volume transport across
26.5°N above 1000 m. We integrate the meridional geostrophic velocity between the
African coast and three mooring sites 500, 40 and 16 km of Abaco, the Bahamas
(moorings sites WB5, WB3, WB2 see Figs. 1 & 2) between April 2004 and October
2006.

The RAPID / MOCHA system began monitoring daily variations in the
strength and vertical structure of the AMOC in April 2004. To monitor mid-ocean
transport the system consists of an array of moorings measuring salinity (S) temperature (T) and pressure (P) near the western boundary (mooring sites WB2, WBH1, WBH2, WB3, WB5) the eastern boundary (EB1, EBHi, EBH0 to EBH5) and on both flanks of the Mid-Atlantic Ridge (MAR1 and MAR2; Figs. 1, 2 & 3). Vertical profiles of density at the western and eastern boundaries are used to compute estimates of meridional geostrophic transport between the eastern boundary and the western boundary moorings (see Table 1). The western boundary end-of-section profile uses data merged from WB2 shallower than 4000 m and from WBH1 / WBH2 (and WB3 after April 2005). Vertical profiles of S, T and P have also been derived separately at WB3 and WB5. At the eastern boundary, S, T and P data from all eastern boundary moorings have been merged into one profile from 4840 m to the shallowest available level during each deployment (Table 1; Kanzow et al., 2007).

Temperature, salinity and pressure are sampled every 15 minutes and then 2 day low-pass filtered and sub-sampled ever 12 hours (Kanzow et al., 2007; www.noc.soton.ac.uk/rapidmoc). Filtered and sub-sampled S, T and P data at each site are vertically interpolated onto a 20 dbar grid (Kanzow et al., 2007) by applying an interpolation technique that makes use of climatological estimates of the vertical shape of the temperature and salinity profile \([dT/dP(T); dS/dP (T)]\) following Johns et al. (2005). Density is then computed from the 12 hourly S, T and P profiles.

Moorings at sites WB0, WB1, WB2 and WB3 are instrumented with current meters (Fig. 3) and at mooring sites WBA and WB0 acoustic Doppler current profilers (ADCP) (Fig. 3) measure the velocity field in the upper 500 m (Johns et al., 2008). Velocity data are 40 hour low-pass filtered and sub-sampled on a 12 hourly grid (Johns et al., 2008). At each time step velocities are interpolated to a spatial grid
with zonal and vertical resolutions of 500 m and 20 m, respectively (Johns et al., 2008). The gridded velocity data are then used to compute profiles of zonally integrated transports per-unit-depth between the Abaco shelf and WB2 and between the Abaco shelf and WB3.

Bottom pressure ($P_{BOT}$) has been measured continuously throughout the 30 months at mooring sites WB3 and WB5. Combined with the density profiles from these two sites we compute geostrophic transports between them. $P_{BOT}$ samples at ten minute intervals are two-day low-pass filtered and subsequently sub-sampled on a 12 hourly basis (Kanzow et al., 2007). A long-term instrumental trend has been removed from each bottom pressure record by subtracting a least-squares exponential-linear fit from the data of each deployment period (Watts and Kontoyiannis (1990)). Finally, the time mean is removed (Fig. 4).

b. methods

Meridional transports across 26.5°N zonally averaged between Florida and Africa and vertically averaged over the top 1000 m are a reliable measure of the strength of the Atlantic Meridional Overturning Circulation (Cunningham et al., 2007) because the mid-depth level of no motion exhibits fluctuations in vertical displacement of only 90 m rms around its time mean of 1041 m for daily transport values. The basin-wide integrated northward transport $T_{BASIN}$ across 26.5°N is computed as the sum of: (i) the zonally averaged Ekman transport $T_{EK}$; (ii) Gulf Stream transport integrated across the Straits of Florida (between Florida and the Bahamas) $T_{GS}$ and; (iii) the mid-ocean transport (the northward geostrophic transport integrated between the Bahamas and the African coast) $T_{MO}$. 
[5] \[ T_{BASIN}(z,t) = T_{GS}(z,t) + T_{EK}(z,t) + T_{MO}(z,t) \]

The notation \( T_{GS}(z,t) \), \( T_{EK}(z,t) \), \( T_{MO}(z,t) \) etc. indicates transports per unit depth \( z \), while \( T_{GS}(t) \), \( T_{EK}(t) \), \( T_{MO}(t) \) etc. indicates vertically integrated transports. As each variable is a function of time \( t \), the explicit mentioning of the time-dependence will be dropped hereafter. Transport fluctuations are anomalies about the time mean, unless otherwise noted.

For the computation of \( T_{EK} \) we used daily wind stress data from the QuikSCAT scatterometer mission (Kanzow et al., 2007) while daily estimates of \( T_{GS} \) are available from the telephone cable based Florida Current monitoring system (Baringer and Larsen, 2001). \( T_{MO}(z,t) \) can be divided into two components: (i) the transport in the 16 km wide “western boundary wedge” between the Abaco shelf (AS) and site WB2 \( T_{AS-WB2}(z) \) is derived from direct current meter measurements on moorings WBA, WB0, WB1 and WB2 (Fig. 2; for details see Johns et al., 2008; Kanzow et al., 2008b) and (ii) the surface-to-4740m geostrophic transport between WB2 and the eastern boundary. The geostrophic transport relative to the reference level \( z_{REF} = -4740 \) m between WB2 and the eastern boundary – referred to as internal transport \( T_{INT} \) - is computed from continuous observations of density profiles at the section end points (\( \rho_{WB2} \) and \( \rho_{EAST} \)) according to

\[
[6] \quad T_{INT}(z) = -\frac{g}{\rho} \int z_{REF}^{z} [\rho_{EAST}(z') - \rho_{WB2}(z')] dz.
\]

To compute absolute geostrophic transports, for each 12 hourly profile of \( T_{INT}(z) \) a reference transport \( T_{C} \) must be derived.

From the first year of the RAPID measurements Kanzow et al. (2007) show that there is a strong negative correlation between \( T_{MO} \) and the sum of \( T_{GS} \) and \( T_{EK} \) at periods in excess of 10 days, such that the residual (zonally and vertically integrated)
transport across 26.5°N displays much smaller fluctuations than either component.

Kanzow et al. (2007) argue that variability of the residual transport is mostly due to measurement uncertainties rather than real fluctuations that could have been caused by changes in the strength of the Arctic throughflow. Here we assume that there is precise compensation among the different flow components, in the sense that the sea surface to sea floor \((z = -h_{BOT})\) integral yields zero residual mass transport across 26.5°N (Kanzow et al., 2007; Cunningham et al., 2007) according to

\[
\int_{-h_{BOT}}^{0} \left[ T_{GS}(z) + T_{EK}(z) + T_{MO}(z) \right] dz = 0.
\]

The net upper mid-ocean transport \(U_{MO}\) is defined as the mid-ocean transport \(T_{MO}(z)\) integrated over the upper 1000 m, according to,

\[
U_{MO} = \int_{-1000}^{0} \left[ T_{INT}(z) + T_{c}(z) + T_{AS-WB2}(z) \right] dz
\]

Formally, there is a small southward mass transport across 26.5°N associated with the sum of Bering Straits flow from the Pacific into the Arctic Ocean and of net precipitation between Bering Straits and 26.5°N. Because each of these components is less than 1 Sv and there is little information on how they vary in time (Woodgate and Aagaard, 2005; Wijffels, 2001) we assume zero residual mass transport at each moment of time and the referencing of \(T_{INT}(z)\) is carried out by computing a compensating transport profile \(T_{C}\) at each time step as follows,

\[
T_{C} = -\int_{-h_{BOT}}^{0} \left[ T_{GS}(z) + T_{EK}(z) + T_{AS-WB2}(z) + T_{INT}(z) \right] dz.
\]

For the computation of the corresponding compensation transport profile \(T_{C}(z)\) we assume that the underlying compensating meridional velocity field \(V_{C}\) is spatially uniform (vertically and zonally) such that,

\[
T_{C}(z) = V_{C} \int_{X_{CR}}^{X_{CR}^{'}} dx = V_{C} L(z)
\]
with $X_E$ and $X_W$ denoting the eastern and western boundary, respectively, and $L$ denoting the effective width of the ocean, that decreases with depth.

Throughout this study we apply a 10-day low-pass filter to $U_{MO}$ (and to all other transport time series) as transport changes at periods shorter than 10 days may not be representative of AMOC fluctuations (Kanzow et al., 2007). $T_c(z)$ defined in [10] is nearly depth-independent shallower than 3500 m, whereas below that the reduction in the effective zonal ocean width (below the crest of the Mid-Atlantic Ridge; Fig. 1) makes the amplitude of $T_c(z)$ decrease with depth. Overall, only about 25% of $T_c(z)$ occurs shallower than 1000 m.

To compute the upper-ocean transport east of WB2 $U_{WB2-EAST}$ we subtract from $U_{MO}$ the transport west of WB2 in the upper 1000 m ($U_{AS-WB2}$).

$$[11] \quad U_{WB2-EAST} = U_{MO} - \int_{-1000}^{0} T_{AS-WB2}(z')dz$$

Similarly, to obtain the upper mid-ocean transport east of WB3 ($U_{WB3-EAST}$) we subtract from $U_{MO}$ the transport west of WB3 ($U_{AS-WB3}$) integrated over the top 1000 m,

$$[12] \quad U_{WB3-EAST} = U_{MO} - \int_{-1000}^{0} T_{AS-WB3}(z')dz.$$ 

$U_{AS-WB3}$ is computed from the direct current meter measurements on WBA, WB1, WB2 and WB3 (Fig. 3; Johns et al., 2008). Finally, to obtain the upper mid-ocean transport east of WB5 ($U_{WB5-EAST}$) we subtract from $U_{MO}$ the sum of $U_{AS-WB3}$ and the transport between WB3 and WB5 above 1000m ($U_{WB3-WB5}$)

$$[13] \quad U_{WB5-EAST} = U_{MO} - \int_{-1000}^{0} [T_{AS-WB3}(z') + T_{WB3-WB5}(z')]dz.$$ 

$U_{WB3-WB5}$ is computed in three steps. Firstly we compute the internal geostrophic transport relative to 4740 m between WB3 and WB5 from the density profiles at both
sites ($\rho_{WB3}$ and $\rho_{WB5}$) as in [6]. Secondly, both mooring sites comprise bottom pressure sensors, from which near-bottom northward external transport (per unit depth) $T_{EXT}(z)$ integrated between WB3 and WB5 can be computed (e.g., Kanzow et al., 2008b) according to

$$ T_{EXT}(z) = \frac{1}{\rho_f} \left[ P_{bot}^{WB5} - P_{bot}^{WB3} \right] $$

Finally, adding internal and external transports gives geostrophic transport fluctuations between WB3 and WB5 $T_{WB3-WB5}(z)$.

We will compare dynamic sea surface height at the mooring locations and the mid-ocean transport anomalies to satellite derived sea surface heights and their zonal differences. We use the gridded multi-satellite merged altimeter data set DT-MSLA “Upd” (Delayed-time maps of sea level anomalies “updated”) described by Dibarboure et al. (2008) and provided by AVISO (http://www.aviso.oceanobs.com). The altimetry data set spans the period between October 1992 and January 2008 and has a nominal resolution of 1/3° in space and 7 days in time. The underlying satellite missions include Topex/Poseidon, Jason-1, ERS-2 and Envisat. We extract the sea surface height $\eta$ along 26.5°N and at each time step we subtract the zonal mean $\eta$ (Bahamas to Morocco) from each grid point. This suppresses the dynamically inactive portion of the seasonal variations of $\eta$ that is mainly associated with seasonal near-sea-surface thermal expansion.

The Jason-1 groundtrack closest to the Abaco shelf approaches to within about 4 km of WB2, crossing 26.5°N east of WB2. It is an ascending track that enters deep waters from the Nassau shelf about 200 km southwest of WB2 (Fig. 1b). The correlation between $\eta$ at the point closest to WB2 from the Jason along-track data set and $\eta$ at WB2 extracted from the gridded data set (with the latter being subject to
spatial and temporal averaging) is R=0.79. A descending Jason-1 track approaches
WB3 to within 15 km (minimum distance) crossing 26.5°N to the east of WB3 (Fig. 1b).

Throughout the paper anomalies of transport and sea surface height will be
discussed such that T and ς represent fluctuations with the time mean removed.

3. Upper ocean transport and sea surface height

In the following we test the extent to which altimetric observations of ς can be
used to gain information on spatio–temporal characteristics of zonally integrated
upper ocean transport. Fig. 5 shows ς for the period between October 1992 and
January 2008 within 700 km of the western boundary along 26.5° N. It reveals
coherent westward propagation of ς anomalies at roughly 6 cm s⁻¹, with periods
between two and six months. However, some of the propagating anomalies entering
the domain at 70° W disintegrate before they reach 76° W. e.g. positive and negative
anomalies appear at 70° W in June 1996 and July 2003 vanishing before crossing 74°
W.

Superimposed on the propagation pattern are numerous isolated stationary
anomalies. that sometimes appear as dipolar structures. Three such events can be seen
between 76° W and 73° W in the period from October 2002 through July 2003.
Interannual variations are also apparent, both propagating and stationary. The
strongest interannual changes are centred around 75.8° W (Fig. 5 B) where ς displays
anomalously negative values from Feb 2001 through June 2002 and from April 2006
Both stationary and propagating anomalies have in common that their rms amplitude and rms variability appears to diminish west of 76° W.

In the western boundary region at 26.5° N there is a gradual westward increase in the rms fluctuations of $\eta$ from 6.5 cm rms at 65°W to its maximum 11.2 cm rms at 75.8°W for the period of altimetric observations between 10/1992 – 01/2008 (solid line in Fig. 6). West of 75.8°W there is a rapid decrease in the variability of $\eta$ from 11.2 cm rms to 5.3 cm rms at 76.75°W. Estimates of the rms amplitude of $\eta$ west of 76.75°W (WB2) are possibly unreliable due to a lack of altimetric observations.

