

North Atlantic ventilation using chlorofluorocarbons and idealised-tracer simulations

By YAN-CHUN HE^{1,2}, YONGQI GAO^{2,3,4*}, MATS BENTSEN³, BENKUI TAN¹ and OLA. M. JOHANNESSEN³, ¹*Department of Atmospheric and Oceanic Sciences, School of Physics, Peking University, Beijing, China;* ²*Nansen-Zhu International Research Centre, Institute of Atmospheric Physics, Chinese Academy of Sciences, Beijing, China;* ³*Nansen Environmental and Remote Sensing Centre, Bergen, Norway;* ⁴*Bjerknes Centre for Climate Research, Bergen, Norway*

(Manuscript received 17 May 2012; in final form 18 September 2012)

ABSTRACT

The simulated chlorofluorocarbon CFC-11 and 29 geographically defined CFC-11 tracers, as well as 29 geographically defined idealised tracers, are used to quantify the regional contribution to the ventilation of the North Atlantic Ocean in a global version of the Miami Isopycnal Coordinate Ocean Model (MICOM) driven by the daily NCEP/NCAR forcing. Age tracers attached to 29 idealised tracers are also used to estimate the timescales for the water masses' transports. Our results show that the simulated overturning circulation matches the available observed data for both intensity and variability, and the simulated distribution of CFC-11 concentration in the subtropical North Atlantic Ocean is in good agreement with the observations, particularly above 800 m in depth. We found that the sandwich-like distribution of CFC-11 concentration in the subtropical North Atlantic in both the observations and simulations is mainly caused by subduction from the western and eastern subpolar North Atlantic, but the contribution of the former (56.0%) is almost four times larger than that of the latter (15.7%). We demonstrated that the ocean dynamics, instead of the source function, determine the annual and inter-annual variability in both dynamically active tracer (such as water temperature and salinity) and passive tracer (such as CFC-11 and idealised tracer) concentrations in the deep North Atlantic. The 'apparent age' distribution shows that the surface water in the western subpolar North Atlantic takes about 12 yr to reach the Nordic Seas and takes 20 yr from the Nordic Seas to the subtropical deep North Atlantic. The transit-time derived by 'optimum time lag' approach shows a 9.3–13.6 yr lag for the signals propagating from the western subpolar North Atlantic to the subtropical North Atlantic, which is generally consistent with that of about 10 yr derived from the 'apparent age'. The study suggests that geographically defined tracers can be used as an efficient tool to investigate the source and spreading pathway of water, and to forecast the spreading and spreading time of environmental accidents such as the radioactive waste in the world's oceans.

Keywords: ocean ventilation, water mass, CFCs, idealised tracers, tracer variability

1. Introduction

The world ocean plays an important role in the Earth climate system because of its large capacity to store and redistribute heat (Levitus et al., 2001; Barnett et al., 2005; Levitus, 2005), and to absorb and therefore regulate atmospheric greenhouse gases, such as CO₂ (Siegenthaler and Sarmiento, 1993; Keeling and Garcia, 2002). Thus, the rate of the ocean's uptake of heat and CO₂ sheds insight

into the role of the oceans in moderating climate change. Ocean ventilation, here understood as the rates, pathways, and timescales for the surface water to enter the ocean interior, is closely associated with water mass formation and is one of the most important processes related to the role of the ocean in climate. Furthermore, ocean ventilation is also a key element to forecast the fate of ocean pollution (Gao and Drange, 2004; Gao et al., 2005, 2009) and to better understand the ocean biogeochemical system (Gnanadesikan, 2004).

Despite the crucial role of ocean ventilation in climate, there is still a limited knowledge of the temporal and spatial distribution of ocean ventilation properties.

*Corresponding author.
email: yongqi.gao@nersc.no

Physical processes that influence ocean ventilation, such as deep convection, vertical/lateral mixing and deep ocean currents, are very difficult to measure directly. Commonly observed and simulated hydrographic parameters (e.g. temperature and salinity) are not the best diagnostics for ocean ventilation since they can offer only limited information on where and when water mass was formed.

Chemical tracers can provide an independent method to understand the processes of ocean ventilation (e.g. Smethie et al., 2000) and provide detailed information on the pathway and rates of water mass renewal (e.g. Rhein et al., 2002; LeBel et al., 2008) when observed data of the chemical tracer are available. However, the temporal and spatial samplings through the large international campaigns so far conducted, such as the World Ocean Circulation Experiments (WOCEs) and the Geochemical Ocean Section Study (GEOSECS), are far from sufficient to fully understand the world's ocean ventilation. Therefore, numerical simulations of chemical tracers have been widely used to complement the study of ocean ventilation (e.g. England, 1995a). Chlorofluorocarbons (CFCs), owing to their well-documented atmospheric histories (Walker et al., 2000), have often been used to estimate ocean ventilation on decadal and inter-decadal timescales (e.g. Dutay et al., 2002; Gao et al., 2003).

However, it is difficult to quantify at any location the contributions from different global water masses due to the nature of the source functions for the chemical tracers (either from the atmosphere or ocean bottom or specific ocean sites). Furthermore, natural and anthropogenic tracers have been used to estimate the ventilation timescales of ocean interior. The 'ages' constructed from tracers, generally, can provide complementary information about ocean ventilation (e.g. Stuiver et al., 1983; Broecker et al., 1991; Smethie, 1993).

In this study, geographically-defined CFC-11 tracers, representing the global surface water masses in different areas of the world ocean, are simulated and used to reveal the transport pathways of ventilated water and quantify the remote contributions to the water mass at a certain location; meanwhile, geographically-defined idealised tracers are also implemented to qualitatively illustrate the impacts of the solubility controlled source function (e.g. global CFC-11) and the constant idealised source function (29 idealised tracers) on the tracer ventilation. The 'apparent age' of a given water parcel, defined as the elapsed time since the water left the source region, was also constructed for idealised tracers to estimate water mass transport timescales (Gao et al., 2005).

The surface world ocean is artificially and geographically divided into 29 areas in our simulation, and this paper focuses on the North Atlantic ventilation. We also investigate the roles of the source function and ocean dynamics in

the variabilities of hydrographic data and passive tracers, and assess on what timescales the anomalies propagate in the deep North Atlantic. The paper is organised as follows: Section 2 provides the model description and experimental setup. Quantification of CFC-11 and idealised tracer ventilation in the North Atlantic are presented in Section 3 and the variabilities and propagation of ocean signals derived from different hydrographic and passive tracer data are analysed in Section 4. Finally, the discussions and conclusions are presented in Sections 5 and 6.

2. Model description and tracer setup

The global version of MICOM (Bleck et al., 1992) is used in this study. MICOM is a primitive equation model using potential density with reference pressure at 2000 db as the vertical coordinate (σ_2 -coordinates). The σ_2 surface is neutral and quasi material for fluid parcels (McDougall and Jackett, 2005), and is favourable to reliable identification of intermediate and deep water. The 34 isopycnic layers from $\sigma_2 = 30.119$ to $\sigma_2 = 37.800$, with values chosen to represent the range of global water masses, comprise the ocean interior; while a bulk-parameterisation for the dissipation of turbulent kinetic energy based on Gaspar (1988) has been incorporated in the uppermost vertically homogeneous mixed layer (ML) where the density varies temporally and spatially. Temperature, salinity and layer thickness are prognostic variables in the ML, whereas in the isopycnic layers below, temperature and layer thickness are the prognostic variables, and salinity is diagnostically determined using the simplified equation of state by Friedrich and Levitus (1972).

The model employed a local orthogonal grid system with the North Pole over central Siberia and the South Pole over Antarctica (see Furevik et al., 2003, Fig. 1b), and it has a coarse resolution with approximately a $1.1^\circ \times 2.4^\circ$ grid by latitude and longitude. The sub-scale eddy-induced tracer transport is parameterised by isopycnal and diapycnal diffusion in the model. The diffusive velocities (diffusivities divided by the size of the grid cell) are set as 0.0035 m/s for tracer dispersion (i.e. for temperature, salinity, and the CFCs) and 0.01 m/s for thickness diffusion and momentum dissipation. The diapycnal mixing coefficient K_d (m^2/s) is parameterised as $K_d = 1.8 \times 10^{-7}/N$, where N (s^{-1}) is the locally stratification-dependent *Brunt-Väisälä* frequency.

The model was initialised from a state of rest with the January salinity and temperature developed by Steele et al. (2001), and forced with daily NCEP/NCAR atmospheric reanalysis fields (Kalnay et al., 1996). The spin-up-phase consisted of three full NCEP/NCAR reanalysis cycles from 1948 to 2007 with each cycle starting from the end of the previous cycle, yielding a total of 177 yr of spin-up. During the spin-up-phase, the ML salinity was relaxed towards the

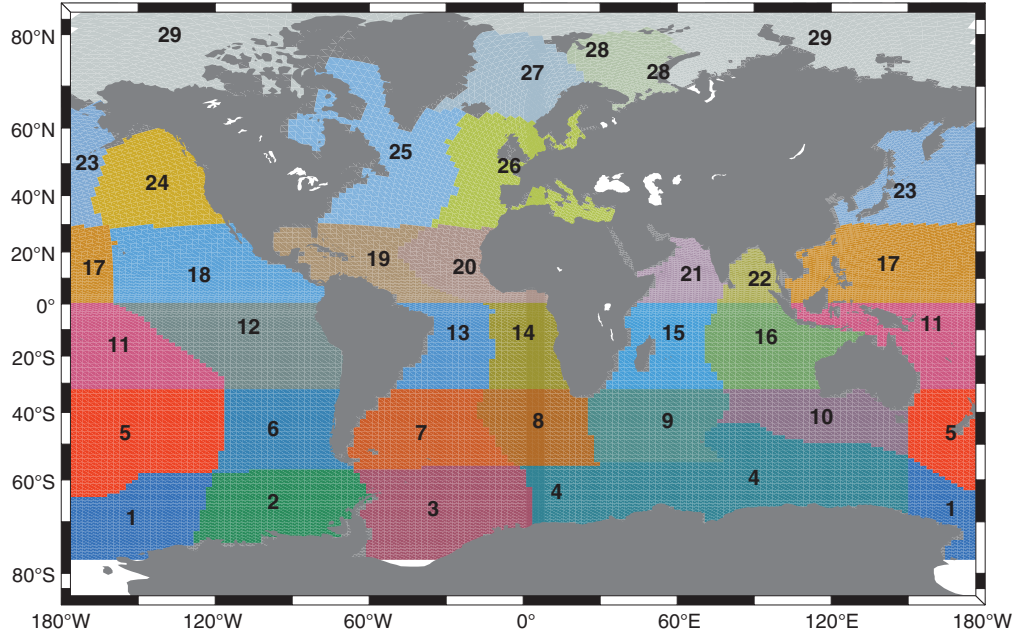


Fig. 1. Locations of 29 geographically defined CFC-11 tracers and 29 geographically defined idealised tracers.

monthly mean climatological values of Levitus et al. (1994). The piston velocity $k(\text{m/s})$ for the salinity relaxation was parameterised as $k = \bar{h}/\tau$, where $\bar{h}(\text{m})$ was the typical mixed layer depth (MLD) and $\tau(\text{s})$ was the relaxation timescale. $\bar{h}$ was set as 50 m when the simulated MLD exceeded 50 m otherwise set as the actual simulated MLD, and τ was set as 30 days. The long-term mean of diagnosed fresh water fluxes, obtained from the 177 yr of spin-up, was applied to the fourth cycle that was also driven by the daily NCAR/NCEP forcing fields. The timescale for the salinity relaxation (τ) was set as 300 days, yielding a weaker relaxation compared with the spin-up phase. The diagnosed time series of Atlantic Meridional Overturning Circulation (AMOC) of these four cycles are similar in terms of the trend, the decadal and inter-annual variabilities (figure not shown). The tracer (CFC-11, CFC-12 and idealised tracers) simulation started from the fifth cycle of the physical model.

To assess the regional contribution to the ventilation of the ocean interior, the surface of the world's oceans was divided into 29 patches (Fig. 1), and each patch was

represented by an independent CFC-11 tracer and an independent idealised tracer. In total, there are 29 regional CFC-11 tracers and 29 idealised tracers, all of which are passive tracers. The division of the ocean surface (Indian, Atlantic and Pacific) meridionally follows their mid-ocean ridges. The equator, 30°N and 30°S are zonal and artificial borders for setting up tracer areas. From 30°N poleward, the Pacific Ocean is divided into two tracer areas with the Bering Strait as the northern boundary. Similarly, in the North Atlantic Ocean, the Greenland–Iceland–Scotland ridge is chosen as the northern boundary. Further north, the Nordic Seas, the Barents Sea and the Arctic Ocean are selected as three different tracer areas for a better tracing of the potential dense water formation in these areas. South of 30°S , the surface ocean is divided into 10 areas with the Weddell Sea and Ross Sea as two independent tracer release areas. Tracers of the first concern (both CFCs and idealised tracers) in this paper are also listed in Table 1.

The simulation of the multiple CFC-11 and idealised tracers started from 1948 with zero concentration in the

Table 1. Location of tracers in the Atlantic Ocean

Tracer 13	Tracer 14	Tracer 19	Tracer 20
Southwest subtropical Atlantic (STA)	Southeast subtropical Atlantic (STA)	Northwest subtropical Atlantic (STA)	Northeast subtropical Atlantic (STA)
Tracer 25	Tracer 26	Tracer 27	Tracer 28
Western subpolar North Atlantic (SPNA)	Eastern subpolar North Atlantic (SPNA)	Nordic Seas	Barents Sea

ocean. We prescribed the source functions for the 29 idealised tracers with a constant and continuous flux. The flux of CFC-11 at the air-sea surface is expressed as: $F = K_w(C_{\text{sat}} - C_{\text{surf}})$, where F (mol/m²/s) is the air-sea flux; K_w (m/s) stands for the gas transfer (or piston) velocity; C_{sat} (mol/m³) is the saturated CFC-11 concentration in moist air near the sea surface; C_{surf} (mol/m³) represents the modelled surface ocean CFC-11 concentration. The parameterisation and implementation of the flux follow the protocol of the Ocean Carbon-cycle Model Intercomparison Project Phase II (OCMIP-2), described in detail by Dutay et al. (2002). The modelled surface ocean CFC-11 concentration, instead of each regional CFC-11 component, was used to calculate the air-sea CFC-11 fluxes. The CFC-11 flux was tagged as individual tracer component according to the regions described above, and each of the tracer components took part in the advective–diffusive process independently in the ocean.

Furthermore, Eulerian age tracers were tagged to the idealised tracers in order to map the timescales for the ventilation. The age was set to zero at the surface in the release area for each idealised tracer. The concept of tracer age has been developed by Thiele and Sarmiento (1990), England (1995b) and Deleersnijder et al. (2001) and implemented in the Ocean General Circulation Model (OGCM) by Gao et al. (2005) and Orre et al. (2008). This tracer age represents concentration-weighted average time for source waters and is dependent on an integrated effect of advection and diffusion processes. The age tracer in this work is similar to the mean value of early developed ocean transit-time distributions (TTDs) or Impulse Boundary Propagator (IBP) (Holzer and Hall, 2000; Haine and Hall, 2002) which usually, however, requires either a prescribed ocean surface tracer concentrations or saturation and a steady-state ocean circulation (Waugh et al., 2004; Peacock and Maltrud, 2006). However, the simulated tracer fields by the TTDs approach can hardly be compared directly with the observed data since it is difficult to maintain both a transient air–sea tracer flux and a transient ocean circulation state (Haine and Hall, 2002).

Therefore, a total of 89 tracers (two global CFC-11 and -12, 29 regional CFC-11 tracer and 29 idealised tracers plus 29 age tracers tagged with idealised tracers) were included in our simulation. These multiple CFC-11 tracers and idealised tracers with different source functions (transient versus constant) can provide unique information about the ocean ventilation including water mass composition and transport pathways. The age tracer, which provides information about water mass transport timescales, is directly derived from the tracer concentration, and the design of the age tracer does not need the assumptions as required by the TTDs approach. These combined multiple tracers and age tracers are expected to provide a joint water

mass composition and transport timescales on the regions of interest.

3. Ventilation in the North Atlantic

3.1. Meridional water mass transport

The northward flowing surface and near-surface water and the southward flowing dense water in the North Atlantic form the so-called Atlantic Meridional Overturning Circulation (AMOC). To illustrate the simulated AMOC, the mean (1948–2009) volume transport for each isopycnal layer across 24.5°N, and the temporal evolution of simulated southward volume transport and available continuous measurement of AMOC are shown in Fig. 2.

In the subtropical North Atlantic, the major part of the water masses with $\sigma_2 < 36.712$ flows northward, while the water mass with $\sigma_2 > 36.846$ flows southward. The transport of water mass with potential density of 36.946 dominates the southward transports, accounting for 41% of the total southward transport. The simulated net transport in the North Atlantic is about 1.42 Sv which is an indication of the inflow from the Pacific to the Arctic Ocean through the Bering Strait and is generally consistent with observational estimates of 1.45 ± 2.5 Sv by Ganachaud (2003) and Roach et al. (1995). The southward transport of the dense water mass (with potential density greater than 36.712 in the model) in the North Atlantic (shown as Fig. 2a) is known as the North Atlantic Deep Water (NADW). The model did not capture the northward flowing dense Antarctic Bottom Water (AABW) at 24.5°N. Therefore, no northward flow appears in layers denser than 36.846 in the simulation.

The temporal variation of the simulated southward volume transport in the whole water column shows an inter-annual to decadal variability with the monthly values ranging from 11.2 to 24.3 Sv (Fig. 2b). The simulated overturning strength is in agreement with the five repeated cruise surveys (1957, 1981, 1992, 1998 and 2004) along 24.5°N (Bryden et al., 2005) except for the years 1957 and 2004. Please note that the observed value in 1957 is possibly due to the large high frequency variability of the observed transport as indicated by Cunningham et al. (2007).

From year 2004 to 2009, there are daily continuous measurement of meridional transport available (Cunningham et al., 2007) shown as the monthly mean in Fig. 2c. The simulated southward volume transport for 2004 to 2009 in the whole water column changes from 14.2 to 21.5 Sv, which is within the range from 8.8 to 26.6 Sv of the monthly mean value of the continuous measurement. The mean value of the simulated overturning transport (18.2 Sv) between 2004 and 2009 is also in the range of 16 ± 2 Sv by Ganachaud (2003) based on WOCE datasets

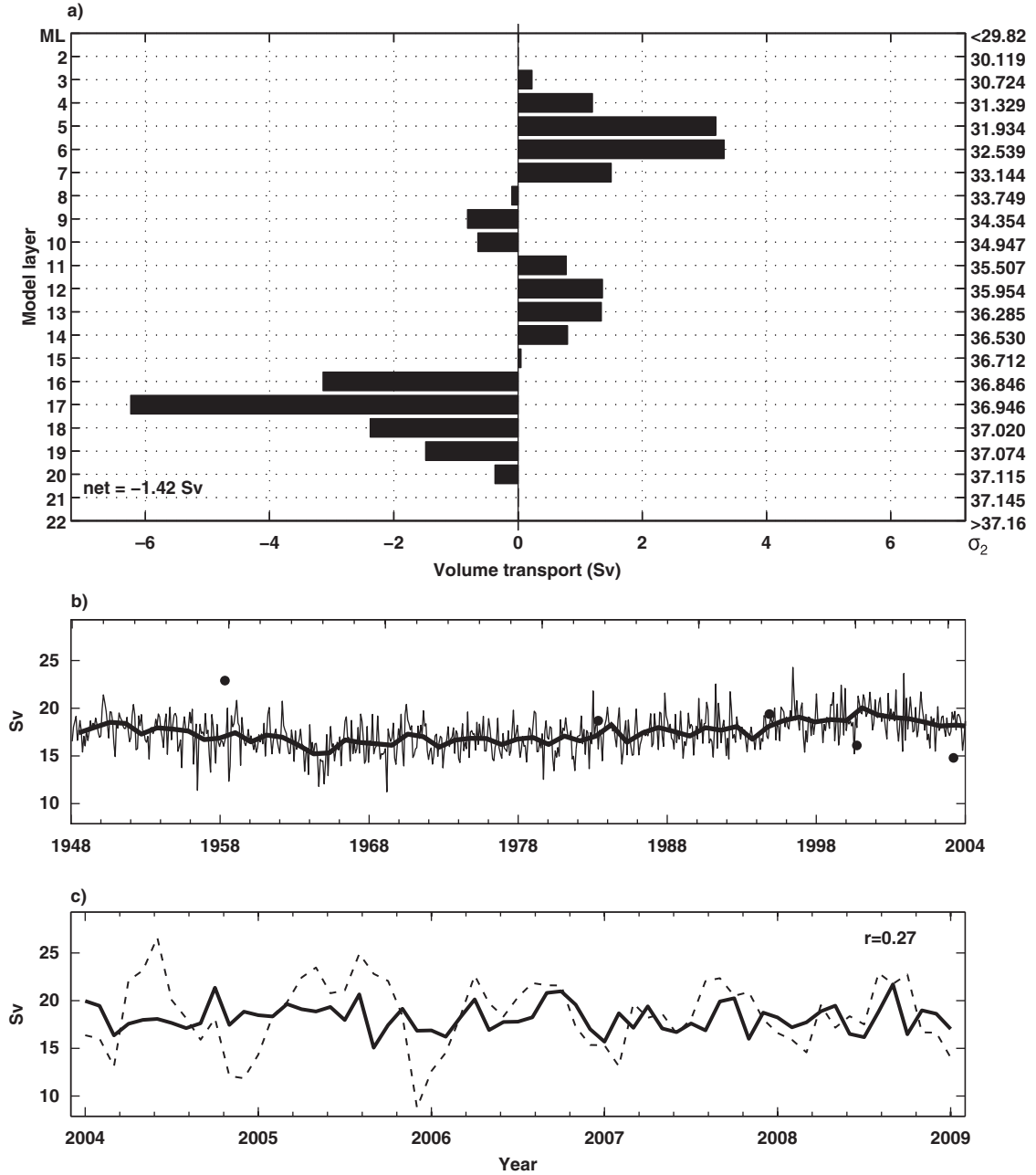


Fig. 2. (a) Long-term mean of simulated volume transport across 24.5°N in the North Atlantic (positive northward). (b) Time series of simulated monthly (thin line) and annually (thick line) mean southward volume transport in the whole water column at 24.5° and available observations (black dots, Bryden et al., 2005) from 1948 to 2004. (c) Time series of monthly mean simulated (solid line) southward volume transport in the whole water column at 26.5°N and monthly mean of continuous measurements of MOC at 26.5°N (dotted line, Cunningham et al., 2007) from 2004 to 2009.

and 18.7 ± 5.6 Sv by Cunningham et al. (2007) based on continuous measurements between 2004 and 2009. The simulated and observed monthly meridional transports have a similar seasonal to inter-annual variability with a correlation factor of 0.27 with statistical significance above the 95% confidence level.

3.2. Observed and simulated tracers in the subtropical North Atlantic

In order to shed light on ocean ventilation in the North Atlantic, observed and simulated CFC-11 in February 1998 along 24.5°N are shown in Figs. 3 and 4. Generally, the

observed subsurface CFC-11 concentrations (below 1000 m) in the western basin is higher than that in the eastern basin where the AABW and the Iceland–Scotland Overflow Water (ISOW) with low CFC-11 concentrations below 0.01 pmol/kg are presented (Smethie Jr et al., 2007; LeBel et al., 2008). The observed CFC-11 exhibits basin-wide surface or near-surface high values in the upper 0–800 m and two localised subsurface maxima around 1800 and 3700 m in the western basin, representing the newly ventilated classic Labrador Sea Water (cLSW) and Overflow Water (OW) (Smethie et al., 2000).

As an interface for the air–sea exchange, the surface water is the earliest to be saturated with dissolvable air gasses, such as CFC-11. However, the observed concentrations of the CFC-11 do not monotonically decrease with depth in the upper 800 m. Instead, a slightly higher CFC concentration of 2.25 pmol/kg lies at 250 m depth across the North Atlantic basin as is shown in the upper right panel of Fig. 3 as well in the vertical profile of CFC-11 concentrations in the upper left panel. This sandwich-like distribution of CFC-11 can also be noticed in Fig. 3 plate 1 of Smethie et al. (2000) and Fig. 2 of LeBel et al. (2008), although it was not highlighted by these

studies. Further examination of the repeated cruise surveys of this section (WOCE A05 section, figure not shown) reveals that this sandwich-like distribution of CFC-11 along 24.5°N in the North Atlantic in the upper 800 m persists during the conducted surveys which cover winter, early spring, and summer (see <http://cchdo.ucsd.edu/groups?id=a05> for detailed spatial and temporal coverage). It is therefore likely that the sandwich-like distribution is not produced by direct vertical mixing or local subduction process induced by basin-scale subtropical gyre (e.g. Qiu and Huang, 1995; Blanke et al., 2002). Hydrographic analysis of this section indicates that the western part of this sandwich-like layer coincides with a weakly stratified water mass with vertically uniform salinity, potential vorticity, oxygen, and temperature of 18°C (figures not shown). This weakly stratified water over a thickness of 200 m is widely recognised as the North Atlantic Subtropical Mode Water (NASTMW), or the so-called Eighteen Degree Water (EDW; Thomson, 1877) dominated by water mass with temperature close to 18°C. The EDW is believed to be important to memorise the climate variability (e.g. Kwon, 2004; Old and Haines, 2006) and substantially contribute to the inter-annual variability

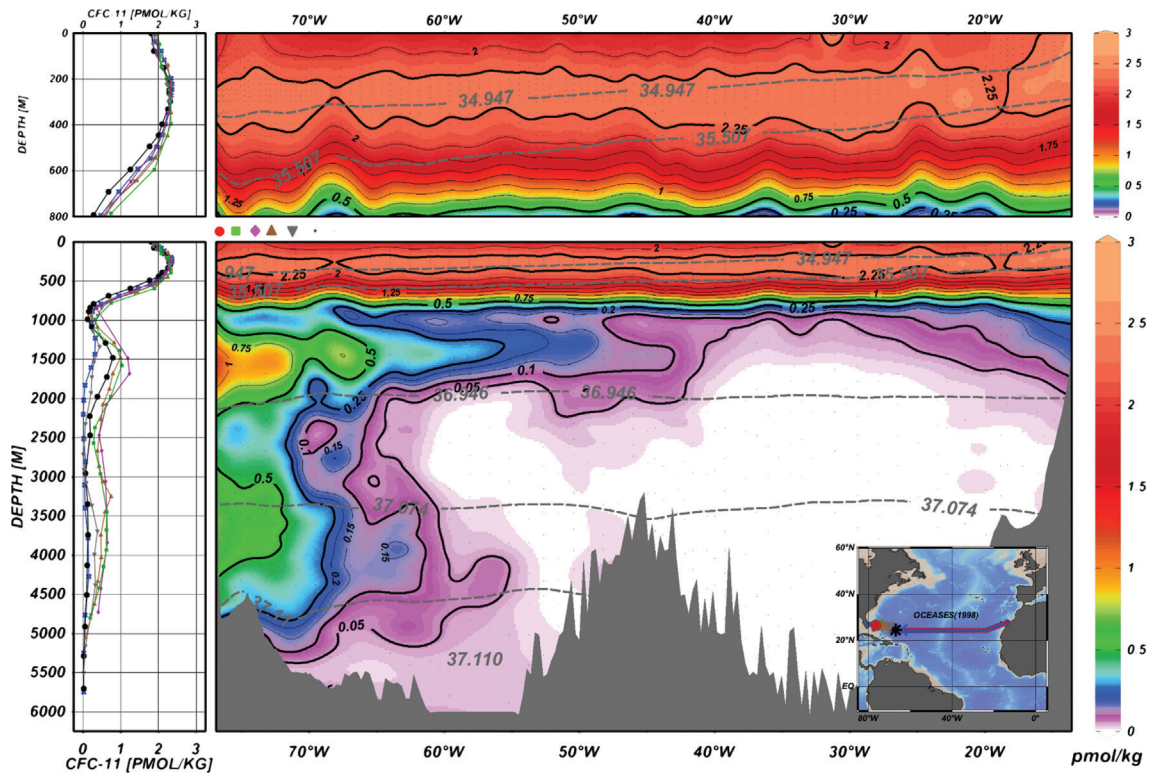


Fig. 3. Observed CFC-11 (pmol/kg) at 24.5°N across the North Atlantic in February, 1998 (LeBel et al., 2008). Upper layers from surface to 800 m are zoomed in the upper panel, and profiles of CFC-11 concentration of selected stations close to the western boundary (denoted by different symbols at the top of axis. e.g. red dot, green square, etc.) are also shown in the left panel. The dashed grey lines denote the locations of 34.947, 35.507, 36.946 and 37.115 isopycnals.

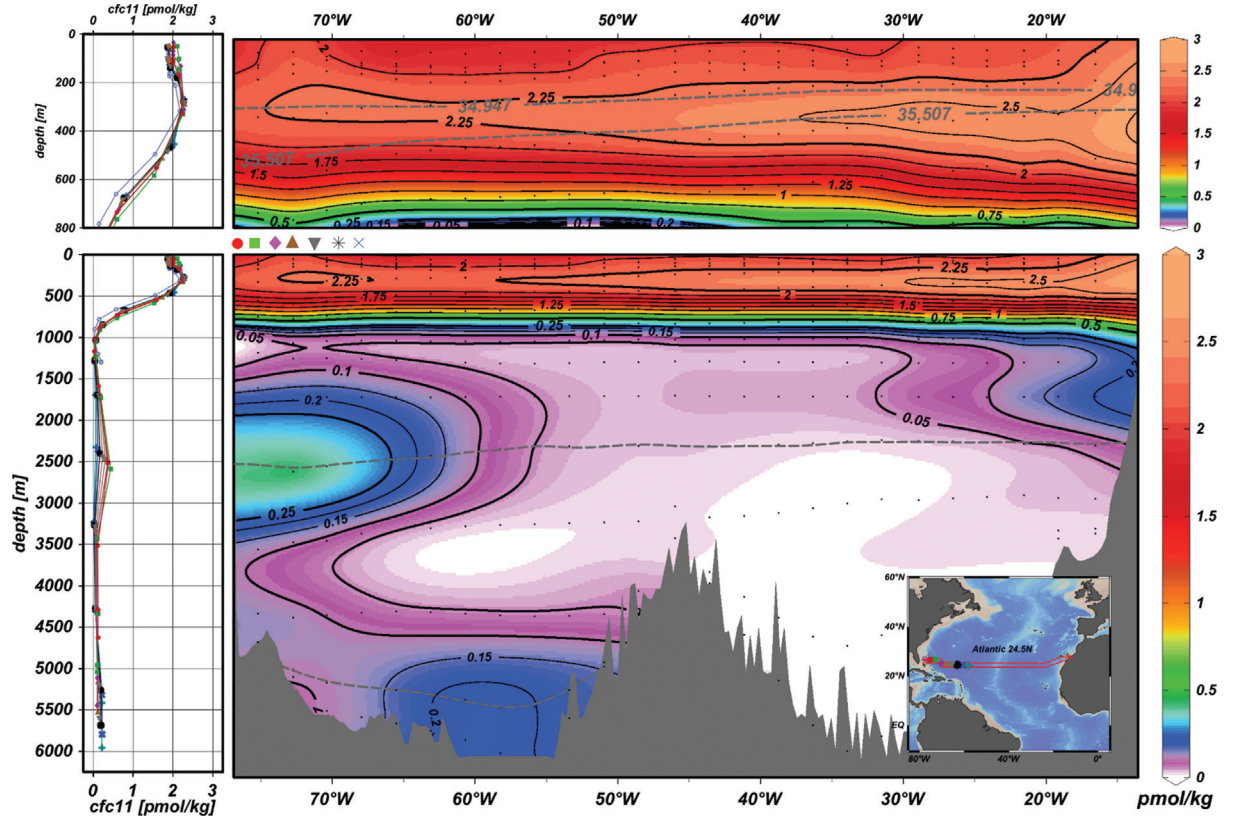


Fig. 4. Same as Fig. 3 but for simulated CFC-11 (pmol/kg) in February, 1998.

in oceanic CO₂ uptake (Bates et al., 2002) and primary production (Palter et al., 2005). The EDW is one of the most studied mode waters regarding its properties, sources and variabilities (e.g. Worthington, 1958; Talley and Raymer, 1982). However, the structure and sources for the EDW related sandwich-like CFC-11 layer, to the best of our knowledge, have not been well documented.

Below around 1000 m depth, the relatively CFC poor upper Labrador Sea Water (uLSW) (Smethie et al., 2000; LeBel et al., 2008) separates the CFC-saturated surface water and the mid-depth classic Labrador Sea Water (cLSW) labelled with higher CFC-11 concentration at about 1800 m depth (Smethie Jr et al., 2007). At 1000 to 2000 m depth close to the eastern boundary, there is also a slightly elevated CFC-11 signal, which is attributed to the Mediterranean Outflow Water (MOW) (LeBel et al., 2008).

The simulated global CFC-11 distribution, which is essentially identical to the sum of the 29 regional CFC-11 simulations, is generally in line with observations, particularly in the range of 0–3000 m depth (Fig. 4). For example, the model well reproduced the sandwich-like CFC-11 distributions in the upper 800 m and captured the local CFC-11 core at 2500 m following the western boundary,

which is 700 m deeper than the core in the observations. In addition, a CFC-11 core at the bottom of the western basin, representing the overflow in the simulation (please refer to Fig. 7), is also present. The model also captured the relatively high CFC-11 signal between 1000 and 2000 m depth in the eastern boundary of the North Atlantic. In addition, the observed structure of thermocline above 1000 m depth is also well captured by the model, e.g. the position and structure of the surface hydrography, the NASTMW and the permanent thermocline (figures not shown).

The distribution of the total concentrations of the 29 independent idealised tracers is shown in Fig. 5. There is no sandwich-like distribution of tracers in the upper 800 m as shown in the observed and simulated CFC-11 (Figs. 3 and 4). Instead, it decreases monotonically with depth. Though the flux for the idealised tracers is different from the spatially and temporally varied CFC-11 flux, the distribution of accumulated tracer concentrations below the upper layer displays some similarities with the simulated CFC-11, e.g. the deep cores at depth of 2500 m and at the bottom of the western basin; relatively higher concentrations between 1000 and 2000 m close to the eastern boundary. If the locally released tracers in the

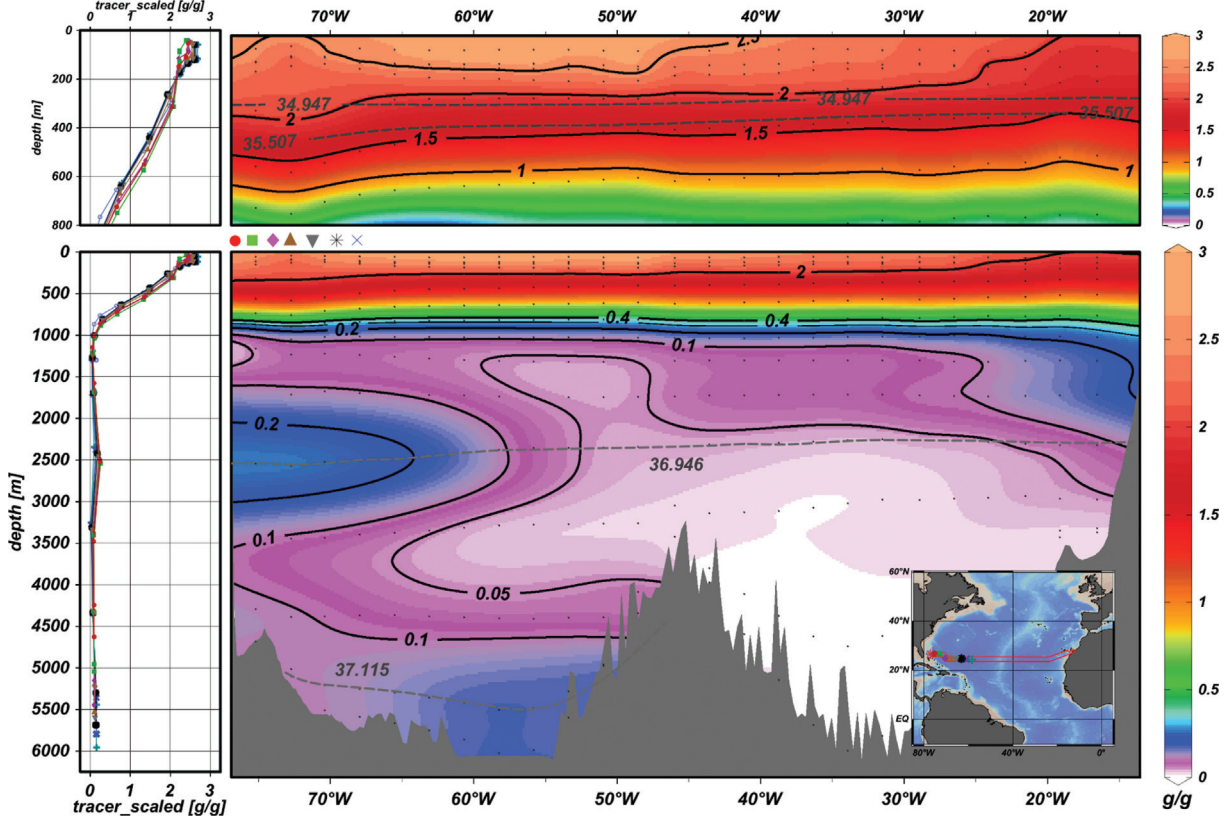


Fig. 5. Same as Fig. 3 but for accumulated concentration (scaled to 0–3 arbitrary unit, herein after) of all 29 idealised tracers at the end of 60-yr simulation.

subtropical North Atlantic (tracers 19 and 20) are excluded and only tracers in the subpolar North Atlantic and Nordic Seas (tracers 25, 26, 27 and 28) are included, the subsurface tracer maximum around 250 m extending horizontally across ocean basin can be successfully captured (figures not shown here), indicating the observed CFC-11 maximum at 250 m is likely not locally generated.

To diagnose the water mass composition and assess the relative contributions from different sources to the ventilated water mass in the subtropical North Atlantic, the accumulated masses of each regional CFC-11 tracer along 24.5°N were firstly calculated. It turns out eight ocean regions represented by tracers 13, 14, 19, 20, 25, 26, 27 and 28 (see Table 1 for detail) were among the top contributors to the total CFC-11 mass, implying that the surface waters in those areas are important for the CFC-11 ventilation in the subtropical North Atlantic. Therefore the contributions from the 8 out of the 29 regions to the ventilation of the North Atlantic Ocean are investigated. To isolate the source water for the CFC-11 ventilation in the North Atlantic, the distributions of CFC-11 concentrations caused by individual geographic CFC-11 area are shown in Fig. 6. The contributions by individual geographic CFC-11 area

to the total CFC-11 and idealised tracer inventory at 24.5°N in the North Atlantic in February 1998 are summarised in Table 2.

As indicated in Fig. 6, the efficient contributors to the CFC-11 ventilation of 0–800 m are the surface waters from the subtropical North Atlantic (tracers 19, 20) and the western (tracer 25) and eastern (tracer 26) subpolar North Atlantic, and to a lesser extent, from the subtropical South Atlantic (tracers 13, 14). However, the concentrations for tracers 13, 14, 19 and 20 decrease with depth quickly whereas the concentrations for tracers 25 and 26 exhibit high values at around 250 m depth originating from the eastern basin and spreading to the western basin, implying that the high observed and simulated CFC-11 concentrations at 250 m is mainly from the western and the eastern subpolar North Atlantic. The surface waters from the western subpolar North Atlantic (tracer 25), the eastern subpolar North Atlantic (tracer 26) and the Nordic Seas (tracer 27) primarily contribute to the CFC-11 ventilation of deep water below 2000 m in the western basin of the North Atlantic, including an intermediate CFC-11 maximum at 2500 m depth and CFC-11 maximum at the bottom.

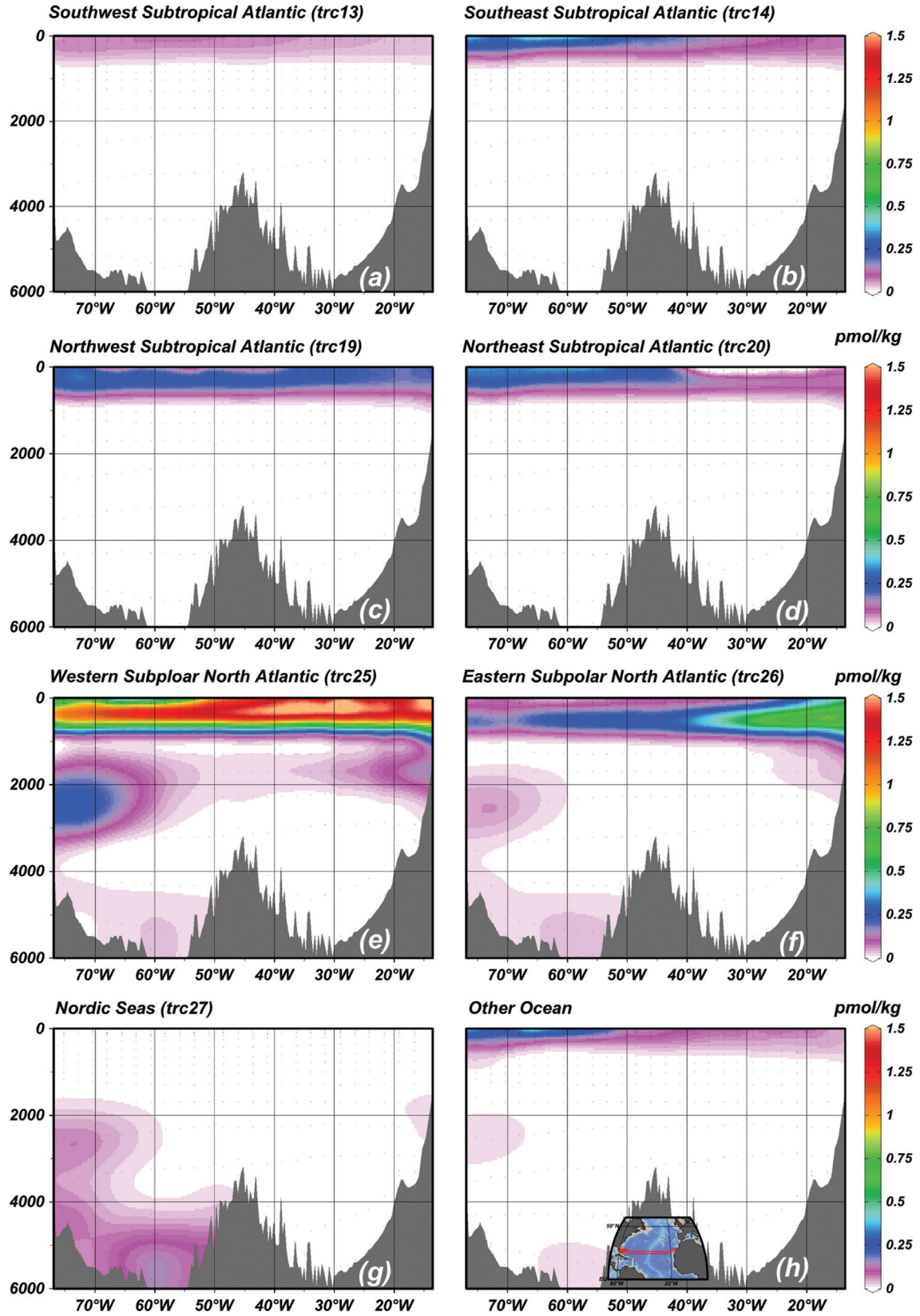


Fig. 6. Concentration of simulated CFC-11 (pmol/kg) at 24.5°N in the North Atlantic from different source regions in February, 1998.

Table 2. Composition for water-mass (a) between layer 0–800 m, (b) at 2500 m of west boundary and (c) at 4500 m depth at sea floor in North Atlantic 24.5°N in February 1998 from different CFC-11 and idealised tracer sources

Depth		Western SPNA (25), (%)	Eastern SPNA (26), (%)	Northwest STA (19), (%)	Northeast STA (20), (%)	Southeast STA (14), (%)	Nordic Seas (27), (%)	Arctic Ocean (29), (%)	Barents Seas (28), (%)	Other Ocean, (%)
0–800 m	(CFC)	56.0	15.7	9.3	7.3	5.6	/	/	/	6.0
	(TRC)	11.4	12.4	21.8	20.3	11.1	/	/	/	23.1
2500 m	(CFC)	64.1	13.5	/	/	/	16.4	2.6	/	3.5
	(TRC)	18.5	7.7	/	/	/	23.5	18.4	/	31.9
4500 m	(CFC)	13.8	16.1	/	/	/	61.0	/	4.0	5.1
	(TRC)	6.8	9.0	/	/	/	44.0	/	18.3	21.8

To quantify the water mass composition for a certain water mass of interest, both the contribution percentages of CFC-11 tracer and idealised tracer mass for three water masses at 24.5°N in the North Atlantic are summarised in Table 2, including water mass between 0–800 m, intermediate tracer-enriched core water mass at around 2500 m and deep tracer-enriched core water mass at the bottom. The fundamental difference of the water mass composition in percentage between CFC-11 and idealised tracer lies in the fact that the air-sea CFC-11 flux is controlled by the CFC-11 solubility (mainly temperature dependent) in sea water whereas the flux of the idealised tracer is prescribed with infinite solubility. Therefore, water mass composition of CFC-11 indicates the combined effects of temperature and advective–diffusive ocean dynamics, while the water mass composition of idealised tracers with a prescribed constant air-sea flux is determined only by the ocean dynamics.

The composition analysis shows that the surface water from the western and eastern subpolar North Atlantic account for 56.0 and 15.7% of the CFC-11 ventilation between the ocean surface and 800 m depth, implying > that the subpolar North Atlantic is a dominant source for the ventilated sandwich-like CFC maximum and other parts of ocean only contribute less than 30%. Idealised tracer derived water mass contributions of different ocean sources, varying from 11.1 to 21.8%, do not exhibit as large contrasts as that derived from CFC-11, and local water mass ventilation from northwest subtropical North Atlantic (tracer 19) and northeast subtropical North Atlantic (tracer 20) have the largest contribution.

Composition of intermediate tracer-enriched core water mass at around 2500 m indicates that the western subpolar North Atlantic (tracer 25) is the dominant source of CFC-11 ventilation (64.1%) while the Nordic Seas and eastern subpolar North Atlantic (tracer 26) are also important sources (16.4 and 13.5%) but only fourth quarter of the contribution from the western subpolar North Atlantic. For idealised water mass composition, sources from

western subpolar North Atlantic and Nordic Seas have an equivalent importance (18.5 and 23.5%). Arctic Ocean (tracer 29) has a larger contribution (18.4%) in idealised tracer ventilation than in CFC-11 ventilation (only 2.6%).

Composition of tracer-enriched water mass deeper than 4500 m shows a similar contribution pattern: the Nordic Seas has very large contributions (61.0 and 44.0% for CFC-11 and idealised tracer respectively); the eastern subpolar North Atlantic, the western subpolar North Atlantic and the Barents Sea contribute below 20% for the CFC-11 ventilation in the deep subtropical North Atlantic.

3.3. Pathways of water masses spreading

The isopycnal model is intrinsically favourable in identifying the sources and pathways of certain water mass since the water mass tends to be transported along isopycnals (Gao et al., 2003). The simulated ocean current and distributions of CFC-11 concentrations along three corresponding isopycnals ($\sigma_2 = 34.947, 36.946, 37.115$) denoted as grey lines in Fig. 5 are therefore shown in Fig. 7.

Note that the sandwich-like CFC-11 distributions in both the observations and simulation (Figs. 3 and 4) are confined to 34.947- and 35.507-isopycnals, which are also within the range of the simulated EDW shown as red line of 18°C contour in Fig. 7b. The 34.947-isopycnal outcrops at sites from Grande Banks around 44°N (Fig. 7b). On this isopycnal, a clockwise current system is dominant. A tracer-enriched tongue spreads from the outcropping areas to the eastern North Atlantic and finally to the southwest following the current with the depth of the isopycnal tilting from 150–300 m. The southwestward spreading of the tracer can partly explain the eastern part of subsurface CFC-11 and idealised tracer maxima at depth of ~250 m shown in Figs. 4 and 5. The distributions of CFC-11 and idealised tracer concentrations on 35.507-isopycnal contribute more to the eastern part of subsurface CFC-11 and idealised tracer maxima (figure not shown

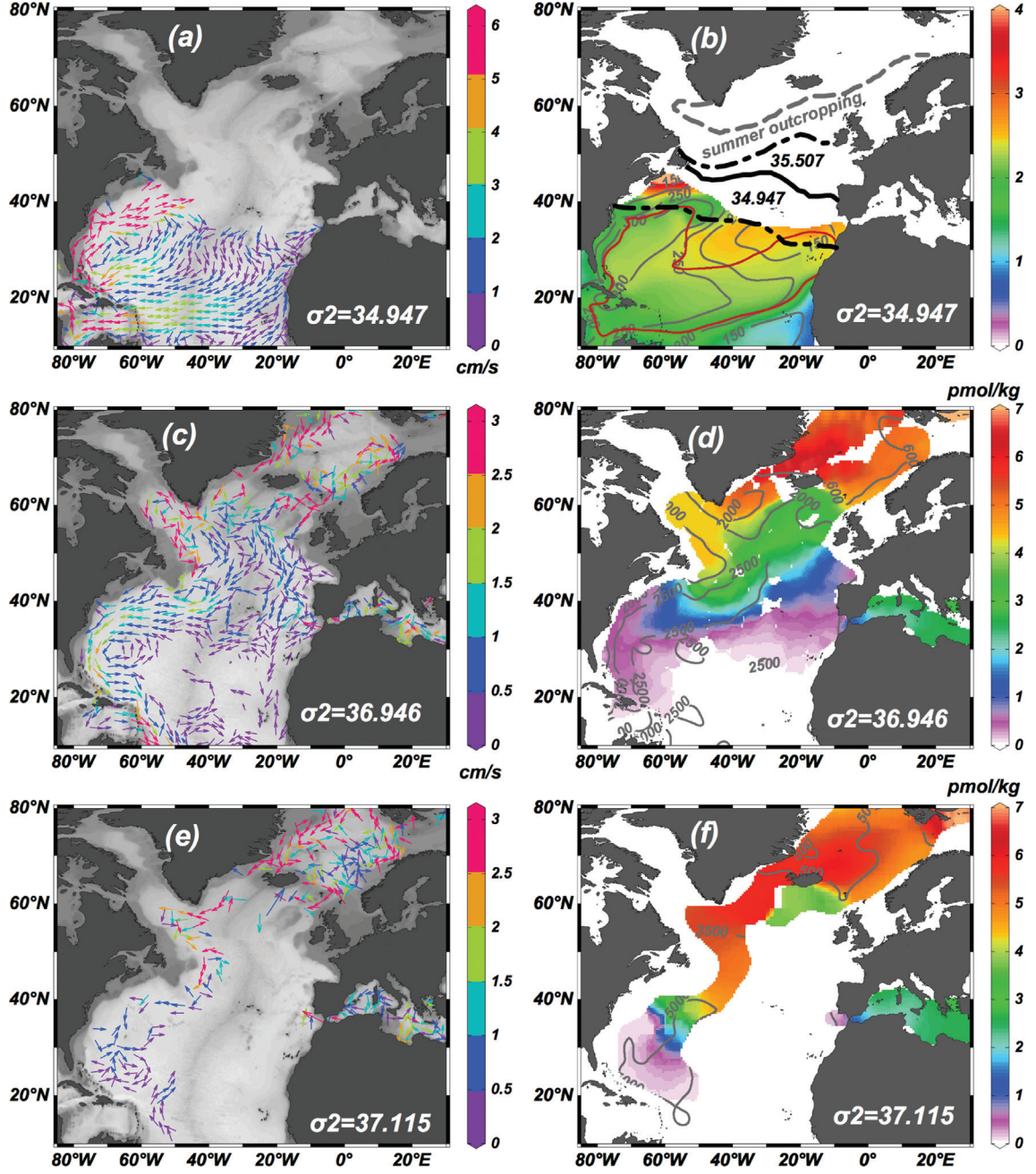


Fig. 7. The simulated ocean currents (a, c, e, unit: cm/s) and CFC-11 concentrations (b, d, f, unit: pmol/kg, note b has different scale) in February, 1998 on 34.947-, 36.946- and 37.115-isopycnals. The grey lines denote the depths of isopycnals. Area between two dash-dotted curves in (b) is the long-term mean outcropping areas for the 34.947 and 35.507-isopycnals during winter (January–February–March). The black solid (grey dashed) curve shows the long-term mean winter (summer) outcropping line for 35.507-isopycnal. The red contour in (b) is the 18°C isotherm line on the 34.947-isopycnal.

here). The long-term mean winter (January–February–March, JFM) and summer (July–August–September, JAS) outcropping areas for the 34.947- and 35.507-isopycnals are shown in Fig. 7b. Water mass subducted into the seasonal thermocline from summer outcropping area would be entrained back to the mixed layer during winter

as suggested by Marshall et al. (1993). It therefore implies that the outcropping areas during summer time would hardly contribute to the ventilation of the sandwich-like CFC-11 layer in the subtropical North Atlantic. On the other hand, isopycnals outcropping during winter would subducted into the permanent thermocline and thus

ventilate the ocean interior (Williams et al., 1995). The simulated winter outcropping area for the 34.947- and 35.507-isopycnals locates immediately south of the Gulf Stream and its extension–North Atlantic Current (figure not shown), where is suggested to be the ventilation sites for the EDW (e.g. Worthington, 1958). We therefore argued that the sandwich-like CFC-11 layer provide a tracer perspective about the EDW structure, and this sandwich-like CFC-11 water mass is ventilated through subduction process south of the Gulf Stream and North Atlantic Current during winter.

On the 36.946-isopycnal, the model clearly captures the DWBC with velocity of 1.5–2 cm/s (Fig. 7c), which is also illustrated by the distribution of CFC-11 concentrations (Fig. 7d). As previously shown, the tracer-enriched tongue at 2000–3000 m depth originates mainly in the western subpolar North Atlantic and the Nordic Seas. Similarly, the distribution of the tracer concentration on the 37.115-isopycnal exhibits a quick southward transport. However, the signal is largely confined to the bottom of the western basin instead of following the western boundary as that in the observations.

Wintertime (JFM) maximum mixed layer potential density and MLD are shown in Fig. 8a and 8b to further check the outcropping of the isopycnals. The maximum MLD in the Labrador Sea, Irminger Sea and central Nordic Seas exceeds 1000 m, indicating the open-ocean deep convective mixing. The isopycnals with potential density between 36.846 and 36.946 outcrops in the Labrador Sea, Irminger Sea and Nordic Seas, which means, according the definition of model's isopycnal layers, the 36.946-isopycnal layer in the model ventilates in these regions during winter. On the other hand, the 37.115-isopycnal layer outcrops in the central Nordic Seas. Therefore, the high concentrations of CFC-11 and idealised tracer along the 36.946-isopycnal between 2000 and 3000 m depth at 24.5°N are likely caused by the direct ventilation in the Labrador and Nordic Seas and the spreading along the 36.946-isopycnal. CFC-11 tracer concentrations from regional oceans shown in Fig. 6 also support that the strong tracer signal between 2000 and 3000 m depth in the subtropical North Atlantic is from both the western subpolar North Atlantic (tracer 25) and the Nordic Seas (tracer 27).

To further investigate the pathways and timescales of the tracer signals transporting from the western subpolar North Atlantic and Nordic Seas to the subtropical North Atlantic, the simulated CFC-11 concentrations contributed from western subpolar North Atlantic (tracer 25) and the simulated apparent age with source region of tracer 25 on the different isopycnals (Fig. 9) were examined. After being ventilated in the central Labrador Sea (high tracer concentration and young age of water mass), the water mass

on the 36.846-isopycnal primarily transports northeastward and passes the Iceland–Scotland Ridge into the Nordic Seas, and partly spreads southeastward into eastern North Atlantic. The spreading into the eastern North Atlantic is slower than that extending to the Nordic Seas. The southeastward flowing water reaches the eastern coast of the subtropical North Atlantic and elevates the CFC-11 concentrations along the eastern coast of the North Atlantic, which appears as the water mass with relatively high CFC-11 concentrations between 1000 and 2000 m depth in the eastern part of the 24.5°N section in Figs. 3 and 4. Therefore, the distribution of CFC-11 on the 36.846-isopycnal suggests that the ventilated water from the western subpolar North Atlantic is an important source for the elevated CFC-11 signal in the eastern subtropical North Atlantic in our simulation.

Early studies suggested that the export of the Labrador Sea Water follows: (i) to the northeast into the Irminger Sea; (ii) eastward into the eastern North Atlantic (Bower et al., 2009); and (iii) southeastward with the DWBC (Talley and McCartney, 1982; Rhein et al., 2002). The model results show that the spreading of the Labrador Sea Water is in general agreement with the early studies (Fig. 9c). The apparent age distribution indicates that the 36.946-isopycnal ventilates in the coastal region of the Labrador Sea and in the Nordic Seas where the apparent age is low (Fig. 9d). The deep 37.115-isopycnal only outcrops in the Nordic Seas (Figs. 8 and 9e, f) illustrated by high CFC-11 regional tracer concentration and low apparent age in the Nordic Seas. Therefore, the tracer-enriched surface water in the Labrador Sea transports into the Nordic Seas where it is transformed to denser water and then flows southward mainly through the Denmark Strait to the subtropical North Atlantic (Fig. 9e, f).

As the combined impact of advection, mixing and recirculation, the surface water from the western subpolar North Atlantic (tracer 25) takes around 12 yr to reach the Nordic Sea where the 36.946-isopycnal outcrops and about 30 yr to reach the subtropical North Atlantic following the 36.946-isopycnal whereas it takes around 36 yr to reach the subtropical North Atlantic mainly via the Denmark Strait and fills the bottom of the western basin of the North Atlantic following 37.115-isopycnal. As is discussed in the Section 3.1, the 36.946-isopycnal dominates the southward volume transports, it is therefore the fastest route for water mass transporting from the subpolar to the subtropics. The transit-time for water mass on the 36.946-isopycnal transporting from the Labrador Sea to 24.5°N in the subtropical North Atlantic is about 10 yr. Similarly, the transit-time for the OW on the 37.115-isopycnal from the Nordic Seas to the subtropics is around 20 yr. The transit-times for the water mass from the subpolar North Atlantic to the subtropical and equatorial Atlantic have large

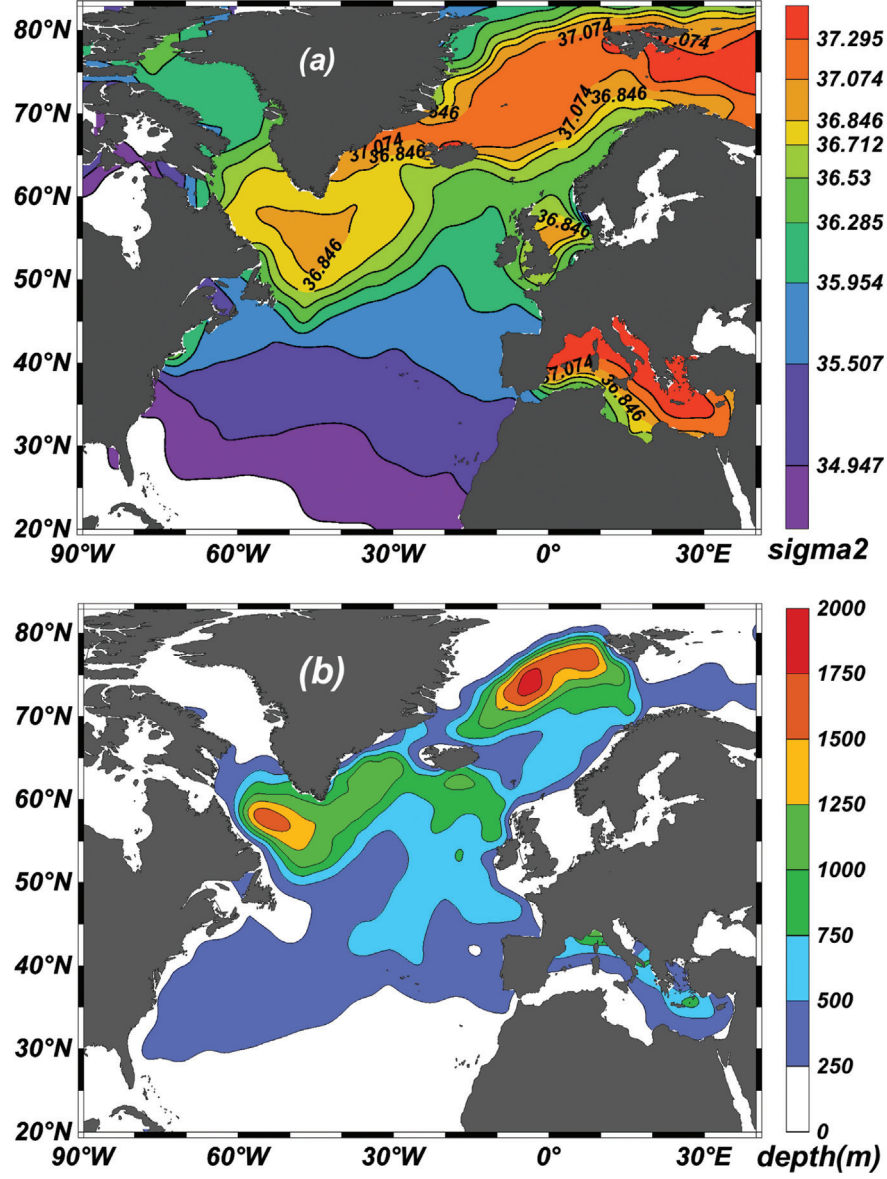


Fig. 8. Simulated winter time maximum mixed layer potential density (a) and depth (b) in the North Atlantic during winter (January to March from 1948–2007). The potential density contours are the same as the model’s isopycnals.

discrepancies among different methods from both observations and simulations, ranging from 8–10 yr to more than 25–30 yr (see Fine et al., 2002, for brief review). Our estimate of the transit-time from the subpolar North Atlantic to the deep subtropical and tropical Atlantic is generally within these ranges.

4. Dominance of ocean dynamics on tracer variability and its propagation

The variations in the overturning circulation can substantially affect meridional ocean heat transport and

therefore has significant impact on global (Knight, 2005; Lynch-Stieglitz et al., 2007) and especially North Atlantic climate (Sutton and Hodson, 2005; Pohlmann et al., 2006). A recent study showed that the prediction of multiyear changes in the AMOC is possible (Matei et al., 2012). Early studies suggested that the variability in hydrographic observations can provide information about the variability of ocean circulation (Mauritzen et al., 2006) and the variabilities in the subpolar North Atlantic and Nordic Seas could be indicators for the changes of AMOC in the subtropical North Atlantic (e.g. Hawkins and Sutton, 2008; Lohmann et al., 2008). Here we examine whether

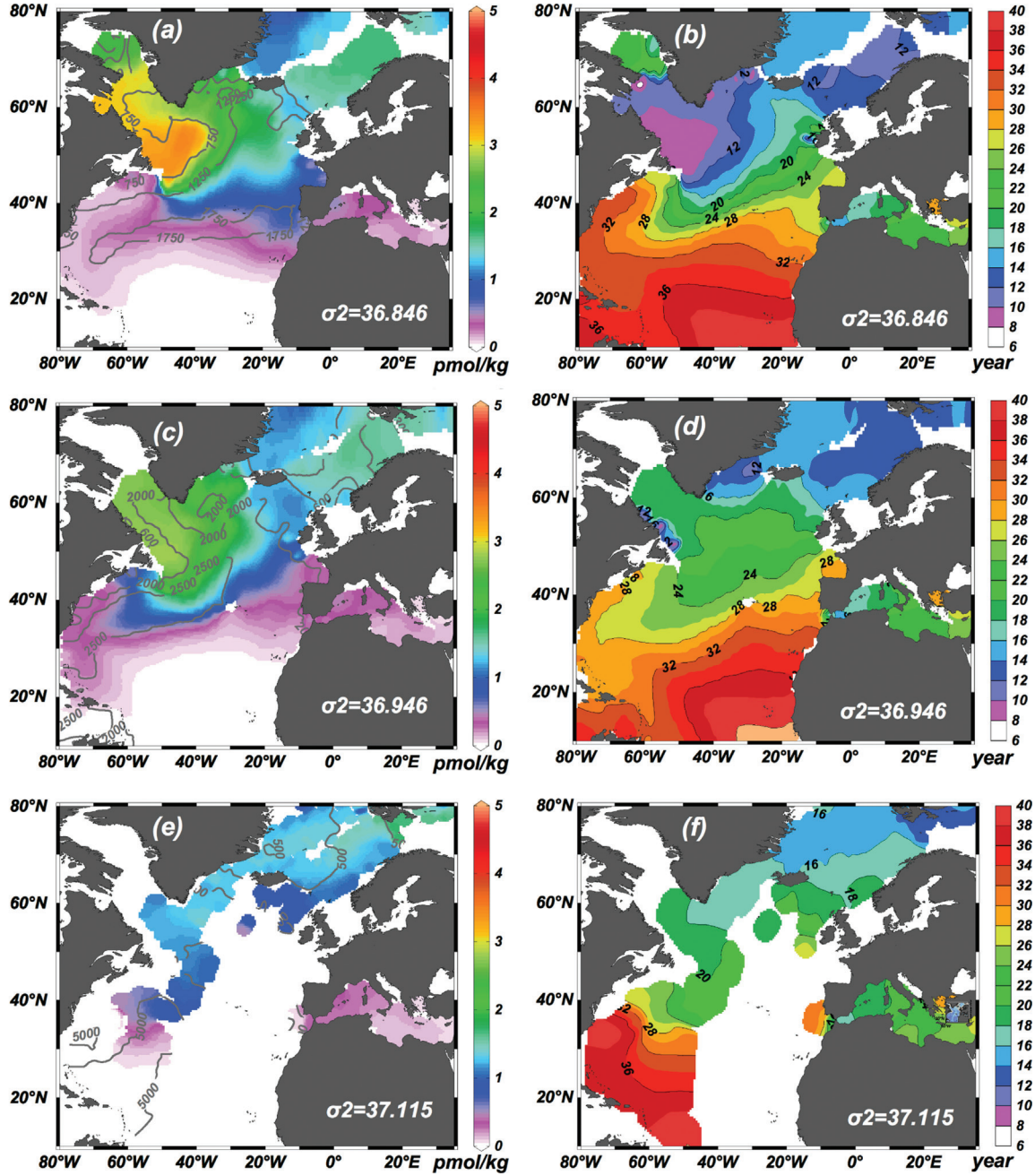


Fig. 9. Simulated CFC-11 released from western subpolar North Atlantic (region 25 denoted in Fig. 1; left panels, unit: pmol/kg) in February 1998 and simulated apparent age (right panel, unit: year) of idealised tracer released from region 25 on 36.846-, 36.946- and 37.115-siopeyanls at the end of 60-yr simulation; contours are depth of the isopycnals (left panel) and age of tracer 25 (right panel) on isopycnals.

the variability in the subpolar North Atlantic can be used to predict the AMOC variability in the subtropical North Atlantic in the simulation and whether the variability in passive tracers (both CFC-11 and idealised tracer) can represent the variability in the hydrography (such as temperature or salinity).

The time evolutions of CFC-11 and idealised tracer concentrations, layer thickness and temperature of 36.946-isopycnal at 24.5°N within the DWBC were shown in Fig. 10a. The salinity below the mixed layer is determined by a simplified state of equation in the model; the simulated salinity therefore has similar variations as the temperature

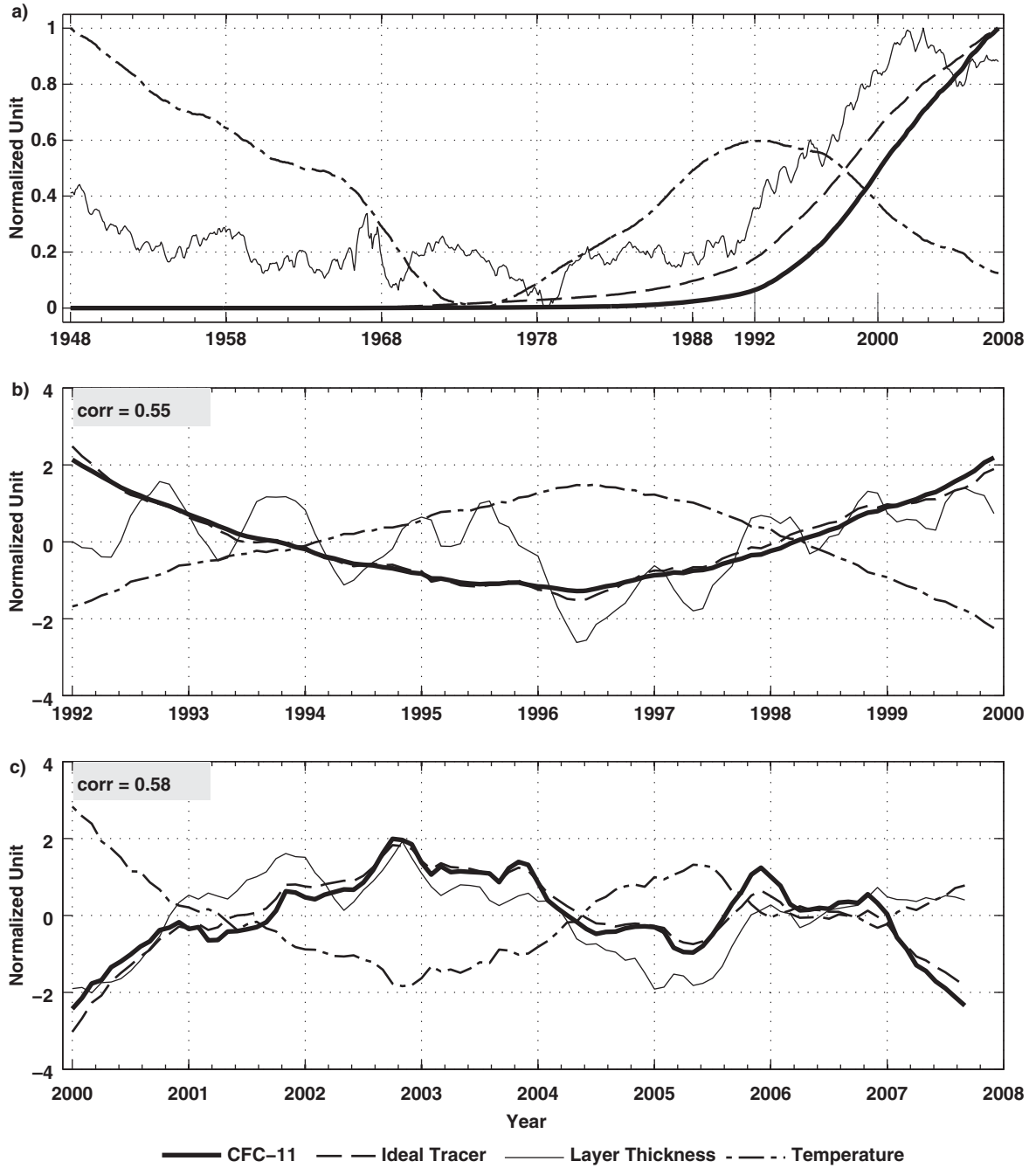


Fig. 10. (a) Time series of simulated monthly mean CFC-11 (thick line), ideal tracer (dashed line), model isopycnal layer thickness (thin line) and ocean temperature (dash-dot line) from 1948 to 2008 for the tracer-enriched part of 36.946-isopycnal at around 2500 m depth of 24.5°N in Figs. 4 and 5. Locations of the involved model grids are also denoted as blue dots in the inset map of Fig. 11. All variables are scaled to 0–1. (b) Same as panel a, but for years from 1992 to 2000 and each variable is detrended and normalised by its standard deviation. (c) Same as panel b but years from 2000 to 2008. Correlations between different variables have been calculated and the minimum one (CFC vs. Layer Thickness) is annotated in each panel.

and will not be discussed here. The idealised tracer and CFC-11 have similar long-term variations though they have different source functions (i.e. spatially and temporally varied CFC-11 flux versus constant flux for idealised

tracer). In order to investigate the short-term variabilities of the tracer concentration, we should firstly remove the long-term variations of the tracer concentration. One straightforward case to remove the long-term variation is

when the time series has a linear trend. Both the simulated idealised tracer and CFC-11 concentrations started to increase almost linearly since 1992, and this increase in CFC-11 around 1992 was also reported by a continuous monitoring in the DWBC at 26.5°N in the subtropical North (see Molinari et al., 1998, Fig. 3). The linear increasing trends in both CFC-11 and idealised tracer concentrations slightly slow down around the year 2000 which could be roughly regarded as an inflection point of the linear trend (Fig. 10a).

To examine the short-term variability in tracer concentrations, the time series of the normalised monthly tracer concentrations from 1992 to 2000 and from 2000 to 2008 are shown in Fig. 10b and 10c after their linear trends are removed. Surprisingly, the variability in temperature (negatively correlated), CFC-11 and idealised tracer concentrations are almost identical, implying the importance of ocean dynamic in shaping the short-term variability. The results suggested that the idealised tracer could be used to trace the short-term variability in CFC-11 and temperature though they have different source functions. It should be noted that the variability of isopycnal thickness well matches that of the tracers from 2000 to 2008. The negative correlation between the temperature and isopycnal thickness likely reflects the impact of wintertime deep convection in the formation of dense water in high latitudes: strong (weak) convection produces a thick (thin), fresh (saline) and cold (warm) layer (Curry and McCartney, 1998). Orre et al. (2007) also suggested that the short-term variability of surface Technetium-99 in the western coast of Norway is governed by the ocean dynamics instead of the source function.

Previous studies also indicated that the temperature and salinity or water density anomalies at one ocean site could possibly be traced upstream with a certain leading time by both the observational evidences (Molinari et al., 1998; Dickson et al., 1999; 2002; Eldevik et al., 2009) and model simulations (Hawkins and Sutton, 2008; Van Sebille et al., 2011). We therefore investigate how and on what timescales the signals in the subtropical Atlantic can be related to its upstream source regions. Three sections (52.0°N, 40.0°N and 24.5°N) are selected to trace the propagation of the anomalies along the DWBC (Fig. 11a and 11b).

The winter time MLD, which is the indicator of deep convection, shows that the Labrador Sea experienced an intense convection in the beginning of the 1990s and was followed by a restratification until around 2006 as revealed by observations (Lilly et al., 1999; Lazier et al., 2002; Louarn et al., 2009). The simulated variabilities of the isopycnal thickness at the 52.0°N sections are highly correlated with that of the MLD on decadal timescales, and the isopycnal thickness among different sections are also highly correlated, implying the propagation of the anomaly

lies from the Labrador Sea to the subtropical Atlantic (Fig. 11a). The simulated temperatures show a multi-decadal variability and also have an obvious time-lag correlation between different sections (Fig. 11b).

The transit-time derived by ‘optimum time lag’ approach (or cross-correlation) (Brown et al., 2002; Delhez and Deleersnijder, 2008) is about 6.3 yr for the simulated isopycnal thickness anomaly to propagate from 52.0°N to 40.0°N, and around 9.3 yr to 24.5°N. The transit-time for propagation of the temperature anomaly is longer than that of the thickness anomaly, taking about 8.9 yr from 52.0°N to 40.0°N and 13.6 yr to 24.5°N. The time series of anomalies between different sections are highly correlated with correlation coefficient larger than 0.91 for isopycnal thickness anomaly and 0.85 for temperature anomaly. It should be mentioned that the magnitude of the anomaly is largely reduced from 52.0°N to 24.5°N, for example, the magnitude of the temperature anomaly at 24.5°N is only about 25% of that at 52.0°N in the simulation. The reduction in amplitude is larger than the observation-based estimate with the amplitude decays by 50% (Van Sebille et al., 2011) but much smaller compared with the analytical model of Waugh and Hall (2005) which shows an anomaly in the subpolar source was reduced to only 2% in the tropics.

However, the long-term near linear increase of CFC-11 and idealised tracer concentrations in the ocean (Fig. 10a) makes it difficult to detect the propagation of signals by comparing the similarities of the long-term variabilities. We used the detrended variability of CFC-11 and idealised tracer at 24.5°N (Fig. 10b and 10c) and search the maximum lagged correlation variability in the upstream sections. The correlated variabilities of idealised tracer and CFC-11 along the DWBC sections are shown in Fig. 12.

The similar variabilities for normalised CFC-11 and idealised tracers, after detrended, confirm the dominance of the ocean dynamics on the tracer variabilities. It also shows the possibility to track the tracer anomalies upstream, and the transit-time is 9.3 to 10.0 yr for the passive tracer to propagate from the 52.0°N to 40.0°N and 11.6 to 13.5 yr to the 24.0°N. The transit-time in 1980–1990 is longer than that in 1990–2000. The observation-based estimates of transit-time between the central Labrador Sea and 26.5°N in the subtropical North Atlantic ranges from 9 to 10 yr (Molinari et al., 1998; Van Sebille et al., 2011).

We notice that transit-time derived by cross-correlation (or time-lag) method is shorter compared with the ‘apparent age’ obtained in Section 3.3. The ‘true mean age’ of a water mass at a certain location is actually a spectrum distribution (Delhez et al., 1999). The cross-correlation (or time-lag) method does not sample the whole age distribution but only part of the age spectrum around its maximum, and therefore suffers from a systematic

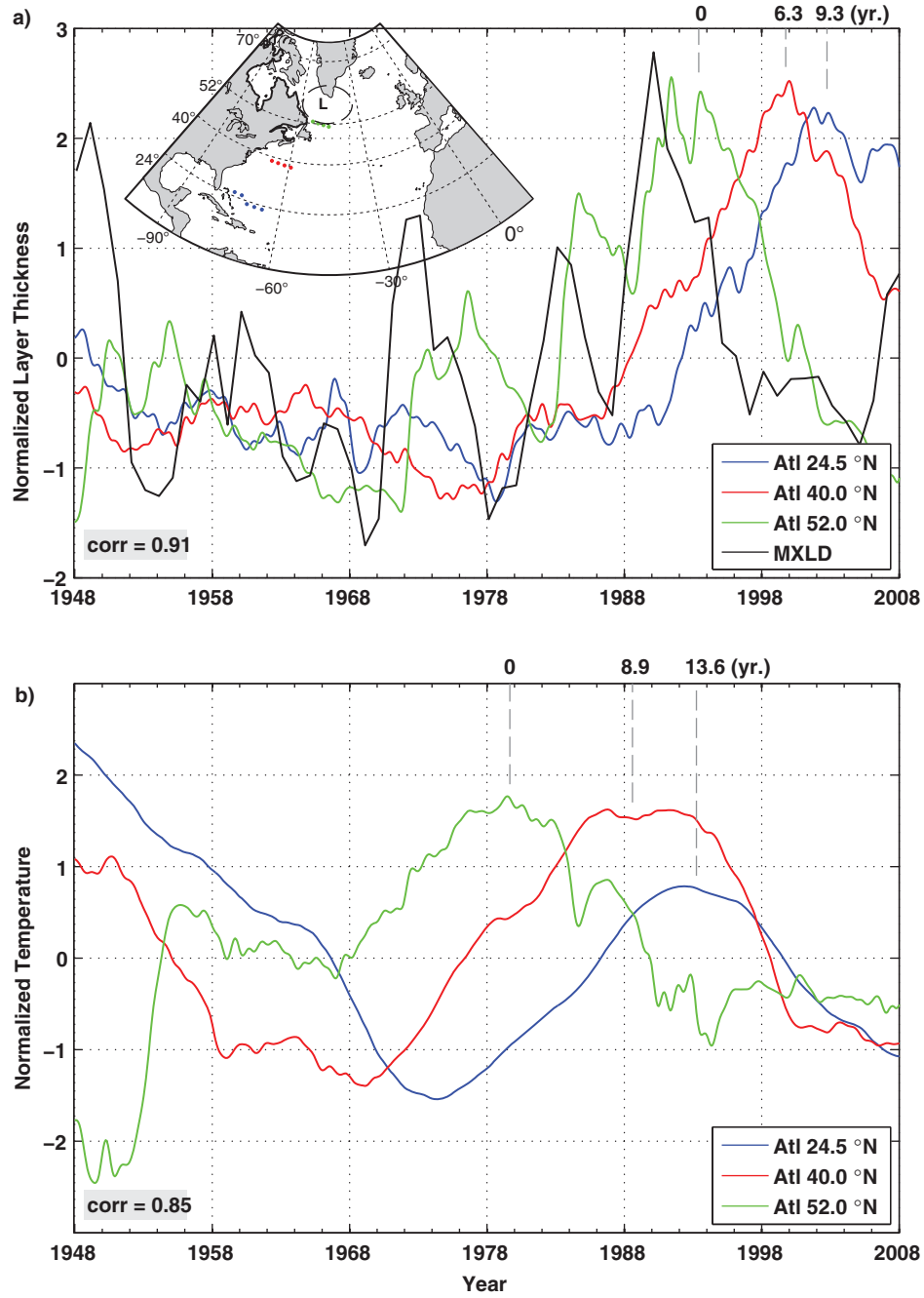


Fig. 11. (a) Time series (smoothed by a 12-month low-pass filter) of simulated monthly mean isopycnal thickness for 36.946-isopycnal at 24.5°N (blue line, the same data as that in Fig. 10a), 40.0°N (red line) and 52.0°N (green line). Model grids selected for each section are denoted in the inset map by blue, red and green dots for 24.5°N, 40.0°N and 52.0°N respectively. Simulated wintertime (JFM) mean MLD in the Labrador Sea is also plotted with black line. The region for calculating the mean MLD is denoted by a dashed ellipse with a bold L inside. (b) Same as panel a, but for simulated ocean temperature. All the variables are normalised and the time lags (in months) between the section at 52.0°N and its downstream section at 40.0°N and 24.5°N are annotated in the top of each panel. Correlation between 24.5°N and 52.0°N is also annotated in each panel.

underestimation of the true mean age. Therefore, the transit-time derived by cross-correlation method is shorter compared with the ‘apparent age’, which is also shown in

Section 3.3. The apparent age is defined as mass-weighted mean age from different sources. The relation between these two methods is clarified by Delhez and Deleersnijder

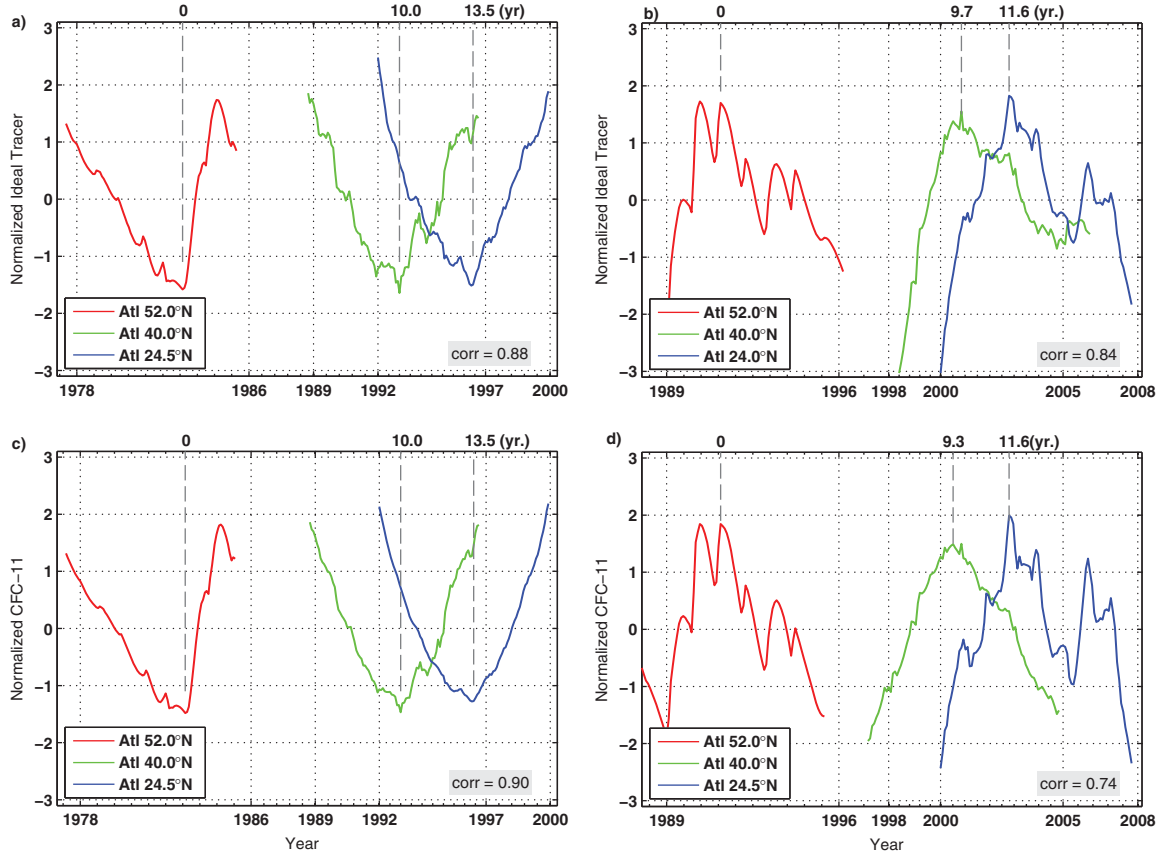


Fig. 12. (a) Variability of ideal tracer concentrations at 24.5°N from 1992 to 2000 (blue curve) from Fig. 10b, and its upstream maximum lag-correlated variability at 40.0°N from 1989 to 1997 (green curve) and that at 52.0°N from 1978 to 1986 (red curve). (b) Same as panel a but for variability from 2000 to 2008 at 24.5°N (blue curve), together with its upstream variability at 40.0°N (green curve) and 52.0°N (red curve). (c) Same as panel a but for simulated CFC-11. (d) Same as panel b but for simulated CFC-11. Correlation between 24.5°N and 52.0°N is also annotated in each panel.

(2008). It also should be noted that the transit-times derived by cross-correlation method from different water mass properties (such as CFC-11, temperature and isopycnal thickness) are also different, and the possible reasons for these discrepancies are discussed in the following section.

5. Summary and discussion

To illuminate the decadal scale ventilation of the North Atlantic, the anthropogenic chlorofluorocarbons (CFC-11 and CFC-12), 29 geographically defined CFC-11 tracers and 29 idealised passive tracers attached with age tracers have been incorporated into a global isopycnal coordinate ocean model to investigate the sources and pathways spreading timescales of water masses in the world oceans.

The simulated CFC-11 along 24.5°N in the North Atlantic in February 1998 is generally consistent with observations, particularly between 0–800 m. Both the observed

and simulated CFC-11 distributions show a sandwich-like pattern in the 0–800 m, which is persistent year-round. The results from our geographically-defined CFC-11 tracer simulation suggest that the observed sandwich-like CFC-11 distribution is mainly caused by the subduction of the surface water from both the western and eastern subpolar North Atlantic with contributions of 56 and 16% respectively. The model also clearly captures the feature of the CFC-11 intrusion between 1000 and 2000 m depth in the eastern North Atlantic, which is mainly originated from the western subpolar North Atlantic in our simulation (Fig. 6e and 9a). On the contrary, LeBel et al. (2008) tended to regard this CFC-enriched water-mass as a signal of the MOW. Below 1000 m depth, the observation exhibits two local CFC-11 maxima at 1500 and 3700 m depth, indicating the so-called ‘Double Flow’ phenomenon which is difficult to be properly captured by the OGCM simulation (Dutay et al., 2002; Gao et al., 2003). Early studies (e.g. England and Holloway, 1998; England and Maier-Reimer, 2001) indi-

cated that a realistic simulation on the southward transport of the LSW and the dense water overflow off the Greenland–Iceland–Scotland Ridge was of crucial importance for capturing the two CFC-bearing cores. Our model, on the other hand, captures the two CFC-11 maxima in the western basin below 1000 m depth, reflecting the reasonable representation of the southward outflow of the LSW and the dense OW.

However, the locations of simulated isopycnal layers (Fig. 4) are lower than those in the observation (Fig. 3), which means the simulated water mass is less dense than the observations, and the presence of the bottom CFC-11 core in the model is also partly because the northward intrusion of the simulated AABW is too weak. Meanwhile, the simulated concentrations of CFC-11 core (0.25–0.3 pmol/kg) at around 2500 m depth are lower than the observations (0.75 pmol/kg), and the simulated eastern basin mid-depth CFC-11 signal is relatively stronger than that in the observations (Figs. 3 and 4). These discrepancies between the observation and simulation are possibly a consequence of the relatively sluggish DWBC and vigorous interior diffusion in the model. For instance, the simulated DWBC velocity of 1.5–2 cm/s is slower than hydrographic data derived effective spreading rate of 2–5 cm/s (Molinari et al., 1998; Stramma and Rhein, 2001; Freudenthal, 2002). A sluggish DWBC is usually related to a coarse model resolution (e.g. England and Holloway, 1998; Beismann, 2003) and can be sensitive to the parameterisations of both the diapycnal and isopycnal mixing (Gao et al., 2003).

The results from the geographically defined CFC-11 tracers and idealised tracers reveal that the western and eastern subpolar North Atlantic and the Nordic Seas are the most efficient areas to ventilate the intermediate and deep subtropical Atlantic Ocean. The simulated CFC-11 core at 2500 m depth exhibits an origin from both the western and eastern subpolar North Atlantic (64.1 and 13.5% respectively) and the Nordic Seas (16.4%) following the simulated DWBC on the 36.946-isopycnal. Composition of tracer-enriched water mass deeper than 4500 m shows that the Nordic Seas dominates (61.0%) the CFC-11 ventilation in the deep subtropical North Atlantic. The observational studies, on the other side, tend to conclude that the Labrador Sea is the only source for tracer-enriched water at 2500 m depth (e.g. Smethie et al., 2000, 2007), but it is hard to estimate how much the CFC-11 maximum comes from Labrador Sea and the Nordic Seas based on the observational data. However, it is also possible that the model does not produce the overflow well, which may lead to spurious mixing with the LSW.

The CFC-11 and idealised tracer have different air–sea gas flux. Therefore, water mass composition of CFC-11 indicates the combined effects of temperature and ocean

ventilation, while the water mass composition of idealised tracers with a prescribed constant air–sea flux is determined only by the ocean ventilation. However, the idealised tracers still reflect the relative importance of different ocean regions for ventilating the subtropical North Atlantic and indicate the important role of ocean dynamics and the possibility to use the idealised tracers to mirror the temporal and spatial variability of other passive tracers. The surface water in the western subpolar North Atlantic flows into the eastern Nordic Seas and the Barents Sea where the 36.946- and 37.115-isopycnal outcrop, and thereafter flow southward and enter the North Atlantic. The apparent age distribution shows that it takes around 12 yr for the surface water from the western subpolar North Atlantic to reach the Nordic Seas where the water mass is transformed and thereafter flows southward. The simulated apparent age shows that the transformed water mass in the Nordic Seas needs about 20 yr to reach the deep subtropical North Atlantic. The apparent age in this paper is an average of tracer concentration-weighted ages of different water masses, and it takes into account recirculation gyres and mixing and advection. Therefore, the apparent age tends to be longer than that estimated by direct velocity measurement (see Fine et al., 2002, for a brief review).

Previous studies have reported that anomalies in hydrographic properties like temperature, salinity and isopycnal thickness or anthropogenic passive tracers like CFC-11 are highly correlated along the transport pathways between the surface subpolar North Atlantic and Nordic Seas (Dickson et al., 1999; 2002; Hawkins and Sutton, 2008; Eldevik et al., 2009) or between subpolar North Atlantic and subtropical deep North Atlantic (Curry and McCartney, 1998; Molinari et al., 1998; Stramma and Rhein, 2001; Freudenthal, 2002; Van Sebille et al., 2011). Our study confirms that the source functions of tracers and the ocean dynamics together determine the trend of passive tracers and that the variabilities of both the dynamically active (such as water temperature and salinity) and passive (such as CFCs and idealised) tracer concentrations are highly correlated after removing their linear trend, indicating the dominant role of ocean dynamics. The results also showed that the variabilities of the tracers (temperature, CFCs and idealised tracers) in the deep subtropical North Atlantic could be traced back to the upstream subpolar region.

The transit-times from subpolar to subtropical North Atlantic derived by cross-correlation or optimum time lag are different among various quantities, for example, 9.3 yr for the isopycnal thickness anomaly, 13.6 yr for the temperature anomaly and 11.6 to 13.5 yr for the passive tracer signals (CFC-11 and idealised tracer). The possible interpretations of these discrepancies in the transit-times for isopycnal thickness anomaly, temperature

(salinity) and passive tracers are, however, under debate. Liu and Shin (1999) pointed out that the propagation of temperature and passive tracer anomalies in a coarse OGCM have comparable pathways as mean flow, but the former could have a slower propagation speed but decay faster, and have a more complex spatial structure than that of the latter. Using a regional OGCM, Nonaka and Xie (2000) even demonstrated that passive tracers and temperature anomaly could propagate in opposite directions when the propagation of temperature anomaly is dominated by high baroclinic modes of Rossby waves. Finally, based on a simplified coupled ocean–atmosphere model, Lazar et al. (2001) suggested a compromise of the role of mean flow advection and wave processes that the mean advection flow determines the first approximation of the timescale and transport pathways for heat anomalies while wave processes tend to influence the evolution of the intensity and the shape of the anomalies. Our study showed that the transit-time for passive tracers (CFC-11 and idealised tracer) is close to that for temperature (salinity), but is longer than that for isopycnal thickness.

In terms of methodology, the geographically defined artificial tracers can potentially be used to trace the spreading pathway and the timescales of environment-harmful pollution (such as radioactive waste) in certain area of the world ocean.

6. Conclusions

By using the simulation on the 29 geographically defined CFC-11 tracers and 29 idealised tracers plus the age tracers, we found that:

- (1) The sandwich-like distribution of CFC-11 concentration in the upper 500 m of subtropical North Atlantic in both the observations and simulations is mainly caused by subduction from western and eastern subpolar North Atlantic, but the contribution of the former (56.0%) is almost four times larger than that of the latter (15.7%).
- (2) We demonstrated that the ocean dynamics, instead of the source function, determine the annual and inter-annual variability in both dynamically active tracer (such as water temperature and salinity) and passive tracer (such as CFC-11 and idealised tracer) concentrations in the deep North Atlantic.
- (3) The ‘apparent age’ distribution shows that the surface water in the western subpolar North Atlantic takes about 12 yr to reach the Nordic Seas and takes 20 yr from the Nordic Seas to the subtropical deep North Atlantic. The transit-time derived by ‘optimum time lag’ approach shows a 9.3–13.6 yr lag for the signals propagating from the western subpolar

North Atlantic to the subtropical North Atlantic, which is generally consistent with that of about 10 yr derived from the ‘apparent age’.

7. Acknowledgments

This study is supported by European Union’s Seventh Framework Programme (FP7/2007-2013) under grant agreement No. 212643 and by the Chinese National Basic Research Program ‘Development and Improvement of Ecosystem Model’ (Grant: 2010CB951800) and Norwegian Research Council supported project ‘NARE’ and ‘NorClim’. The authors are also grateful to Prof. Helge Drange for his valuable help and two anonymous reviewers for their comments to improve the manuscript. Data from the RAPID-WATCH MOC monitoring project are funded by the Natural Environment Research Council and are freely available from <http://www.noc.soton.ac.uk/rapidmoc>. We thank Schlitzer, R. (2011) for providing Ocean Data View (available at <http://odv.awi.de>).

References

- Barnett, T. P., Pierce, D. W., Achutarao, K. M., Gleckler, P. J., Santer, B. D. co-authors. 2005. Penetration of human-induced warming into the world’s oceans. *Science*. **309**, (5732), 284–287. DOI: 10.1126/science.1112418.
- Bates, N. R., Pequignat, A. C., Johnson, R. J. and Gruber, N. 2002. A short-term sink for atmospheric CO₂ in subtropical mode water of the North Atlantic Ocean. *Nature*. **420**(6915), 489–493. DOI: 10.1038/nature01253.
- Beismann, J.-O. 2003. Model simulations of CFC uptake in North Atlantic deep water: effects of parameterizations and grid resolution. *J. Geophys. Res.* **108**(C5), 1–16. DOI: 10.1029/2001JC001253.
- Blanke, B., Speich, S., Madec, G. and Maugé, R. 2002. A global diagnostic of interior ocean ventilation. *Geophys. Res. Lett.* **29**(8), 1–4. DOI: 10.1029/2001GL013727.
- Bleck, R., Rooth, C., Hu, D. and Smith, L. T. 1992. Salinity-driven thermocline transients in a wind- and thermohaline-forced isopycnal coordinate model of the North Atlantic. *Phys. Oceanogr.* **22**(12), 1486–1505. DOI: 10.1175/1520-0485(1992)022<1486:SDTTIA>2.0.CO;2.
- Bower, A. S., Lozier, M. S., Gary, S. F. and Böning, C. W. 2009. Interior pathways of the North Atlantic meridional overturning circulation. *Nature*. **459**(7244), 243–7. DOI: 10.1038/nature07979.
- Broecker, W. S., Blanton, S., Smethie, W. M. and Ostlund, G. 1991. Radiocarbon decay and oxygen utilization in the Deep Atlantic Ocean. *Global Biogeochem. Cy.* **5**(1), 87. DOI: 10.1029/90GB02279.
- Brown, J., Iospje, M., Kolstad, K., Lind, B., Rudjord, A. and co-authors. 2002. Temporal trends for ⁹⁹Tc in Norwegian coastal environments and spatial distribution in the Barents Sea. *J.*

- Environ. Radioact.* **60**(1–2), 49–60. DOI: 10.1016/S0265-931X(01)00095-9.
- Bryden, H. L., Longworth, H. R. and Cunningham, S. A. 2005. Slowing of the Atlantic meridional overturning circulation at 25 degrees N. *Nature*. **438**(7068), 655–7. DOI: 10.1038/nature04385.
- Cunningham, S. A., Kanzow, T., Rayner, D., Baringer, M. O., Johns, W. E. and co-authors. 2007. Temporal variability of the Atlantic meridional overturning circulation at 26.5 degrees N. *Science*. **317**(5840), 935–8. DOI: 10.1126/science.1141304.
- Curry, R. and McCartney, M. 1998. Oceanic transport of subpolar climate signals to mid-depth subtropical waters. *Nature*. 575–577. DOI: 10.1038/35356.
- Deleersnijder, E., Campin, J.-M. and Delhez, E. J. M. 2001. The concept of age in marine modelling I. Theory and preliminary model results. *J. Mar. Syst.* **28**(3–4), 229–267. DOI: 10.1016/S0924-7963(01)00026-4.
- Delhez, E. J., Campin, J.-M., Hirst, A. C. and Deleersnijder, E. 1999. Toward a general theory of the age in ocean modelling. *Ocean Model.* **1**(1), 17–27. DOI: 10.1016/S1463-5003(99)00003-7.
- Delhez, E. J. and Deleersnijder, E. 2008. Age and the time lag method. *Continental Shelf Res.* **28**(8), 1057–1067. DOI: 10.1016/j.csr.2008.02.003.
- Dickson, B., Meincke, J., Vassie, I. and Jungclauss, J. 1999. Possible predictability in overflow from the Denmark Strait. *Nature*. **397**(January), 243–246. DOI: 10.1038/16680.
- Dickson, B., Yashayaev, I., Meincke, J., Turrell, B., Dye, S. and co-authors. 2002. Rapid freshening of the deep North Atlantic Ocean over the past four decades. *Nature*. **416**(6883), 832–7. DOI: 10.1038/416832a.
- Dutay, J., Bullister, J. L., Doney, S. C., Orr, J. C., Najjar, R. and co-authors. 2002. Evaluation of ocean model ventilation with CFC-11: comparison of 13 global ocean models. *Ocean Model.* **4**(2), 89–120. DOI: 10.1016/S1463-5003(01)00013-0.
- Eldevik, T., Nilsen, J. E. O., Iovino, D., Anders Olsson, K., Sandø, A. B. and co-authors. 2009. Observed sources and variability of Nordic seas overflow. *Nat. Geosci.* **2**(6), 406–410. DOI: 10.1038/ngeo518.
- England, M. H. 1995a. Using chlorofluorocarbons to assess ocean climate models. *Geophys. Res. Lett.* **22**(22), 3051. DOI: 10.1029/95GL02670.
- England, M. H. 1995b. The age of water and ventilation timescales in a global ocean model. *J. Phys. Oceanogr.* **25**(11), 2756–2777. DOI: 10.1175/1520-0485(1995)025<2756:TAOWAV>2.0.CO;2.
- England, M. H. and Holloway, G. 1998. Simulations of CFC content and water mass age in the deep North Atlantic. *J. Geophys. Res.* **103**(C8), 15885–15901. DOI: 10.1029/98JC00147.
- England, M. H. and Maier-Reimer, E. 2001. Using chemical tracers to assess ocean models. *Rev. Geophys.* **39**(1), 29–70. DOI: 10.1029/1998RG000043.
- Fine, R. A., Rhein, M. and Andri , C. 2002. Using a CFC effective age to estimate propagation and storage of climate anomalies in the deep western North Atlantic Ocean. *Geophys. Res. Lett.* **29**(24), 1–4. DOI: 10.1029/2002GL015618.
- Freudenthal, S. 2002. The arrival of a new Labrador sea water signal in the tropical Atlantic in 1996. *Geophys. Res. Lett.* **29**(15), 29–32. DOI: 10.1029/2002GL015062.
- Friedrich, H. and Levitus, S. 1972. An approximation to the equation of state for sea water, suitable for numerical ocean models. *J. Phys. Oceanogr.* **2**(4), 514–517. DOI: 10.1175/1520-0485(1972)002<0514:AATTEO>2.0.CO;2.
- Furevik, T., Bentsen, M., Drange, H., Kindem, I. K. T., Kvamst , N. G. and co-authors. 2003. Description and evaluation of the Bergen climate model: ARPEGE coupled with MICOM. *Clim. Dynam.* **21**(1), 27–51. DOI: 10.1007/s00382-003-0317-5.
- Ganachaud, A. 2003. Large-scale mass transports, water mass formation, and diffusivities estimated from World Ocean Circulation Experiment (WOCE) hydrographic data. *J. Geophys. Res.* **108**(C7). DOI: 10.1029/2002JC001565.
- Gao, Y. and Drange, H. 2004. The effect of diapycnal mixing on the ventilation and CFC-11 uptake in the Southern Ocean. *Adv. Atmos. Sci.* **21**(5), 755–766. DOI: 10.1007/BF02916372.
- Gao, Y., Drange, H. and Bentsen, M. 2003. Effects of diapycnal and isopycnal mixing on the ventilation of CFCs in the North Atlantic in an isopycnal coordinate OGCM. *Tellus B.* **55**(3), 837–854. DOI: 10.1034/j.1600-0889.2003.00068.x.
- Gao, Y., Drange, H., Bentsen, M. and Johannessen, O. M. 2005. Tracer-derived transit time of the waters in the eastern Nordic Seas. *Tellus B.* **57**(4), 332–340. DOI: 10.1111/j.1600-0889.2005.00153.x.
- Gao, Y., Drange, H., Johannessen, O. M. and Pettersson, L. H. 2009. Sources and pathways of 90Sr in the North Atlantic-Arctic region: present day and global warming. *J. Environ. Radioact.* **100**(5), 375–95. DOI: 10.1016/j.jenvrad.2009.01.003.
- Gaspar, P. 1988. Modeling the seasonal cycle of the upper ocean. *J. Phys. Oceanogr.* **18**(2), 161–180. DOI: 10.1175/1520-0485(1988)018<0161:MTSCOT>2.0.CO;2.
- Gnanadesikan, A. 2004. Oceanic ventilation and biogeochemical cycling: understanding the physical mechanisms that produce realistic distributions of tracers and productivity. *Global Biogeochem. Cy.* **18**(4). DOI: 10.1029/2003GB002097.
- Haine, T. W. N. and Hall, T. M. 2002. A generalized transport theory: water-mass composition and age. *J. Phys. Oceanogr.* **32**(6), 1932–1946. DOI: 10.1175/1520-0485(2002)032<1932:AGTTWM>2.0.CO;2.
- Hawkins, E. and Sutton, R. 2008. Potential predictability of rapid changes in the Atlantic meridional overturning circulation. *Geophys. Res. Lett.* **35**(11), 1–6. DOI: 10.1029/2008GL034059.
- Holzer, M. and Hall, T. M. 2000. Transit-time and tracer-age distributions in geophysical flows. *J. Atmos. Sci.* **57**(21), 3539–3558. DOI: 10.1175/1520-0469(2000)057<3539:TTATAD>2.0.CO;2.
- Kalnay, E. E., Kanamitsu, M., Kistler, R., Collins, W., Deaven, D. and co-authors. 1996. The NCEP/NCAR 40-year reanalysis project. *Bull. Am. Meteorol. Soc.* **77**(3), 437. DOI: 10.1175/1520-0477(1996)077<0437:TNYRP>2.0.CO;2.
- Keeling, R. F. and Garcia, H. E. 2002. The change in oceanic O(2) inventory associated with recent global warming. *Proceedings of the National Academy of Sciences of the United States of America* **99**(12), 7848–7853. DOI: 10.1073/pnas.122154899.

- Knight, J. R. 2005. A signature of persistent natural thermohaline circulation cycles in observed climate. *Geophys. Res. Lett.* **32**(20), L20708. DOI: 10.1029/2005GL024233.
- Kwon, Y.-O. 2004. North Atlantic subtropical mode water: a history of ocean-atmosphere interaction 1961–2000. *Geophys. Res. Lett.* **31**(19). DOI: 10.1029/2004GL021116.
- Lazar, A., Murtugudde, R. and Busalacchi, A. J. 2001. A model study of temperature anomaly propagation from the subtropics to tropics within the South Atlantic thermocline. *Geophys. Res. Lett.* **28**(7), 1271. DOI: 10.1029/2000GL011418.
- Lazier, J., Hendry, R., Clarke, A., Yashayaev, I. and Rhines, P. 2002. Convection and restratification in the Labrador Sea, 1990–2000. *Deep Sea Res. Oceanogr. Res. Paper.* **49**(10), 1819–1835. DOI: 10.1016/S0967-0637(02)00064-X.
- LeBel, D. A., Smethie, W. M. Jr, Rhein, M., Kieke, D., Fine, R. A. and co-authors. 2008. The formation rate of North Atlantic deep water and eighteen degree water calculated from CFC-11 inventories observed during WOCE. *Deep Sea Res. Oceanogr. Res. Paper.* **55**(8), 891–910. DOI: 10.1016/j.dsr.2008.03.009.
- Levitus, S. 2005. Warming of the world ocean, 1955–2003. *Geophys. Res. Lett.* **32**(2). DOI: 10.1029/2004GL021592.
- Levitus, S., Antonov, J. I., Wang, J., Delworth, T. L., Dixon, K. W. and co-authors. 2001. Anthropogenic warming of Earth's climate system. *Science*. **292**(5515), 267–270. DOI: 10.1126/science.1058154.
- Levitus, S., Burgett, R. and Boyer, T. 1994. World Ocean Atlas, Vol. 3, Salinity. NOAA Atlas NESDIS 3, U.S. Gov. Printing Office, Washington, DC, 99 pp.
- Lilly, J. M., Rhines, P. B., Visbeck, M., Davis, R., Lazier, J. R. N. and co-authors 1999. Observing deep convection in the Labrador Sea during winter 1994/95. *J. Phys. Oceanogr.* **29**(8), 2065–2098. DOI: 10.1175/1520-0485(1999)029<2065:ODCITL>2.0.CO;2.
- Liu, Z. and Shin, S.-I. 1999. On thermocline ventilation of active and passive tracers. *Geophys. Res. Lett.* **26**(3), 357. DOI: 10.1029/1998GL900315.
- Lohmann, G., Haak, H. and Jungclauss, J. H. 2008. Estimating trends of Atlantic meridional overturning circulation from long-term hydrographic data and model simulations. *Ocean Dynam.* **58**(2), 127–138. DOI: 10.1007/s10236-008-0136-7.
- Louarn, E., Mercier, H., Morin, P., de Boisseson, E. and Bacon, S. 2009. Upper Labrador Sea water in the Irminger Sea during a weak convection period (2002–2006). *Ocean Sci. Discuss.* **6**(3), 2085–2113. DOI: 10.5194/osd-6-2085-2009.
- Lynch-Stieglitz, J., Adkins, J. F., Curry, W. B., Dokken, T., Hall, I. R. and co-authors. 2007. Atlantic meridional overturning circulation during the last glacial maximum. *Science (New York, N.Y.)* **316**(5821), 66–69. DOI: 10.1126/science.1137127.
- Marshall, J. C., Williams, R. G. and Nurser, A. J. G. 1993. Inferring the subduction rate and period over the North Atlantic. *J. Phys. Oceanogr.* **23**(7), 1315–1329. DOI: 10.1175/1520-0485(1993)023<1315:ITSRAP>2.0.CO;2.
- Matei, D., Baehr, J., Jungclauss, J. H., Haak, H., Muller, W. A. and co-author. 2012. Multiyear Prediction of Monthly Mean Atlantic Meridional Overturning Circulation at 26.5°N. *Science* **335**(6064), 76–79. DOI: 10.1126/science.1210299.
- Mauritzen, C., Hjøllø, S. S. and Sandø, A. B. 2006. Passive tracers and active dynamics: a model study of hydrography and circulation in the northern North Atlantic. *J. Geophys. Res.* **111**(C8), 1–15. DOI: 10.1029/2005JC003252.
- McDougall, T. J. and Jackett, D. R. 2005. An assessment of orthobaric density in the global ocean. *J. Phys. Oceanogr.* **35**(11), 2054–2075. DOI: 10.1175/JPO2796.1.
- Molinari, R. L., Fine, R. A., Wilson, W. D., Curry, R. G., Abell, J. and co-authors. 1998. The arrival of recently formed Labrador sea water in the deep western boundary current at 26.5°N. *Geophys. Res. Lett.* **25**(13), 2249. DOI: 10.1029/98GL01853.
- Nonaka, M. and Xie, S. 2000. Propagation of North Pacific interdecadal subsurface temperature anomalies in an ocean GCM. *Geophys. Res. Lett.* **27**(22), 3747. DOI: 10.1029/2000GL011488.
- Old, C. and Haines, K. 2006. North Atlantic subtropical mode waters and ocean memory in HadCM3. *Journal of Climate* **19**(7), 1126–1148. DOI: 10.1175/JCLI3650.1.
- Orre, S., Gao, Y., Drange, H. and Deleersnijder, E. 2008. Diagnosing ocean tracer transport from Sellafield and Dounreay by equivalent diffusion and age. *Adv. Atmos. Sci.* **25**(5), 805–814. DOI: 10.1007/s00376-008-0805-y.
- Orre, S., Gao, Y., Drange, H. and Nilsen, J. E. O. 2007. A reassessment of the dispersion properties of 99Tc in the North Sea and the Norwegian Sea. *J. Mar. Syst.* **68**(1–2), 24–38. DOI: 10.1016/j.jmarsys.2006.10.009.
- Palter, J. B., Lozier, M. S. and Barber, R. T. 2005. The effect of advection on the nutrient reservoir in the North Atlantic subtropical gyre. *Nature* **437**(7059), 687–692. DOI: 10.1038/nature03969.
- Peacock, S. and Maltrud, M. 2006. Transit-time distributions in a global ocean model. *J. Phys. Oceanogr.* **36**(3), 474–495. DOI: 10.1175/JPO2860.1.
- Pohlmann, H., Sienz, F. and Latif, M. 2006. Influence of the multidecadal Atlantic meridional overturning circulation variability on European climate. *Journal of Climate* **19**(23), 6062–6067. DOI: 10.1175/JCLI3941.1.
- Qiu, B. and Huang, R. X. 1995. Ventilation of the North Atlantic and North Pacific: subduction versus obduction. *J. Phys. Oceanogr.* **25**(10), 2374–2390. DOI: 10.1175/1520-0485(1995)025<2374:VOTNAA>2.0.CO;2.
- Rhein, M., Fischer, J., Smethie, W. M., Smythe-Wright, D., Weiss, R. F. and co-authors. 2002. Labrador Sea Water: pathways, CFC inventory, and formation rates. *J. Phys. Oceanogr.* **32**(2), 648–665. DOI: 10.1175/1520-0485(2002)032<0648:LSWPCI>2.0.CO;2.
- Roach, A. T., Aagaard, K., Pease, C. H., Salo, S. A., Weingartner, T. and co-authors. 1995. Direct measurements of transport and water properties through the Bering Strait. *J. Geophys. Res.* **100**(C9), 18443–18457. DOI: 10.1029/95JC01673.
- Siegenthaler, U. and Sarmiento, J. L. 1993. Atmospheric carbon dioxide and the ocean. *Nature* **365**(6442), 119–125. DOI: 10.1038/365119a0.
- