Variability of the meridional overturning circulation at the Greenland–Portugal OVIDE section from 1993 to 2010

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A B S T R A C T
The meridional overturning circulation (MOC) in the North Atlantic transports heat from the subtropics to high latitudes and hence plays an important role in the Earth’s climate. A region crucial for the MOC is the northern North Atlantic and the adjacent Nordic Seas, where waters transported northward in the MOC upper limb progressively cool, gain density and eventually sink. Here we discuss the variability of the gyre circulation, the MOC and heat flux as quantified from a joint analysis of hydrographic and velocity data from six repeats of the Greenland to Portugal OVIDE section (1997–2010), satellite altimetry and Argo float measurements. For each repeat of the OVIDE section, the full-depth absolute circulation and transports were assessed using an inverse model constrained by ship-mounted Acoustic Doppler Current Profiler data and by an overall mass balance. The obtained circulation patterns revealed remarkable transport changes in the whole water column and evidenced large variations (up to 50% of the lowest value) in the magnitude of the MOC computed in density coordinates (MOCr). The extent and time scales of the MOCr variability in 1993–2010 were then evaluated using a monthly MOCr index built upon altimetry and Argo. The MOCr index, validated by the good agreement with the estimates from repeat hydrographic surveys, shows a large variability of the MOCr at OVIDE on monthly to decadal time scales. The intra-annual variability is dominated by the seasonal component with peak-to-peak amplitude of 4.3 Sv (1 Sv = 10⁶ m³ s⁻¹). On longer time scales, the MOCr index varies from less than 15 Sv to about 25 Sv. It averages to 18.1 ± 1.4 Sv and shows an overall decline of 2.5 ± 1.4 Sv (95% confidence interval) between 1993 and 2010. The heat flux estimates from repeat hydrographic surveys, which vary between 0.29 and 0.70 ± 0.05 PW, indicate that the heat flux across the OVIDE section is linearly related to the MOCr intensity (0.054 PW/Sv).

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1. Introduction

Climate models suggest that the continuing increase in human emissions of carbon dioxide to the atmosphere will cause a slowdown of the Atlantic meridional overturning circulation (MOC) in the 21st century (Schmittner et al., 2005; Weaver et al., 2007; IPCC, 2007). The associated decline in the meridional oceanic heat flux will likely modify the heat loss from the ocean to the atmosphere in the North Atlantic region and affect the climate of the northern hemisphere (e.g., Vellinga et al., 2002; Latif et al., 2007; Semenov et al., 2010). The climate simulations also show that the Atlantic MOC is subject to energetic natural variability occurring on a variety of time scales (e.g., Schmittner et al., 2005), which masks the low-frequency signal and renders the detection of the predicted MOC decline challenging (Zhang et al., 2010).

Insights about the MOC decadal variability in the North Atlantic have been gained from model studies (e.g., Eden and Willebrand, 2001; Gulev et al., 2003; Marsh et al., 2005; Böning et al., 2006; Balmaseda et al., 2007; Zhang, 2010; Huang et al., 2012). Ocean models forced by atmospheric reanalysis (Marsh et al., 2005; Böning et al., 2006; Balmaseda et al., 2007; Huang et al., 2012; Desbruyères et al., 2013) have consistently suggested that the Atlantic MOC weakened by a few Sv (1 Sv = 10⁶ m³ s⁻¹) since the mid-1990s in response to a weakening of the surface forcing associated with the decline in the North Atlantic Oscillation (NAO) index. The model studies have more broadly suggested a positive correlation on interannual to decadal time scales between the NAO, the MOC, the deep convection intensity in the Labrador Sea and the strength...
of the subpolar gyre cyclonic circulation (Eden and Willebrand, 2001; Gulev et al., 2003; Böning et al., 2006).

Because of the scarcity of historical transoceanic measurements and despite the recent progress in the development of observational networks, only a few observation-based estimates of the MOC variability in the North Atlantic are currently available. Based on the analysis of a set of hydrographic sections carried out in the North Atlantic in the late 1950s, early 1980s and early 1990s, Kottmann et al. (1999) have reported coherent decadal variability of the MOC at 48°N and 36°N, as well as the lack of significant decadal signal at 24.5°N. These authors suggested that in the 1950s – 1990s the MOC at 48°N and 36°N was negatively correlated with the deep convection intensity in the Labrador Sea and positively correlated with the southward transport of the overflow-derived deep waters. Focusing on the period from 1993 to 2000, Lumpkin et al. (2008) have found no significant trend in the MOC magnitude from the analysis of 5 repeats of the 48°N section. By combining satellite altimetry and Argo profiling float data, Willis (2010) has constructed a time series of the transport in the MOC upper limb at 41°N. No trend in the MOC magnitude from 2002 to 2009 has been found, while altimetry data alone have suggested strengthening of the MOC by 2.6 Sv since 1993. A long-term slowdown of the MOC from 1957 to 2004 has been inferred by Bryden et al. (2005) from the analysis of 5 repeats of the 24.5°N section. This result has however been recently revisited based on data obtained from a transatlantic array of moored instruments deployed in the framework of the RAPID program, which has provided a continuous MOC time series at 26.5°N with daily resolution since 2004 (Cunningham et al., 2007). Using these data, Kanizow et al. (2010) have concluded that the aliasing due to the seasonal signal in the MOC magnitude possibly accounts for most of the trend reported by Bryden et al. (2005). Recently, based on data from a moored array located further to the south, at 16°N, Send et al. (2011) have inferred a ~3 Sv weakening of the MOC from January 2000 through June 2009.

In the northern North Atlantic, some variability of the MOC between August–September 1997, June–July 2002 and June–July 2004 has been reported by Lherminier et al. (2007, 2010) based on data from 3 repeats of the A25-OVIDE Greenland to Portugal section. The 2002–2008 mean summer (May–September) state of the MOC has been estimated from hydrographic measurements combined with satellite altimetry data. However, no information about the decadal variability of the MOC in the northern North Atlantic is available from observations.

In this study, we made an assessment of the variability of the MOC and heat flux across the A25-OVIDE Greenland to Portugal line in the northern North Atlantic (Fig. 1) over 1993–2010 from 6 occupations of the line (1997–2010) in conjunction with satellite altimetry (1993–2010) and Argo float (2002–2010) data sets. As an additional source of hydrographic data extending back to 1993, we used the World Ocean Atlas (WOA) temperature and salinity fields. While earlier studies of the MOC variability in the extra-tropical North Atlantic have employed either repeat full-depth hydrography (e.g. Lumpkin et al., 2008; Lherminier et al., 2007, 2010) or a combination of satellite and float measurements (Willis, 2010), we took advantage here of all these three data sources. Based on CTD and absolute velocity data from the 6 repeats of the A25-OVIDE line, we obtained synoptic estimates of the large-scale circulation, the MOC magnitude and vertical structure, and the associated heat flux across the line at the time of the cruises. Then, we used these estimates as benchmarks for validating a monthly MOC index built upon a combination of altimetry and Argo float data. The results are discussed with a focus on the seasonal to decadal variability of the MOC since the early 1990s.

The manuscript is organized as follows. The data are described in Section 2. The large-scale circulation features, the MOC and their variability at A25-OVIDE, as estimated from repeat hydrographic measurements, are reported on in Section 3. The monthly time series of the MOC magnitude derived from Argo and altimetry data is presented and discussed in Section 4. Heat flux variability and its relation to the MOC variability are discussed in Section 5. We conclude the paper in Section 6.

2. Data sets

2.1. OVIDE Greenland–Portugal section

The primary data source was composed of (i) the five repeats of the Greenland–Portugal line occupied since 2002 every other year in the framework of OVIDE (Observatoire de la Variabilité Interannuelle à Décadenne) project and (ii) the nearby A25-FOUREX line occupied in 1997 (Fig. 1, Table 1). FOUREX measurements, referred to in the following as FOUREX 1997, have been described by Bacon (1998) and analyzed by Alvarez et al. (2002) and Lherminier et al. (2007). OVIDE 2002, 2004 and 2006 measurements have been presented and analyzed by Lherminier et al. (2007, 2010) and Gourcuff et al. (2011). New data collected along the OVIDE section in 2008 and 2010 complement our data set. Those six sections are referred to as the “OVIDE” sections hereafter.

About ninety hydrographic stations were occupied during each cruise. Prior to 2008, the CTD measurements were carried out using a Neil brown Mark III CTD probe, then a Sea-bird Electronics 911 plus CTD probe was used. Overall, the CTD measurement accuracies are 1 dbar for pressure, 0.002 °C for temperature and 0.003 for salinity (see e.g. Billant et al., 2004 and Branellec et al., 2011). The station spacing was nominally of 25 nautical miles (NM) and was reduced to 16 NM in the Irminger Sea and 12 NM or less over steep topographic features.

Ship-mounted Acoustic Doppler Current Profiler (S-ADCP) measurements were combined with hydrography to estimate the absolute transports across the OVIDE line (see Section 3.1). The S-ADCP measurements were obtained using a RD Instruments 150 kHz ADCP during FOUREX 1997 and a RD Instruments 75 kHz ADCP during OVIDE. A high level of quality was achieved for the S-ADCP velocity data by correcting for the misalignment of the transducer orientation relative to the vessel axis and systematic bias in navigation heading. This was done by using the “bottom track” calibration procedure for FOUREX 1997 (Bacon, 1998) and by minimizing the correlation between the ship velocity and the current component along the ship trajectory during acceleration/deceleration phases for the OVIDE 2002 to OVIDE 2010 (Joyce, 1989; Lherminier et al., 2007, 2010). The estimated error on 1 min averaged mean velocity is of the order of 0.03 m s⁻¹. The velocity due to the barotropic tide was estimated and removed from the ADCP signal as explained by Lherminier et al. (2007).

For each survey, the Ekman transport orthogonal to the section was estimated from wind stress averaged over the months of the cruise. Depending on the availability of the wind stress products at the time of the cruise, different sources of wind stress were used: the European Centre Medium-Range Weather Forecast reanalysis ERA40 (Uppala et al., 2005) for FOUREX 1997 and OVIDE 2002, the CERSAT gridded products (available at http://cersat.ifremer.fr) for OVIDE 2004 OVIDE 2006 and OVIDE 2008, and the monthly products of the NCEP/NCAR reanalysis for OVIDE 2010 (Kalnay et al., 1996). For the computation of the MOC time series presented in Section 4, a homogeneous time series of the Ekman transport was derived from the NCEP/NCAR reanalysis wind stress data.
2.2. Temperature and salinity gridded fields

The monthly In Situ Analysis System (ISAS) provides gridded fields of temperature and salinity over the period 2002–2010 on a square horizontal grid of side equal to $0.5\times \cos(\text{latitude})$, on 152 levels in the upper 2000 m (see http://wwz.ifremer.fr/lpo/SO-Argo/Products/Global-Ocean-T-S). The vertical resolution is 5 m from the surface to 100 m depth, 10 m from 100 m to 800 m depth and 20 m from 800 m to 2000 m depth. The AN_V6-D2CA2S0 version of the ISAS fields was used to compute the sea-surface referenced monthly geostrophic velocities orthogonal to the OVIDE section. The ISAS fields were obtained by optimal interpolation of the large in situ data set provided by the Argo array of profiling floats and complemented by measurements from drifting buoys, CTDs and moorings (Gaillard et al., 2009; Gaillard, 2012). The data set was provided by the Coriolis data center, one of the Argo Global Data Acquisition Centers. In total, the Argo measurements account for at least 90% of the data employed by ISAS before and for more than 95% since 2006. The CTD data collected along the OVIDE line were excluded from the data base before performing the ISAS AN_V6-D2CA2S0 analysis, so that estimates derived from ISAS are fully independent from OVIDE data.