Similar reductions in rms variability are confirmed for the shorter RAPID measurement period between 04/2004 and 10/2006 (red dashed line) and the ACCP-3 measurement period (Johns et al., 2005) between 10/1995 and 06/1997 (blue dashed line).

The rapid decrease of eddy energy toward the boundary may be due to the abrupt, cliff-like topography of the Bahamian escarpment, that deepens from 500 m to 5000 m depth over 25 km. We show however, that a decrease in eddy energy close to the western boundary is a general expectation from theory. This decrease in eddy energy within a few tens of kilometers of the western boundary differs from the general tendency for eddy energy to decrease away from energetic western boundary regions toward the ocean interior on scales of several hundred kilometers (as seen, for example, in the MODE-1 Experiment a few degrees farther north (Mode-1 Atlas Group, 1977)).

Also shown in Fig. 6 are rms fluctuations of dynamic height at 200 m relative to the sea floor from the RAPID mooring measurements (red crosses) and from mooring C of the ACCP-3 experiment (Johns et al., 2005). Using dynamic height at
200 m rather than at sea level suppresses the seasonal cycle of the near-surface thermal expansion. Dynamic heights at WB5, C, WB3, and WB2 vary by 9.1, 11.0, 5.9 and 3.9 dynamic cm rms, respectively, and are in agreement with the altimetric observations (Fig. 6). In particular, the sharp decrease to the west of site C is reproduced. If rms fluctuations of dynamic height from mooring C are representative of those during the RAPID measurement period, the implied decrease in dynamic height variations (or $\eta$) is 7.1 cm reducing from 11 cm at 76.1°W to 3.9 cm rms at 76.75°W (i.e. over 65 km). This may be a steeper gradient than the gridded altimeter data set can resolve with its resolution of about 33 km.

If $\eta$ or dynamic height near the sea surface can be used as a proxy for upper ocean transport, then according to [4] meridional transport variations integrated between WB5 and the eastern boundary should be significantly larger than between WB2 and the eastern boundary. Not in passing that at the eastern boundary the rms fluctuations of $\eta$ amount to only 2.2 cm rms at EBH4 (Fig. 2).

The 10-day low-pass filtered basin-wide integrated upper mid-ocean transport $U_{MO}$ exhibits fluctuations of 3.0 Sv rms over the 30 months of observations (Fig. 7, Table 2). It shows pronounced seasonal variations that suggest a reduced southward flow in autumn and enhanced southward flow in spring. The standard error of $U_{MO}$ (based on the degrees of freedom in table 2) is 0.6 Sv for the 30 month long time series and 0.9 Sv for a 12 month segment.

The upper ocean transport between WB2 and the eastern boundary ($U_{WB2\text{-}EAST}$) displays a variability of 3.6 Sv rms and a correlation of 0.81 with $U_{MO}$ which is significant at 5% error probability (Table 2, Fig. 8). Transports between WB3 and the eastern boundary ($U_{WB3\text{-}EAST}$) and between WB5 and the eastern boundary ($U_{WB5\text{-}EAST}$)
exhibit variations of 6.0 Sv and 10.8 Sv rms, respectively (Table 2, Fig. 8). The correlations of $U_{WB3-EAST}$ and $U_{WB5-EAST}$ with $U_{MO}$ are 0.45 and 0.10, respectively, with the first value being barely significant and the second not exceeding the 5% error probability threshold of 0.42, (Table 2). The differences $U_{WB2-EAST} - U_{MO}$ and $U_{WB3-EAST} - U_{MO}$ are 2.1 and 5.3 Sv rms.

The following picture emerges within roughly 100 km of the western boundary of the Atlantic at 26.5° N. The further away from the Abaco shelf the western end point of the section is moved to the east, the larger the variations of the zonally integrated transports become. $U_{WB2-EAST}$, $U_{WB3-EAST}$ and $U_{WB5-EAST}$ exceed $U_{MO}$ by a factor of 1.2, 2 and 3.6 in terms of rms fluctuations. The increase in transport variations is in qualitative agreement with the offshore increase in the rms amplitude of $\eta$ in the vicinity of the western boundary. This holds both for intra-seasonal and interannual fluctuations (Fig. 6). The correlation analysis suggests that $U_{WB2-EAST}$ is moderately representative of $U_{MO}$ with an rms uncertainty of 2.1 Sv (which is not small compared to the 3.0 Sv rms fluctuations of $U_{MO}$). The 2.1 Sv rms is, according to [8], the rms variability that the upper ocean transports west of WB2 ($U_{AS-WB2}$) contributes. A negative correlation of -0.55 between $U_{WB2-EAST}$ and $U_{AS-WB2}$ is observed - possibly resulting from Antilles Current meanders and eddies passing over WB2 – and explains why $U_{WB2-EAST}$ displays a larger transport fluctuations than does $U_{MO}$. The transport components further offshore (i.e., $U_{WB3-EAST}$, $U_{WB5-EAST}$) cannot serve as a reliable measure of $U_{MO}$. Thus to estimate variations in $U_{MO}$ reliably (as part of the AMOC at 26.5°N) density measurements have to be obtained as close to the boundary as possible (and not several tens of kilometres away from it).
We now test whether zonal differences in $\eta$ provide a straightforward way of estimating zonally integrated upper-ocean transports across 26.5°N. The correlation of the difference in $\eta$ between the eastern boundary and WB5, WB3, WB2 and WB1 ($\eta_{EAST} - \eta_{WBX}$, with X=5, 3, 2 and 1 denoting the different WB mooring sites) with the corresponding transports $U_{WB5\text{-}EAST}$, $U_{WB3\text{-}EAST}$, $U_{WB2\text{-}EAST}$, and $U_{MO}$ is 0.85, 0.71, 0.44 and 0.12 respectively (Fig. 9). The difference $\eta_{EAST} - \eta_{WB1}$ was used to compared to $U_{MO}$ rather than $\eta$ at the Abaco shelf break, because (i) variations of $U_{WB1\text{-}EAST}$ (i.e., upper ocean transport east of WB1) are representative of $U_{MO}$ with their correlation being 0.95 and, (ii) sea surface variations from the gridded altimetry data set to west of WB2 are unreliable due to lack of measurements (section 2). A linear regression of the form $U_{WBX\text{-}EAST} = s_{WBX} \times (\eta_{EAST} - \eta_{WBX})$ gives $s_{WBX} = 0.91$, 0.68 and 0.32 Sv cm$^{-1}$ for $U_{WB5\text{-}EAST}$, $U_{WB3\text{-}EAST}$ and $U_{WB2\text{-}EAST}$ respectively. Subtracting the fit from the respective upper-ocean transports yields transport residuals $U_{WBX\text{-}EAST} - s_{WBX} \times (\eta_{EAST} - \eta_{WBX})$ of 5.7, 4.2 and 3.2 Sv rms for X = 5, 3 and 2.

The linear fit using $\eta_{EAST} - \eta_{WBX}$ explains 72, 51 and 19% of the variance of $U_{WB5\text{-}EAST}$, $U_{WB3\text{-}EAST}$, $U_{WB2\text{-}EAST}$. Although the transport variance explained by $\eta_{EAST} - \eta_{WBX}$ is largest at WB5, the absolute residual between the linear fit and the corresponding transport is largest, too. Both relative prediction skill and the absolute error decrease toward the boundary. At WB2 the error of the transport prediction (3.2 Sv rms) has roughly the same amplitude as $U_{MO}$. That means reliable estimates of $U_{MO}$ cannot be obtained by combining a current meter based boundary current array (measuring the transports between the shelf and, say, WB2) with altimetric observations of $\eta_{EAST} - \eta_{WB2}$. 
The fact that $\eta_{EAST} - \eta_{WB1}$ has no skill in determining $U_{MO}$ could partly be related to the fact that the Jason altimeter (or Topex/Poseidon) takes no measurements between Abaco and WB2. However, since there is a Jason ground track four km east of WB2, we consider the relatively low level of the variance of $U_{WB2-EAST}$ explained by $\eta_{EAST} - \eta_{WB2}$ to be a robust result.

It is interesting to note that it is particularly the low-frequency transport variability of periods in excess of two months that is gradually suppressed, as the western end-point of the section is moved closer to the western boundary. The reduction in rms transport variability is a factor four comparing two-month low pass filtered time series of $U_{WB5-EAST}$ and $U_{MO}$ from 10.2 Sv to 2.5 Sv rms (Table 2). For 2-month high-pass filtered data while there is only a reduction by a factor of 2 between $U_{WB5-EAST}$ and $U_{MO}$ from 3.0 Sv to 1.7 Sv rms (Table 2). This will become relevant in section 6.

To investigate why the correlation between $U_{WB2-EAST}$ and $\eta_{EAST} - \eta_{WB2}$ is lower than the correlation between $U_{WB5-EAST}$ and $\eta_{EAST} - \eta_{WB5}$, an EOF mode analysis is carried out. The first three leading EOF modes for $U_{MO}(z)$, $U_{WB2-EAST}(z)$, $U_{WB3-EAST}(z)$, and $U_{WB5-EAST}(z)$ (which are the transport per-unit-depth profiles in the upper 1000 m corresponding to the net upper-ocean transports defined in [8, 11, 12, 13]) are displayed in Fig. 10 a-d. The variances explained by the first modes are 69, 76, 83 and 92%, respectively, while those of the second modes amount to 22, 18, 17 and 7%.

For all four transport cases the first modes (blue lines in Fig. 10 a-d) do not exhibit a zero crossing between the sea surface and 1000 m. Therefore, the transport at the sea surface associated with the first mode is of the same sign as the upper-ocean transport explained by the first mode (i.e. the vertical integral of that mode over the
top 1000 m). All of the second modes (red lines in Fig. 10 a-d) display a zero crossing shallower than 250 m. As a result transports associated with the second modes at the sea surface have opposite signs to the vertically integrated transports. The first modes of $U_{WB5-EAST}(z)$, $U_{WB3-EAST}(z)$, $U_{WB2-EAST}(z)$ and $U_{MO}(z)$ explain 10.8, 5.9, 3.5 and 2.8 Sv rms above 1000m, respectively, according to,

$$\langle U_{WB5-EAST}^i \rangle^{\frac{1}{2}} = \langle A_{WB5-EAST}^i \rangle^{\frac{1}{2}} \int_{-1000}^{0} M_{WB5-EAST}^i(z)dz,$$

while the second modes accounts for 0.8, 0.9, 1.0 and 1.2 Sv rms, respectively. Here, $M^i$ denotes the i-th EOF mode and $A^i$ the corresponding principal component, and the brackets represent variance operators.

The gradient in sea surface height is regarded as a measure of the surface geostrophic flow. For the former to be also a good indicator of depth-integrated upper ocean transport requires surface and upper ocean transport to fluctuate in phase. It is not surprising that $\eta_{EAST} - \eta_{WB5}$ is a good proxy of $U_{WB5-EAST}$ (Fig. 9) given that the first EOF mode explains 92 % of the variance (Fig. 10 d). Flow reversals in the upper 1000 m - as represented by the second EOF modes - gradually gain in importance as the western endpoint of the transport integration is moved westward. As has been shown above, all of the second modes imply an out-of-phase relationship between surface and depth-integrated upper-ocean transport. In the particular case of the Bahamas western boundary, this has to do with the superposition of the upper ocean Antilles Current on the DWBC below, which have been shown to vary largely independently from one another (Lee et al., 1990; 1996). Further, the core of the Antilles Current, which is found near 400 m in the mean, is not strongly coupled to variations in the surface currents just above it (op. cit.).
The change of the modal structure over the continental slope implies that the
clear relationship between surface and upper-ocean transport found offshore is
gradually lost. As a consequence, variations in $\eta_{\text{EAST}} - \eta_{\text{WB2}}$ do not provide a reliable
estimate of the temporal evolution of $U_{\text{WB2-EAST}}$ (Fig. 9). In the case of $U_{\text{MO}}$ the second
mode has an even higher relative importance than the first one, when compared to
$U_{\text{WB2-EAST}}$ (Fig. 10 a,b). The fact that altimetric measurements effectively do not exist
west of WB2 represents an additional practical problem, when trying to infer the
temporal evolution of $U_{\text{MO}}$ (as part the AMOC) from the basin-wide gradient of sea
surface height.

4. The signature of eddies in basin-wide integrated transports – a model
simulation

Why are upper ocean transport variations integrated between the western and
eastern boundaries so much smaller than WB5-to-eastern boundary transports? And
what is the dynamical reason for the sharp decline in rms variations of sea surface
height anomalies within 100 km of the western boundary? Analysis of the altimeter
data suggests that westward propagating eddies are a major source of variability near
the western boundary (Fig. 5). In this section we present numerical results from a non-
linear reduced-gravity model. We aim at elucidating the fate of Rossby waves and
eddies as they approach the western boundary, and to investigate how they affect the
sea surface height near and on the boundary. This allows us to put the observations
into a dynamical context.

The model is similar to that described in Johnson and Marshall (2002) but is
eddy permitting with a lateral resolution of 14 km. The domain extends 4500 km in
the meridional direction (with the southern boundary at the equator) and 2800 km in
the zonal direction. A beta-plane approximation is made, with $f_0 = 0.65 \times 10^{-4}$ s$^{-1}$
equal to the Coriolis parameter at 26.5°N. The background model layer thickness is
750 m, and the reduced gravity is 0.015 m s$^{-2}$ (broadly consistent with the North
Atlantic density distribution). Dissipation is achieved through lateral friction, with a
coefficient of 30 m$^2$ s$^{-1}$, and no-slip and no-normal flow boundary conditions are
applied. A sponge region ramps up over the southernmost 1000 km of the domain, in
which layer thickness is relaxed to the background value at a maximum rate of
1/6 hours$^{-1}$; this efficiently damps any fast boundary waves as they approach the
equator, preventing them from propagating right around the basin and contaminating
the results.

The first integration (Fig. 11) is initialized with a single, anticyclonic eddy,
located roughly in the centre of the domain at $y=0$ (equivalent to 26.5°N). The eddy
has an approximate diameter of 500 km, and a layer thickness anomaly of 130 m,
equivalent to a sea surface height anomaly, $\eta' = (g'/g)h'$, of roughly 20 cm, as plotted
in Fig 11a. The eddy propagates westward at a speed of about 5 cm s$^{-1}$ (Fig 11b)
although it is somewhat dispersive. On reaching the western boundary, a fast Kelvin
wave is generated which propagates rapidly southwards; subsequently the eddy itself
propagates northwards along the boundary due to the nonlinear image effect (Minato,
1982) and rocket effect (Nof, 1988) and short Rossby waves are also generated. The
maximum amplitude of the free surface height disturbance, $\eta$, on the boundary is
roughly a factor of four smaller than that in the basin interior.

Fig. 11c & d show integrated meridional transports within the moving layer of

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the reduced gravity model, as a function of latitude and time. In Fig. 11c the transport
is integrated from 56 km off the western boundary to the eastern boundary ($U_{56KM-EAST}$); Fig. 11d shows the meridional transport integrated coast-to-coast, from the
western to the eastern boundary ($U_{WEST-EAST}$). These two transport values differ widely
in terms of both their structure and their amplitude. While $U_{56KM-EAST}$ is dominated by
the signature of the northward moving eddy close to the western boundary, $U_{WEST-EAST}$
is dominated instead by the much smaller amplitude, southward propagating boundary
wave signals on the western boundary itself. It is clear from these two panels that the
net northward transport anomaly associated with an individual eddy is very different
when the western endpoint of the section lies a short distance off the boundary
compared with on the boundary. This is a direct consequence of the reduction in
amplitude of variations in $\eta$ at the boundary.

Fig. 12 shows the same diagnostics for a calculation in which the model is
initialized with a field of cyclonic and anticyclonic eddies (Fig 12a). The Hovmoeller
plot shows the westward propagation of the eddies along $y=0$ and subsequent
propagation along the western boundary (Fig. 12b); again, the maximum amplitude of
$\eta$ on the boundary is smaller than that in the basin interior. This is consistent with the
observed reduction in the amplitude of dynamic height variability between site C and
WB2 (Fig. 6). As with the single eddy, $U_{56KM-EAST}$ exhibits large-amplitude variability
(Fig. 12c) most of which disappears in $U_{WEST-EAST}$ (Fig. 12d) the remaining transport
anomaly being associated with boundary waves generated when each eddy reaches
the western boundary. Fig.13a shows variations of $\eta$ along 26.5°N within 500 km
from the western boundary (blow-up of image detail in Fig. 12b). Like in the
observation (Fig. 5) the simulated $\eta$ anomalies display maximum amplitudes away
from the western boundary. The rms variability of $\eta$ peaks at 100 km offshore (Fig. 13b) and there is then a decay of roughly a factor of five toward the western boundary. Both the scale over which the decay takes place and the degree of the decay are in good agreement with the observations (Fig. 6).

Note that the 1.5 layer reduced gravity model used here includes no representation of topographic effects or boundary complexity, and as such does not accurately reproduce the details of the boundary propagation, which in more complex models involves a combination of Kelvin, mixed Kelvin-Rossby and topographic Rossby wave dynamics (Huthnance 1978, Clarke and Shi 1991, Hallberg et al. 1996). However, we believe that the fact that the model exhibits a similar decay in the amplitude of variability as that observed, and over a similar distance from the boundary, suggests that these details are not necessary to explain the observed behavior. O'Rourke (PhD thesis, Liverpool 2009) finds that the presence of topography in fact increases the speed of coastally-trapped waves; provided that there are boundary waves with a group speed significantly faster than the interior Rossby wave speed, then a reduction in the amplitude of variability close to the boundary will occur.

The “eddy-filled ocean” experiment in Fig. 12 also highlights a potential attribution problem. The fast equatorward communication of transport anomalies may make it very difficult to distinguish between local and remotely forced fluctuations of $U_{MO}$ (or the AMOC) because - at any given latitude – in addition to local anomalies $U_{MO}$ also contains transport contributions resulting from eddies impinging on the western boundary poleward of the latitude of interest.
5. A theory for the reduction of SSH fluctuations close to the western boundary

The observations are in good qualitative agreement with the reduced-gravity model. In addition the model has highlighted boundary waves as a potential mechanism for reducing variability in $\eta$ at the boundary. On the basis of linear dynamics we now derive a relationship between the amplitude of $\eta$ (or layer thickness anomaly) on the western boundary and the amplitude offshore.

First consider an analytical model of linear Rossby waves in a $\beta$-plane basin. In the limit $\omega \ll \beta L$, the reduced gravity equations admit long and short Rossby wave solutions $^2$, satisfying the dispersion relations $\omega/k_l = -\beta L_d^2$ (for long waves) and $\omega k_s = -\beta$ (for short waves) where $\omega$ is the angular frequency, $k_l$ and $k_s$ are the long and short wavenumbers, $L_d$ is the deformation radius, and $\beta$ is the meridional gradient of $f$.

Now suppose that incoming long Rossby waves are balanced by short Rossby waves generated at a meridional western boundary ($x = 0$). The total layer thickness anomaly is thus

$$h' = A_l(y)e^{ik_l x - \omega t} + A_s(y)e^{ik_s x - \omega t}.$$  

The wave amplitudes, $A_l(y)$ and $A_s(y)$, may be complex and include phase information. The zonal component of velocity is zero at the western boundary. Thus, the linear meridional momentum equation (e.g. Cushman-Roisin, 1994, eq. 12-35, for the non-linear case) takes the following form on the boundary

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$^2$ Clarke and Shi (1991) discuss this scaling. The distinction between long and short waves only works if this condition is met. Waves with higher frequencies become trapped at the boundary as Kelvin waves.
Here $g'$ is the reduced gravity, we assume that the meridional velocity is geostrophic to leading order, and we neglect any frictional dissipation. Substituting for $h'$, and noting $k_s \gg k_l$, we find

$$\frac{\partial}{\partial t} \left( \frac{g'}{f} \frac{\partial h'}{\partial x} \right) + g' \frac{\partial h'}{\partial y} = 0. \tag{15}$$

This, in turn, may be rewritten,

$$-\frac{\beta A_s f}{f} + \frac{d}{dy} (A_l + A_s) \approx 0. \tag{16}$$

Finally, assuming that the total wave amplitude, $A_s + A_l$, vanishes to the north of the incoming long waves, and integrating the above expression, we obtain

$$A_l + A_s \approx f \int \frac{\beta A_l}{f^2} dy - \frac{\beta \Delta y}{f} |A_l|. \tag{17}$$

Thus, the total wave amplitude on the boundary, $A_s + A_l$, is reduced relative to the interior by a factor $\beta \Delta y / f$, equal to the fractional change in Coriolis parameter across the band of incoming long waves. The amplitude reduction is thus greater the smaller the meridional scale of the incoming wave (or eddy). If the incoming waves are confined to a small latitude band then the wave amplitude, and hence the variability in sea surface height, is greatly reduced on the boundary. For $\Delta y = 500$ km the reduction will be by a factor of $f / (\beta \Delta y) = 6$ (using $f_{26.5^oN} = 6.4 \times 10^{-5} \text{ s}^{-1}$ and $\beta = 2.1 \times 10^{-11} \text{ s}^{-1} \text{ m}^{-1}$).

Physically, this amplitude reduction occurs because an along-boundary pressure gradient is unable to be maintained, since the Coriolis force to balance it would require a flow into the boundary. Instead a fluid acceleration is generated, which
rapidly propagates any pressure anomalies along the boundary. This is precisely the same mechanism that gives rise to coastal Kelvin waves, although the character of the wave motion is different in this Rossby wave limit (Clarke and Shi, 1991). The result that the pressure anomaly is reduced on the boundary holds across a wide range of physical parameters, although precise details depend on frictional dissipation, boundary conditions and wave frequency (Marshall and Johnson, 2008).

6. Discussion

a. Estimating the western boundary eddy contribution to upper mid-ocean transports from altimetry

Wunsch (2008) attempted to estimate the rms variability of $U_{MO}$ by projecting variations in $\eta$ onto dynamical vertical modes of horizontal velocities. He argued that $\eta$ near the western boundary of the subtropical Atlantic displayed typical variations of 16 cm rms while exceeding those at the eastern boundary by a factor of four. Upon making assumptions about the spectral properties of $\eta$, he simulated an 80-year-long time series of monthly values of $U_{MO}$ (excluding Gulf Stream and Ekman transports) which exhibited fluctuations of 16 Sv rms, distributed over a broad range from intra-seasonal to decadal time scales. According to Wunsch (2008) the variations in $U_{MO}$ were due to eddy variability with a decorrelation scale of not more than several tens of kilometres. He argued that the high degree of eddy variability in his simulation had the potential to mask possible climate change related trends in the strength of the MOC (Schmittner et al., 2005) for more than a century.
Our results imply that variability in $\eta$ of 16 cm rms may be regarded as typical of conditions found at 26.5°N between 100 and 500 km away from the western boundary, but not of conditions over the steep continental slope. Both altimetry and our measurements of dynamic height suggest that a value between three and five cm rms is appropriate. Wunsch assumes that the amplitude of $\eta$ of 16 cm rms is linked to the amplitude of $U_{MO}$ which readily explains why his estimate of the amplitude of $U_{MO}$ of 16 Sv rms is too large by a factor of three to five, when compared to the 3.0 Sv rms we have observed. The actual impact of eddy-variability on the observed intra-seasonal and seasonal variability of $U_{MO}$ is much smaller than implied by Wunsch's interpretation.

We have shown from in-situ observations, altimetry, and numerical modelling that there is a rapid westward decrease in the variability of $\eta$ within 100 km of the Abaco shelf. Boundary waves play an important role in explaining this decline, as they provide a fast mechanism for along-boundary export of transport anomalies originating from eddies and Rossby waves impinging on the western boundary. The statistical approach of Wunsch (2008) does not include western boundary dynamics and therefore cannot account for the decline.

Wunsch (2008) argues that an observed time series of $U_{MO}$ of only 12 months length (Cunningham et al., 2007) may display significantly less variability than the same timeseries spanning years or decades. Over the RAPID observational period (04/2004 to 10/2006) the variability of $\eta$ amounts to 3.8 cm rms at mooring site WB2 corresponding to $U_{MO}$ variations of 3.0 Sv rms. The 15-year-long timeseries of $\eta$ from (01/1992 to 01/2008) has an rms variability of 5.3 cm. We estimate that $U_{MO}$ spanning one or two decades could display fluctuations of $3Sv \times (5.3cm/3.8cm) = 4.2Sv$. We
therefore expect the actual influence of eddy variability on a decade-long time series of $U_{MO}$ to be four times less than suggested by Wunsch – and possibly smaller, since not all the contributions to variability in $U_{MO}$ stem from eddy-generated density variations at the western boundary.

The standard error of the mean gives an indication of changes in $U_{MO}$ that can be resolved over any interval ($SE = \sigma/\sqrt{DOF}$ where $\sigma$ is standard deviation and DOF is the number of degrees-of-freedom). For the 30-month record (Table 2) we estimate that there are 22 degrees-of-freedom – or effectively nine independent measurements per year. Assuming measurement errors to be negligible, we estimate that we could resolve year-to-year changes of 1.4 Sv $[(3.0^2 / 9 \times 2)^{\frac{1}{2}}]$ and 0.6 Sv $[(4.2^2 / 90 \times 2)^{\frac{1}{2}}]$ for decade-to-decade changes.

For $U_{WB5\text{-}EAST}$ the western end point of the zonal transport integration is located east of the western boundary Antilles current system (Johns et al, 2008) and therefore it may be taken as a measure of the wind-driven subtropical gyre at 26.5° N. Due to much larger and longer-period transport variations in $U_{WB5\text{-}EAST}$ compared to $U_{MO}$ only changes exceeding 6.2 Sv on interannual time scales and 2.2 Sv on decadal time scales can likely be resolved. It seems probable that year-to-year changes in $U_{WB5\text{-}EAST}$ are dominated by eddy-noise, which may mask interannual changes in the Sverdrup gyre.

b. Eddy contribution to the meridional overturning circulation estimated from an eddy resolving numerical model

Fig. 14 shows AMOC ($U_{MO}$ plus Ekman plus Gulf Stream) variations at 26°N from 1988 through 2004 from an integration of the 1/12° OCCAM ocean general
circulation model that is driven by realistic wind and buoyancy forcing (Marsh et al., 2008). The 30-day low-pass filtered time series (black line) displays fluctuations of 3.0 Sv rms.

In the subtropical North Atlantic it is assumed that the meridional coherence scale of zonal wind stress exceeds 100 km (Josey et al., 2002) and that eddy-related density signals exhibit de-correlation scales of several tens of kilometres. We also expect the buoyancy driven AMOC signal to have a meridional coherence exceeding 100 km. Therefore, the difference between AMOC fluctuations between 26° N and 27° N likely reflect the influence of eddies on each section and it amounts to 1.5 Sv rms (blue line in Fig. 14). For 365 day low-pass filtered values (red line in Fig. 14) the variability in the AMOC difference goes down to 0.9 Sv. Thus, there is nothing reminiscent of pronounced interannual eddy-related variability in the MOC in this model.

The small eddy contribution to the MOC variability found in OCCAM is consistent with results from Biastoch et al. (2008) who show that eddies only leave a small imprint on the MOC at 26.5°N in their 1/12° model. Other eddy-permitting models also exhibit a decrease of the SSH variability close to the western boundary (e.g. Hurlburt and Hogan, 2000; Maltrud and McClean, 2005) and therefore suggest an attenuation of the eddy variability at the western boundary itself, in agreement with the observation presented in this study (Fig. 6)

c. AMOC monitoring using altimetry?

Within 100 km kilometres of the western boundary there is a rapid westward reduction the variance of upper-ocean transport that is explained by the linear
regression to the zonal $\eta$ difference ($U_{WBX\text{-}EAST} = (\eta_{EAST} - \eta_{WBX}) * s_{WBX}$). This finding is in line with results from Hirschi et al. (2008) who studied the problem using an eddy-permitting numerical model. In their model simulation, $\eta$ is available at each grid point and in the model $\eta$ near the coast is not degraded by land effects. The agreement between observations and numerical modelling strengthens the argument that there is a dynamical reason for the failure of the relationship between the boundary-to-boundary gradient in $\eta$ and $U_{MO}$. For the relationship to fail means that no stable (i.e. time-invariant) relationship between surface geostrophic flows (as measured by the gradient in $\eta$) and $U_{MO}$ exists. The gradual loss of this relationship - as the western end point approaches the boundary has been shown from our mode analysis to be due to the a gradual increase in the importance of the second mode (at the expense of the first mode). For the second mode the surface geostrophic transport has the opposite sign to that of the depth-integrated upper-ocean transport. We speculate that this is a result of (i) complicated mixed barotropic / baroclinic flow over sloping bathymetry, (ii) the gradual suppression of the eddy variability and, (iii) the more pronounced influence of coastally-trapped waves near the boundary.

More sophisticated methods of using $\eta$ to infer $U_{MO}$ (or the strength of the MOC) should be tested as the benefit of a successful technique would be high: we could establish a relationship to derive a continuous time series of the AMOC at 26.5°N from the starting point of the TOPEX/POSEIDON mission in 1992, given that daily estimates Gulf Stream transport and Ekman transport are also available for this period.

Model simulations seem to imply the observed pronounced multi-decadal variations of sea surface temperature over the North Atlantic (e.g. Mann et al., 1995)
are directly linked to changes in the strength of the MOC (Latif et al., 2004; Knight et al., 2005). While observational evidence for this link is missing because a reliable MOC time series spanning several decades does not exist, it is possible that on multi-decadal time scales $\eta$ averaged over the North Atlantic may serve as a proxy of MOC variations.

The 0.3° resolution of the gridded AVISO data set may represent a practical limitation of the altimetry in estimating the rapid westward decrease in $\eta$ over 100 km. Measurements near the coast have to be viewed with caution, as standard tidal and atmospheric corrections may have larger errors than offshore, and faster and smaller-scale dynamics than encountered in the open ocean might add additional challenges to the interpretation of coastal data (Bouffard et al., 2008). However, we interpret the finding that the rms variability in $\eta$ at mooring site WB2 is in agreement with the density observations as strong indication that $\eta$ is not particularly affected by these problems at this location.

7. Summary

The main results of this study can be summarized as follows. The meridional upper mid-ocean transport $U_{MO}$ integrated between Morocco and the Bahamas varies by 3.0 Sv rms between April 2004 and October 2006. Zonally integrated upper-ocean transports between Morocco and stations 16, 40 and 500 km off the shelf of the Bahamas show fluctuations of 3.6, 6.0 and 10.8 Sv rms, respectively, and thus exceed those of $U_{MO}$. In agreement with that, both the mooring and altimetric measurements of sea surface height variability show a threefold decline within roughly 100 km of (and towards) the shelf of the Bahamas. The findings imply that in order to capture
the variability of $U_{MO}$ (and the AMOC) reliably in-situ density measurements need to be taken right at the ocean boundaries.

775 There is a gradual decrease in correlation between upper-ocean transport (integrated between the eastern boundary and different sites within 500 km of the western boundary) and the difference in $\eta$ between the corresponding section endpoints, as the western endpoint moves westward. The correlation coefficient drops from 0.85, for a station located 500 km offshore, to a statistically insignificant value of only 0.12 for the basin-wide integration. We consider the gradual decrease in correlation - as the western section endpoint is moved westward - to be mainly a consequence of changes in the vertical structure of the integrated flow, because of flow reversals in the upper 1000 m becoming increasingly important over the continental slope compared to offshore conditions. The results imply that the basin-wide difference of $\eta$ does not provide reliable estimates of the temporal evolution of $U_{MO}$.

780 The non-linear reduced-gravity model used to study eddy-boundary interaction is able to reproduce both the observed decline in the amplitude of $\eta$ within 100 km from the western boundary and the corresponding reduction in the integrated transport fluctuations. It shows that eddies impinging on the western boundary kick off fast boundary waves that propagate equatorward and account for the observed decline.

790 An analytical linear model based on a simple budget of incoming long and reflected short Rossby waves at the western boundary provides a scaling for the reduction of thermocline thickness (or $\eta$) on the western boundary compared to offshore conditions.
according to $f/ (\beta \Delta y)$. At 26.5°N and for a meridional scale $\Delta y = 500$ km of the incoming wave yields a decrease by a factor of 6.

Physically, this amplitude reduction occurs because an along-boundary pressure gradient cannot be maintained by a Coriolis force (since there cannot be flow into the boundary). Instead a fluid acceleration is generated, which rapidly propagates any pressure anomalies along the boundary.

Overall, our results suggest that the local eddy field does not have the potential to dominate AMOC variability at 26.5°N on interannual to decadal time scales. If a monitoring system based on boundary–to–boundary density gradient observations can be maintained in the ocean continuously over climate relevant time scales, it will document the temporal evolution of the strength of the AMOC, whether it be random fluctuations, regular oscillations or a long-term trend.

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from CERSAT, at IFREMER, Plouzané (France). We would like to thank Darren Rayner (NOCS) for coordinating and planning the seagoing activities and leading the mooring data recovery. The competent help from Julie Collins and John Molina in processing the mooring data is highly appreciated. The OCCAM MOC time series has been made available to us by Bob Marsh. Two of the authors (TK, JJMH) were supported in the framework of the NERC funded Rapid Climate Change Programme (Grant NER/T/S/2002/00481). HLJ is funded by a Royal Society University Research Fellowship, for which she is grateful.
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Table 1: Instrument depths of MicroCAT CTDs and current meters on the different moorings at the western boundary during the different deployment periods.

<table>
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<tr>
<th>Variable</th>
<th>transport standard deviation [Sv]</th>
<th>D.O.F</th>
<th>Corr. to $U_{MO}$</th>
<th>Corr. to $\eta_{EAST} - \eta_{WBX}$</th>
<th>rms error using $\eta_{EAST} - \eta_{WBX}$ [Sv]</th>
<th>variance explained by $\eta_{EAST} - \eta_{WBX}$ [%]</th>
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<td>1</td>
<td>0.12 / WB1</td>
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<td>0.44 / WB2</td>
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<td>0.71 / WB3</td>
<td>4.2</td>
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<td>15</td>
<td>0.10</td>
<td>0.85 / WB5</td>
<td>5.7</td>
<td>72</td>
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Table 2: Transport statistics for upper-ocean transport (shallower than 1000 m).

Columns 2 - 4 show transport standard deviation (the values in bracket correspond to
the 60 day low-pass and high-pass filtered time series: degrees of freedom for the 30 month records, and correlation of the transport to upper mid-ocean transport $U_{MO}$ respectively. Column 5 indicate the correlation of transports with $\eta$ at the mooring locations. Column 6 shows rms transport error if $\eta$ is used to predict transport, assuming a linear relation

$$U_{WBX\text{-EAST}} = s_{WBX} \cdot (\eta_{EAST} - \eta_{WBX})$$

Column 7 gives the transport variance [Sv] explained by $\eta$ based on the aforementioned linear relationship. Calculations are based on 10 – day low pass filtered data. The integral time scales underlying the degrees of freedom in column 3 represent the integral of the auto-correlation function to the first zero crossing.
Fig. 1: Panel A: The distribution of moorings along 26.5°N in the subtropical North Atlantic. Panel B: Location of western boundary moorings WB1 – WB3 (Crosses). Black dots indicate Topex / Poseidon (Jason-1) ground track.
Fig. 2: Section of density (and bottom pressure) moorings along 26.5°N. The current meter moorings west of WB2 are not shown here for clarity (see Fig. 3).

Fig. 3: Moorings near the western boundary (off Abaco, the Bahamas). Density sensors, bottom pressure recorders and current meters are denoted as crosses, squares and circles, respectively. The dots at WBA and WB0 indicate the part of the water column covered by Acoustic Doppler Current Profiler (ADCP) measurements. WBH1 and WBH2 were only deployed during the period 04/2004 to 04/2005.
Fig. 4: 48 hour low-pass filtered time series of bottom pressure fluctuations (time mean removed) at WB3 and WB5. An offset of 0.1 dbar has been added to the WB5 pressure record.
Fig. 5: Left: Sea surface height anomalies $\eta$ [cm] along 26.5°N in the Atlantic from October, 1992, to January, 2008. The zonal mean (Bahamas to Morocco) has been subtracted at each time step. Right: Same as left panel but 730-day low-pass filtered. The red lines in both panels indicate the 04/2004 to 10/2006 interval during which mooring measurements at WB2, WB3 and WB5 were made. The blue line marks the interval of the density measurements made at mooring C from the ACCP-3 experiment (Johns et al., 2005) deployed at 26.5°N and 76.1°W between 10/1995 and 06/1997.
Fig. 6: RMS amplitude of $\eta$ along 26.5° N for the intervals 10/1992 to 01/2008 (solid black line: 04/2004 to 10/2006 (red dashed line) and 10/1995 to 06/1997 (blue dashed line). Also shown is the amplitude of rms dynamic height fluctuations (dyn. cm; i.e., geopotential anomaly divided by the Earth's gravitational acceleration) at 200 m determined from the mooring density measurements at WB2, WB3 and WB5 (red crosses). The blue cross denotes dynamic height computed from density at mooring “C” of the ACCP-3 experiment (Johns et al., 2005). The green line show the rms amplitude of $\eta$ along 26.5° N for 2 year low-pass filtered data for the interval 10/1992 to 01/2008.
Fig. 7: Fluctuations of northward mid-ocean transport above 1000 m ($U_{MO}$) between April 2004 and October 2006, 10-day low-pass filtered using a 6-pole Butterworth filter.
Fig. 8: Northward transport fluctuation [Sv] above 1000 m east of WB2 ($U_{WB2\_EAST}$ green: east of WB3 ($U_{WB3\_EAST}$ red) and east of WB5 ($U_{WB5\_EAST}$ blue). For better visualization the transports are offset by 25 Sv (dashed lines represent the zero lines). Each of the three grey lines represents fluctuations of the upper mid-ocean transport (to which the same offset has been applied to facilitate comparisons to the coloured time series).
Fig. 9: Transports [Sv] of $U_{MO}$ (black), $U_{WB2-EAST}$ (green), $U_{WB3-EAST}$ (red) and $U_{WB5-EAST}$. Superimposed (orange) are fluctuations of the difference in sea surface height [cm] between the eastern boundary ($27^\circ 54.5'$ N & $13^\circ 22.0'$ W) and the locations of WB1, WB2, WB3 and WB5 extracted from the gridded altimeter data set ($\eta_{EAST} - \eta_{WBX}$ with X=5, 3, 2 and 1 denoting the different WB mooring sites). The difference $\eta_{EAST} - \eta_{WBI}$ was used to compare to $U_{MO}$ rather than $\eta$ at the Abaco shelf break (see text).
Fig. 10: First three EOF modes for 10-day low-pass filtered transport per-unit-depth profiles of $U_{MO}$ (panel A), $U_{WB2-EAST}$ (panel B), $U_{WB3-EAST}$ (panel C) and $U_{WB5-EAST}$ (panel D). The time averaged transport profile has been removed prior to the computation of the modes. The modes exhibit units Sv m$^{-1}$, since they have been multiplied by the standard deviation of the corresponding principal components. This way the amplitudes of the different modes can be compared at each depth level. The legends indicate the variance explained by each mode.
Figure 11: Results of a numerical calculation with a reduced-gravity model, initialized with a single anticyclonic eddy of 500 km in diameter at $y=0$ (equivalent to 26.5°N) and 1400 km from the western boundary. Panel (a) shows the initial sea surface height $\eta$ in cm (contour interval 2 cm). Panel (b) shows a Hovmoeller diagram of $\eta$ along the dashed line in panel (a) along $y=0$ and then down the western boundary (contours at ±0.5, ±1, ±2 and then every 2 cm). Panel (c) displays a Hovmoeller plot of net northward transport within the upper model layer, integrated zonally from 56 km away from the western boundary to the eastern boundary, as a function of latitude and
time (contour interval 2 Sv). Panel (d) shows the net northward transport integrated
from coast to coast, again as a function of latitude and time; the color scale is the
same as in panel (c) and the contour interval is 0.2 Sv.

Figure 12: As in Figure 11, but for an integration initialized with a field of cyclonic
and anticyclonic eddies. Contour intervals are (a) 4 cm, (b) ±1, ±4 then every 4 cm, (c)
8 Sv and (d) 2 Sv.
Fig. 13: (a) Hovmöller diagram of $\eta$ along 26.5°N within 500 km of the western boundary, and (b) rms fluctuations of $\eta$ from panel (a) as a function of distance from the western boundary.
Fig. 14: Strength of the AMOC at 26°N from 1988 through 2004, defined as maximum upper ocean northward transport in the OCCAM 1/12 degree model for 30-day low-pass filtered values (black line). The blue and red lines represent the differences in AMOC strength between 26°N and 27°N for 30-day and 365-day low-pass filtered values.