For the 1993–1996 time span, we used temperature and salinity fields constructed from objectively analyzed pentadal anomalies down to a depth of 2000 m and associated mean fields (Levitus et al., 2005; Boyer et al., 2005) available respectively at http://www.nodc.noaa.gov/OC5/DATA_ANALYSIS/anomaly_data.html and http://www.nodc.noaa.gov/OC5/DATA_ANALYSIS/sal_intro.html and released as part of the World Ocean Atlas 2005. These fields are referred to hereafter as the WOA data set.

2.3. AVISO mapped altimetry product

Absolute sea-surface geostrophic velocities orthogonal to the OVIDE section were computed from the absolute dynamic topography, which is the sum of a sea level anomaly (SLA) and a mean dynamic topography (MDT). SLAs were from the multi-mission altimetry products provided by AVISO (gridded delayed time “upd” products) on a $1/3\times \cos(\text{latitude})$ grid, with data every 7 days since October 1992 (http://www.aviso.oceanobs.com/en/data/products.html). We used the 2005 version of the MDT available on a $1/2$ grid (Rio and Hernandez, 2004).

3. Large scale circulation and MOC estimated from OVIDE hydrographic surveys

3.1. Method for the estimation of the circulation

A linear box inverse model was used for the estimation of the absolute geostrophic field orthogonal to the OVIDE hydrographic
section at the time of the cruise. The inverse model is based on the least squares formalism and follows the method of Jackson (1979), also described by Mercier (1986) and Lux et al. (2000). First, geostrophic velocities referenced to selected levels were computed for each pair of stations from CTD data. Then, the unknown velocities at the reference level were estimated by minimizing the weighted sum of (i) the squared departures from a priori values of the reference-level velocities, (ii) the squared residuals of transport constraints derived from the S-ADCP direct velocity measurements, and (iii) the squared residual of an overall mass conservation constraint. The weight of each constraint is inversely proportional to its uncertainty, such that constraints with large uncertainties bring less information than those with small uncertainties. The inverse model configurations have been described in detail by Lherminier et al. (2007) for FOUREX 1997 and OVIDE 2002, by Lherminier et al. (2010) for OVIDE 2004 and by Gourcuff et al. (2011) for OVIDE 2006. Our study builds on the circulation estimates reported by these authors as well as on new estimates obtained from OVIDE 2008 and OVIDE 2010 data. Model configurations used for the latter two cruises are not detailed here since they are similar to those previously reported for FOUREX 1997 and OVIDE 2002–OVIDE 2006. The least-squares formalism provides errors on the circulation and associated quantities such as the MOC, magnitude and the heat flux (see Lherminier et al., 2010).

The inverse model estimates of transports across OVIDE have been validated by favorable comparisons with independent measurements. The East Greenland-Irminger Current (EGIC, see Fig. 1) transports from the inverse model for OVIDE 2004 and OVIDE 2006 have been shown to agree well with those obtained from concurrent and co-located moored current measurements (Daniault et al., 2011a). The EGIC transports from the inverse model for FOUREX 1997 and OVIDE 2002 are likewise consistent with transports obtained by combining the mooring data with altimetry (Daniault et al., 2011b). Gourcuff et al. (2011) have developed a method to estimate absolute transports across a hydrographic section by using altimetry instead of S-ADCP data to constrain the inverse model. They have applied the method to FOUREX 1997 and OVIDE 2002–OVIDE 2006 and have reported on the North Atlantic Current (NAC), EGIC and heat transports. A good agreement has been shown between the results obtained with the S-ADCP data and those obtained with altimetry, although using altimetry resulted in errors 50% larger when using S-ADCP data (Gourcuff et al., 2011).

3.2. Large-scale circulation

The main circulation features intersected by OVIDE (Fig. 1) can be identified from the barotropic streamfunctions, which were computed for the six occupations of the section by accumulating the top-to-bottom integrated transports from Greenland to Portugal (Fig. 2).

In the Irminger Sea, the barotropic streamfunction shows a well-documented cyclonic circulation (Fig. 2, Fratantoni, 2001; Reverdin et al., 2003; Flatau et al., 2003; Lavender et al., 2005; Våge et al., 2011; Sarafanov et al., 2012). The southward limb is represented by the western boundary current observed above the Greenland continental slope (0–250 km in Fig. 2). The northward limb of the cyclonic pattern (from ~250 km to ~700 km in Fig. 2) is composed of two distinct branches associated, from west to east, with the eastern rim of the Irminger Gyre and with the Irminger Current (Figs. 1 and 2; Våge et al., 2011; Sarafanov et al., 2012).

In the Iceland Basin, between the crest of the Reykjanes Ridge and the Eriador Seamount (Figs. 1 and 2), the depth-integrated circulation is cyclonic (Fig. 2, Sarafanov et al., 2012). Its northward limb is the western branch of the NAC. Surface drifters (Fratantoni, 2001), subsurface floats (Bower et al., 2002) and modeling studies (Treguier et al., 2005) have shown that the southwestward flow over the eastern side of the Reykjanes Ridge crosses the Ridge and joins the Irminger Current. A substantial flow of the NAC-derived waters across the Reykjanes Ridge, from the Iceland Basin to the Irminger Sea, has also been inferred from the volume balance north of the OVIDE section (Lherminier et al., 2010) and north of the 59.5°N section (Sarafanov et al., 2012). East of the Eriador Seamount, the barotropic streamfunction is dominated by the NAC, which is composed of a western and an eastern branch (Figs. 1 and 2) with embedded energetic meanders and eddies. By an analysis of the upper ocean transports and hydrographic properties, Lherminier et al. (2010) have identified a westward shift of the western branch of the NAC between OVIDE 2002 and OVIDE 2004. This shift is apparent in the vertically integrated transports (Fig. 2), which also show that at the time of OVIDE 2004-OVIDE 2010 the western branch of the NAC was located in the vicinity of the Eriador Seamount. The position of the NAC eastern branch is more variable. In the West European basin, the depth-integrated circulation is anticyclonic: the NAC feeds a southward flow in the vicinity of the Azores-Biscay Rise (Figs. 1 and 2).

The barotropic streamfunction variability from one cruise to another is large. The western boundary current transport varied between 24 Sv (OVIDE, 2006) and 43 Sv (OVIDE 2008) southward (Fig. 2), which is within the range of variability expected from transports reported in the literature (see, e.g., Sarafanov et al., 2012, their Fig. 10). The weak western boundary current transport, as obtained from the OVIDE 2006 data, has been corroborated by Daniault et al. (2011a,b) and Gourcuff et al. (2011). The transport between the Reykjanes Ridge crest and the minimum in the barotropic streamfunction in the central Iceland Basin varied between 13 Sv (OVIDE 2002, OVIDE 2006) and 18 Sv (OVIDE 2008) southward. The depth-integrated northward transport computed between the minimum in the barotropic streamfunction in the Iceland Basin and the Portugal coast varied between 20 Sv (OVIDE 2010) and 48 Sv (OVIDE 2008). For comparison, Sarafanov et al. (2012) have found 23 Sv for the 2002–2008 mean depth-integrated northward transport in between the central part of the Iceland Basin and the European coast at 59.5°N. The southward recirculation of the NAC in the West European basin across OVIDE was ~10 Sv in agreement with the ~8 Sv reported by Paillet and Mercier (1997, their Fig. 9).

3.3. Meridional overturning circulation

In order to quantify the thermohaline circulation that comprises the northward flow of warm and saline upper-ocean waters and the southward return flow of colder and fresher waters across the section (Marsh et al., 2005; Lherminier et al., 2007), we computed the meridional overturning streamfunction for each repeat of the OVIDE line, by integrating the Greenland to Portugal transports in density space, from the bottom to the sea surface (Fig. 3). This streamfunction is designated as the MOC (e.g., Lherminier et al., 2007); its maximum (referred to hereafter as the magnitude of the MOCr, Table 2) corresponds to the net southward transport in the lower limb of the overturning cell.

The MOCb is a better measure of the thermohaline circulation at the latitudes of OVIDE than the MOC estimated in depth space (MOC) (Marsh et al., 2005; Lherminier et al., 2007; Zhang, 2010). This is because the northward flow of warm waters transported by the NAC and the southward flow of colder and denser waters carried by the EGIC reside at overlapping depths so that they partially cancel each other out in the MOCb. In density space, the NAC and EGIC are respectively ascribed to the warm (light) and cold (dense) limbs of the overturning cell (e.g., Lherminier et al., 2010). As a result, the magnitude of the MOCb is substantially larger than that of the MOC (Table 2) and unlike the latter the MOCb...
The magnitude of the MOC$_r$ at OVIDE varies between 11.4 Sv (OVIDE 2006) and 18.5 Sv (FOUREX 1997) (Table 2). The error on each value of the MOC$_r$ magnitude, based on the uncertainties of the lower limb velocities, is estimated at 2.4 Sv (Lherminier et al., 2010). The standard deviation is 2.5 Sv. Since our cruises were carried out between May–June and August–September (Table 1), this variability is characteristic of summer, when, in models, the intra-annual variability of the MOC$_r$ at OVIDE is at a yearly minimum (Treguier et al., 2006). Models also suggest that because of the difference in the FOUREX 1997 and OVIDE 2002–OVIDE 2010 section paths (Fig. 1), the MOC$_r$ magnitude at FOUREX 1997 might differ from that across the OVIDE section by ~1 Sv on average (Treguier et al., 2006), which is smaller than the typical uncertainty in the MOC$_r$ estimates (~2.4 Sv). The 1997–2010 mean MOC$_r$ magnitude, as estimated from the 6 repeats of OVIDE, is 16.0 ± 1 Sv. The density at which the overturning streamfunction reaches a maximum ($\sigma_{\text{MOC}}$) varied between $\sigma_1 = 32.1$ (OVIDE 2004) and $\sigma_1 = 32.2$ (OVIDE 2002) with a mean value of 32.14 (Fig. 3, Table 2). There is no correlation between $\sigma_{\text{MOC}}$ and the MOC$_r$ magnitude. This result can be understood in the light of Desbruyères et al. (accepted for publication) who have shown that the MOC$_r$ magnitude at OVIDE is correlated with the slope of $\sigma_{\text{MOC}}$ east of the Reykjane Ridge rather than with its intrinsic value.

The magnitude of the MOC$_r$ upper limb is defined as the transport between $\sigma_{\text{MOC}}$ and the sea surface. Its mean value is estimated at 17.0 ± 1 Sv, which is 1.0 Sv larger than the mean magnitude of the MOC$_r$ lower limb (Table 2). This difference comes from a northward transport of 1 ± 3 Sv that was imposed in the inverse model as a constraint on the net transport across OVIDE in order to account for the flow of the upper-ocean waters that does not overturn north of OVIDE but circulates in the surface layer through the Arctic Ocean and the Canadian Archipelago (see Lherminier et al., 2007). After inversion, the net transport across OVIDE and, hence, the difference between the magnitudes of the MOC$_r$ upper and lower limbs varied between 0.3 Sv southward (OVIDE 2004) and 2.2 Sv northward (FOUREX 1997) (Table 2).

4. MOC variability estimated from altimetry and Argo

In the previous section, variability of the MOC$_r$ at OVIDE was assessed from the 6 synoptic surveys. Hereafter, we quantify the MOC$_v$ variability at OVIDE from 1993 to 2010 by combining altimetry with ISAS or WOA temperature and salinity data sets. To do so, we estimated the net northward transport in the MOC$_r$ upper limb using an approach similar to that proposed by Willis (2010). The method is described in Section 4.1, the evaluation of the results is provided in Section 4.2 and the reconstructed MOC$_v$ time series is discussed in Section 4.3.

4.1. Method for the estimation of the MOC$_v$ variability from altimetry and Argo

The monthly ISAS temperature and salinity gridded fields (available from January 2002 to December 2010 on 152 levels) were interpolated at 50 km spaced locations along the OVIDE section and the monthly dynamic heights referred to the sea surface were computed at each point of the horizontal/vertical grid. The weekly SLA and the MDT provided by AVISO were interpolated...
define the monthly magnitudes of the MOC\(_r\) upper limb (see Section 3.3). The time series of these magnitudes (Fig. 5) is referred to in the following as the MOC\(_r\) index. The associated uncertainty was estimated as the standard deviation of an ensemble of 100 MOC\(_r\) estimates obtained by random perturbations of both the ISAS temperature and salinity fields and the altimetry-derived surface dynamic heights, according to their respective error statistics. The standard errors for the ISAS fields were obtained from the objective analysis used for their estimation (Gaillard et al., 2009). These errors depend on the data distribution in time and space, the variance of the field and the noise in the data. The standard error for the surface dynamic height was chosen so as to result in an error of 0.03 m s\(^{-1}\) in surface velocity when calculated over a distance of 100 km, as suggested by Gourcuff et al. (2011) from a comparison of AVISO altimetry products with S-ADCP data. We assumed that errors on the ISAS and altimetry fields were correlated over distances corresponding respectively to the correlation length used for constructing the ISAS fields (250 km) and the correlation distance for the AVISO data at OVIDE (100 km, Gourcuff et al., 2011).

In the absence of Argo measurements and, hence, ISAS fields before 2002, the MOC\(_r\) index was extended back to 1993 (Fig. 6) by combining altimetry data with the monthly mean density fields obtained by averaging the ISAS fields over 2002–2010. When using this approach (see also Willis, 2010), we assumed that most of the upper ocean transport changes at OVIDE are captured by the geostrophic surface velocities measured by the altimetry, at least on seasonal and longer time scales. To test the sensitivity of the MOC\(_r\) index to the selected mean density field, we performed an additional estimate of the MOC\(_r\) index for the 1993–1996 period by replacing the 2002–2010 mean ISAS fields with the pentadal WOA climatology (Fig. 6). The mean difference between the two estimates is 0.4 Sv, such that this substitution had a very minor impact on the results.

### 4.2. Evaluation of the reconstructed fields

The summer mean temperature and velocity fields estimated from ISAS and altimetry (Fig. 7c and d) are qualitatively similar to the mean fields derived from the 5 hydrographic surveys OVIDE 2002–OVIDE 2010 (Fig. 7a and b). The shape of the 32.14 isopycnal (the average \(\sigma_{\text{MOC}}\), see Section 3.3) is likewise correctly reproduced by ISAS (Fig. 7a–d). Closer inspection reveals that there is indeed no considerable bias between the two estimates east of 27\(^\circ\)E (Fig. 7e). Around 30\(^\circ\)W, however, the objective interpolation used for generating the ISAS fields smoothed the bowl associated with the anticyclonic circulation above the Reykjanes Ridge. This resulted in a local difference of 50–100 m in the mean depths of the 32.14 isopycnals between ISAS and repeat hydrography. The effect of this bias on the net transport in the MOC\(_r\) upper limb is minor, as follows from the comparison of the mean transports accumulated from Greenland to Portugal within the MOC\(_r\) upper

### Table 2

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Another indication of the reliability of the MOC$_{\sigma}$ reconstruction comes from the comparison in Fig. 8 of the observed and reconstructed overturning streamfunctions for August–September 1997 (FOUREX 1997) and June–July 2006 (OVIDE 2006), when the MOC$_{\sigma}$ magnitude was at its observed maximum and minimum, respectively (Table 2). The observed and reconstructed streamfunctions are qualitatively similar, and the difference in the magnitudes of the MOC$_{\sigma}$ upper limb between the two synoptic estimates is 7–8 Sv for both estimates (Fig. 8). As no monthly ISAS fields are available for 1997, the MOC$_{\sigma}$ reconstruction for August–September 1997 relies on the monthly mean density fields from ISAS averaged over 2002–2010. This suggests that the MOC$_{\sigma}$ index built for both the 2002–2010 and 1993–2010 time spans is able to reproduce the observed extreme synoptic states of the overturning circulation. The observed and reconstructed streamfunctions averaged over 2002–2010 (Fig. 8) also show a remarkable similarity, and the mean $\sigma_{\text{MOC}}$ in the reconstruction from altimetry and ISAS ($\sigma_1 = 32.14$) matches the estimate based on the full-depth measurements at OVIDE (see Section 3.3). Finally, while being fully independent, the magnitudes of the observed and reconstructed MOC$_{\sigma}$ agree within error bars (Fig. 5).

To conclude, we find a good overall agreement between the reconstructed velocity fields from altimetry and ISAS and the ship-based observations in terms of meridional overturning streamfunction and MOC$_{\sigma}$ magnitude. We will now focus on discussing the MOC$_{\sigma}$ index.

### 4.3. The MOC$_{\sigma}$ index

The 2002–2010 mean magnitude of the reconstructed MOC$_{\sigma}$ (Fig. 5) is $18.1 \pm 1.4$ Sv. The error on the mean 2002–2010 MOC$_{\sigma}$ was estimated as the standard deviation of the MOC$_{\sigma}$ index over this time span divided by $\sqrt{N}$, where $N$ is the number of independent estimates in the time series, to which we added 1 Sv to take into account errors in the MDT not accounted for by the method. Considering that no reliable estimate of the error in the MDT is available (see Gourcuff et al., 2011), this ad hoc procedure was implemented based on the differences $\sim$1 Sv found between MOC estimates based on the MDT used herein and those based on the MDT deduced from the OVIDE inversion-derived surface circulations.

The MOC$_{\sigma}$ index averaged over summer months (May–September) from 2002 to 2008 (16.9 $\pm$ 1.5 Sv) nearly matches the 2002–2008 mean summer (May–September) magnitude of the MOC$_{\sigma}$ upper limb at 59.5°N (16.6 $\pm$ 1.1 Sv) reported by Sarafanov et al. (2012). The estimate by Sarafanov et al. (2012) and the MOC$_{\sigma}$ index presented herein are not fully independent, as both estimates employ the MDT by Rio and Hernandez (2004) and altimetry data. Yet, the hydrographic data sets used in the two estimates are different: the ISAS fields in the present analysis and the full-depth repeat hydrography at 59.5°N in the study by Sarafanov et al. (2012).

The 2002–2010 time series of the reconstructed MOC$_{\sigma}$ suggests that the magnitude of the MOC$_{\sigma}$ upper limb at OVIDE declined from 20.8 $\pm$ 1.7 Sv in 2002–2003 to 15.7 $\pm$ 1.7 Sv in 2006–2007 and has recovered since then. Energetic higher-frequency variability is superimposed on this signal: on intra-seasonal time scales, the MOC$_{\sigma}$ magnitude varied by more than 5 Sv.
The MOC$_r$ index extended back to 1993 using altimetry and the ISAS temperature and salinity fields averaged over 2002–2010 is compared in Fig. 6 with the MOC$_r$ index built upon altimetry and the pentadal WOA analysis (1993–1996) and the 6 synoptic estimates from repeat hydrographic measurements at OVIDE (FOUREX 1997, OVIDE 2002–OVIDE 2010). Despite a slight overestimation of the variability by altimetry alone, the comparison is quite favorable and altimetry alone clearly captures the main features of the variability. This is likely due to the fact that the MOC$_r$ upper limb at OVIDE represents a relatively thin upper layer (see Fig. 7) in which the horizontal velocity anomalies are vertically correlated.

The 1993–2010 MOC$_r$ index exhibits apparent low-frequency variability at a timescale of 7–9 years: the index was higher than 20 Sv in 1993–1996, 2002–2003 and 2009–2010, and lower than 15 Sv in 1998–1999 and 2006–2007 (Fig. 6). The steep decline in the MOC$_r$ index between 1996 and 2000 is consistent with the results from numerical modeling (e.g., Böning et al., 2006; Xu et al., 2013) that suggested a fast response of the MOC magnitude to the NAO-related weakening of the surface forcing. The 1993–2010 linear trend in the MOC$_r$ index estimated at $-2.5 \pm 1.4$ Sv (95% confidence level) suggests a significant decline of the MOC$_r$ at OVIDE since the early 1990s.

The intra-annual variability in the MOC$_r$ index shows a pronounced seasonal cycle (Fig. 9) with maximum in winter, minimum in summer and peak-to-peak amplitude of 4.3 Sv. This seasonality explains why the mean MOC$_r$ magnitude estimated from summer hydrographic measurements at OVIDE (17.0 $\pm$ 1 Sv for the MOC$_r$ upper-limb) is smaller than the annual mean of the MOC$_r$ from the 2002–2010 reconstruction (18.1 $\pm$ 1.4 Sv). The seasonal signal in the MOC$_r$ (Fig. 9) is mostly due to the geostrophic MOC$_r$ seasonal signal, the seasonal Ekman signal being much weaker. Likewise, a study of the seasonal variability of the MOC at 26.5°N from the RAPID data set has revealed that the contributions of the variability of the Gulf Stream and interior geostrophic flow to the MOC seasonal variability were substantially larger than that of the Ekman component (Kanzow et al., 2010).
at constant pressure, \( \theta = \theta(x, z) \) is potential temperature, \( x(x, z) \) is the velocity orthogonal to the section, \( x \) is the along section coordinate and \( z \) is depth. HF estimated from the 6 occupations of the OVIDE line ranges between 0.70 ± 0.5 PW (FOUREX 1997) and 0.29 ± 0.5 PW (OVIDE 2006) (Table 3), with a mean value of 0.51 ± 0.06 PW and a standard deviation of 0.14 PW. This mean HF value is within the HF range reported by Ganachaud and Wunsch (2003) for the 40°N–60°N latitude interval.

Classically, to get insight on the elements of circulation that influence HF variability, \( x(x, z) \) and \( \theta(x, z) \) are decomposed as \( x(x, z) = \theta(x, z) + \phi(x, z) \) and \( \theta(x, z) = \phi(x, z) + \theta'(x, z) \), respectively, where \( \theta(x, z) \) is the section average velocity corresponding to the net transport across the section, \( \phi(x, z) \) and \( \phi(x, z) \) are horizontally averaged components and \( \theta'(x, z) \) and \( \theta'(x, z) \) are anomalies (Böning and Herrmann, 1994; Álvarez et al., 2004). Denoting hereafter the area integration over the OVIDE section by an overbar, HF can then be written as:

\[
HF = \rho c_p \frac{\partial}{\partial z} \frac{\langle u \rangle}{<\theta>} + \rho c_p \langle \nabla \theta \rangle + \rho c_p \langle \tau \rangle <\theta>
\]  

where the right hand side (r.h.s.) term is the sum, from left to right, of an overturning component (HF

\( \text{dia} \)), a horizontal or gyre component (HF

\( \text{gyre} \)) and a component associated with the net transport across the section (HF

\( \text{net} \)). To be consistent with the use of the MOC,

\( \text{rf} \), the HF decomposition was performed herein in potential density \( \sigma_1 \) space rather than by using depth as vertical coordinate. Accordingly, the brackets in Eq. (1) now denote an average at constant density and the r.h.s. of Eq. (1) is the sum of a diapycnal component (HF

\( \text{dia} \)), an isopycnal component (HF

\( \text{iso} \)) and HF

\( \text{net} \) (Table 3).

HF and its variability at OVIDE are largely explained by HF

\( \text{dia} \) (Table 3). Accordingly, HF

\( \text{iso} \) and HF

\( \text{net} \) are small, ranging between -0.07 PW and 0.08 PW (Table 3). To indentify the mechanisms behind HF variability, HF

\( \text{dia} \) was effectively approximated as \( \rho c_p A\emptyset \) (Table 3), where \( A\emptyset \) is the difference between the area averages of the temperature over the upper and lower limbs of the MOC,

\( \rho \) and \( c_p \) are constant and equal to 1027 kg m

\(-3\) and 3989 J kg

\(-1 K\)

\(-1\), respectively. The mean of \( A\emptyset \) is 6.87°C and its standard deviation is 0.22°C. If we impose the OVIDE hydrography-derived 1997–2010 mean MOC,

\( 17.0 \) Sv for the upper limb) as volume transport in the above approximation, a change in \( A\emptyset \) by one standard deviation (0.22°C) would lead to a change of 0.014 PW in HF, which is an order of magnitude less than the standard deviation of HF

\( \text{dia} \) (0.1 PW). The variability of HF

\( \text{dia} \) at OVIDE is thus driven primarily by the variability of the MOC,

\( \rho \) and not by the temperature variability in the water column. In agreement with this inference, HF and MOC,

\( \rho \) are highly correlated (\( r = 0.93 \), \( p < 0.01 \)), with a linear dependence of 0.054 PW/Sv. As expected, the correlation between HF and MOC,

\( \rho \) is smaller and significant at a lower confidence level (\( r = 0.73 \), \( p = 0.1 \)). Accordingly, at high latitudes, the MOC,

\( \rho \) is a quantity better suited than the MOC,

\( \emptyset \) for assessment of HF variability.

6. Summary and discussion

In this study, we analyzed six high-resolution repeats of the OVIDE Greenland–Portugal hydrographic section carried out in spring and summer between 1997 and 2010. For the first time from observations in the subpolar North Atlantic, we quantified the variability of the MOC,

\( \rho \), whose magnitude, defined as the net southward transport in the MOC,

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agreement between the MOC$_{\alpha}$ index and the hydrographic estimates of the MOC$_{\alpha}$ magnitude. The MOC$_{\alpha}$ index was extended back to 1993 using altimetry data in combination with the ISAS monthly density fields averaged over 2002–2010 and the WOA data set for 1993–1996. At interannual to decadal scales, the MOC$_{\alpha}$ variability (Fig. 6) is barely sensitive to the choice of the hydrographic fields used in the analysis. That means that the changes in the velocity shears in the upper water column have a minor effect on the MOC$_{\alpha}$ at OVIDE. The MOC$_{\alpha}$ index averaged over 1993–2010 and 2002–2010 gives 18.7 ± 1.3 Sv and 18.1 ± 1.4 Sv, respectively. Because of the distinct seasonal signal (the MOC$_{\alpha}$ is stronger in winter, see Fig. 9), these mean values are larger than the summer mean magnitude of the MOC$_{\alpha}$ upper limb (17.0 ± 1 Sv) derived from the OVIDE repeat hydrography. The summer (May–September) MOC$_{\alpha}$ index averaged over 2002–2008 (16.9 ± 1.5 Sv) practically matches the recent estimate of the 2002–2008 mean summer MOC$_{\alpha}$ at 59.5°N (16.6 ± 1.1 Sv) (Sarafanov et al., 2012).

At 41°N, Willis (2010) has reported a mean MOC$_{\alpha}$ magnitude of 15.5 ± 2.4 Sv for 2004 through 2006 in agreement with the 15.5 ± 1.8 Sv found for the MOC$_{\alpha}$ index at OVIDE for the same period. This agreement is likely accidental because the vertical coordinates used in the two studies are different, and also because some contribution of the water mass transformation to the MOC magnitude is expected to occur in between 41°N and the Ovide line (e.g., Pickart and Spall, 2007). Furthermore, our estimate suggests a decrease in the MOC$_{\alpha}$ strength of 2.5 ± 1.4 Sv over 1993–2010 at OVIDE, which contrasts with the estimate by Willis (2010) who inferred a strengthening of MOC$_{\alpha}$ by 2.6 Sv since 1993 at 41°N. Yet, our observation agrees with a similar decrease in MOC$_{\alpha}$ magnitude found with the high-resolution Hybrid Coordinate Ocean Model at AR19 (about 46°N) on the same time period (Xu et al., 2013, their Fig. 13b). According to those authors, the drop in MOC$_{\alpha}$ observed between 1996 and 2000 (that is mainly responsible for the trend found in this study) would be part of a decadal variability, and their simulation shows no significant trend for 1978–2011.

From the analysis of the decadal variability in the 4°ocean general circulation model simulation ORCA025-G70, Desbruyères et al. (2013) have shown a decrease in the MOC$_{\alpha}$ magnitude at OVIDE between the mid-1990s and 2004. Using a Lagrangian method, they also identified two vertical overturning cells contributing to the MOC$_{\alpha}$ at OVIDE: a “subtropical cell” connecting low and high latitudes (12 Sv) and a cell internal to the subpolar gyre (4 Sv). The significance of this result is that the comparison of the mean magnitude of the MOC$_{\alpha}$ at OVIDE and at lower latitudes is not straightforward.

The MOC$_{\alpha}$ index exhibits variability from monthly to decadal time scales. On intra-seasonal time scales, the MOC$_{\alpha}$ magnitude variations exceed 5 Sv. The peak-to-peak amplitude of the seasonal cycle of the MOC$_{\alpha}$ is of about 4 Sv, which agrees with the results from an analysis of the 1/6° simulation of the Clipper ocean model (Treguier et al., 2006). A lower-frequency variability of the MOC$_{\alpha}$ at OVIDE in 1993–2010 exhibits a signal at a time scale of 7 to 9 years and an overall decline since 1993. Remarkably, the drop in the MOC$_{\alpha}$ magnitude after the change in the NAO from a positive phase to a negative phase in 1996 is found in most numerical models (Marsh et al., 2005; Treguier et al., 2006; Huck et al., 2008; Desbruyères et al., 2013). Analyzing the decadal variability in the ORCA025-G70 simulation, Desbruyères et al. (2013) have shown that the 10-year low-passed filtered simulated MOC$_{\alpha}$ variability at OVIDE is correlated with the NAO (r = 0.7, significant at the 95% confidence level), the NAO leading the MOC$_{\alpha}$ by 4 to 5 years. Here we find that the winter NAO leads the annual average MOC$_{\alpha}$ index by 2 years (r = 0.45, p = 0.03), which is in line with previous results that have indicated that the circulation in the eastern subpolar gyre responds to the NAO with a 2 year lag (Bersch, 2002).

Hydrographic estimates of HF across OVIDE ranges between 0.29 ± 0.05 PW and 0.70 ± 0.05 PW. HF is more closely correlated with the MOC$_{\alpha}$ (r = 0.93) than with the MOC$_{\alpha}$ (r = 0.73), and, hence, in the northern North Atlantic, variability of the MOC estimated into density space is a better proxy for oceanic heat transport variability. By decomposing the HF in a diapycnal component (HF$_{\text{dia}}$) and an isopycnal component (HF$_{\text{iso}}$), we showed that HF$_{\text{dia}}$, associated with the water mass conversion, explains 87% of the heat flux. By contrast, Álvarez et al. (2002) have reported an equal partition of HF between HF$_{\text{gyre}}$ and HF$_{\text{ov}}$ by decomposing the heat flux across FOUREX 1997 in z coordinates. The mean value of HF at OVIDE, 0.5 ± 0.06 PW, is in general agreement with the regional estimates of HF based on hydrographic data (Koltermann et al., 1999; Ganachaud and Wunsch, 2003) and atmospheric reanalysis (Trenberth and Caron, 2001).

The inferred ~2.5 Sv decrease in the magnitude of the MOC$_{\alpha}$ upper limit at OVIDE since 1993 should be balanced by a similar decrease in the net southward transport in the MOC$_{\alpha}$ lower limb. An essential element of the MOC lower limit at OVIDE being the Deep Western Boundary Current (DWBC, see Fig. 1), one might therefore expect that a decline in the MOC$_{\alpha}$ is accompanied by a decrease in the DWBC transport. Recent observation-based studies (Sarafanov et al., 2009, 2010; Våge et al., 2011) have shown, however, an increase in the DWBC transport between the early 1990s and the late 2000s by about 2 Sv at the location of OVIDE. The estimate by Våge et al. (2011) suggests an increase in the DWBC transport at a rate of 1.5 Sv / 10 years from 1991 to 2007. Over the same time span, the net top-to-bottom transport of the Western Boundary Current (WBC) decreased by about 2 Sv due to a weakening of the circulation above the DWBC (Våge et al., 2011), which agrees with the decline of the large scale cyclonic circulation in the sub-polar gyre of the North Atlantic since the mid-1990s (Hakkinen and Rhines, 2004, 2009). Accordingly, these results suggest that the decrease in the northward transport in the MOC$_{\alpha}$ upper limit at OVIDE in the 1990s–2000s was balanced, at least partially, by a decrease in the southward export of intermediate waters in the western Irminger Sea within the MOC$_{\alpha}$ lower limb. The spatial

| Table 3 |
| Decomposition of HF (PW). HF$_{\text{dia}}$ and HF$_{\text{iso}}$ are the diapycnal and isopycnal components of HF; HF$_{\text{net}}$ is the heat flux component associated with the net volume transport across the section. The estimation error on HF is 0.05 PW. $\Delta T$ is the temperature difference in °C between the upper and lower limbs of the MOC$_{\alpha}$ and $\mu$ is the MOC$_{\alpha}$. $\mu$ MOC$_{\alpha}$ is a linearization of HF$_{\text{dia}}$. The mean values are also indicated.

<table>
<thead>
<tr>
<th>HF (PW)</th>
<th>HF$_{\text{dia}}$ (PW)</th>
<th>HF$_{\text{iso}}$ (PW)</th>
<th>HF$_{\text{net}}$ (PW)</th>
<th>$\Delta T$ (°C)</th>
<th>$\mu$ (PW)</th>
</tr>
</thead>
<tbody>
<tr>
<td>FOUREX 1997</td>
<td>0.70</td>
<td>0.61</td>
<td>0.05</td>
<td>0.04</td>
<td>7.26</td>
</tr>
<tr>
<td>OVIDE 2002</td>
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<td>0.41</td>
<td>0.03</td>
<td>0.01</td>
<td>6.59</td>
</tr>
<tr>
<td>OVIDE 2004</td>
<td>0.51</td>
<td>0.44</td>
<td>0.08</td>
<td>0.01</td>
<td>6.91</td>
</tr>
<tr>
<td>OVIDE 2006</td>
<td>0.39</td>
<td>0.34</td>
<td>0.07</td>
<td>0.02</td>
<td>6.81</td>
</tr>
<tr>
<td>OVIDE 2008</td>
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<td>0.47</td>
<td>0.01</td>
<td>0.03</td>
<td>6.84</td>
</tr>
<tr>
<td>OVIDE 2010</td>
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<td>0.53</td>
<td>0.05</td>
<td>0.02</td>
<td>6.80</td>
</tr>
<tr>
<td>Mean</td>
<td>0.51</td>
<td>0.46</td>
<td>0.03</td>
<td>0.02</td>
<td>6.87</td>
</tr>
</tbody>
</table>
pattern of temporal changes in the MOC lower limb in the 1990s–2000s needs further investigation though, which is beyond the scope of the present study.

A general conclusion from the analysis performed herein is that a monitoring of the MOC in the northern North Atlantic is possible from sustained measurements. The method we presented can be applied for estimation of the MOCr variability at sections, such as OVIDE, where the MOCr upper limb represents a broad northward flow not interacting with bathymetry. In this case, the variability of the upper-ocean northward transport can be efficiently resolved by satellite altimetry and Argo float measurements. The method might not be applicable at latitudes where the northward transport in the MOCr upper limb is concentrated at the western boundary.

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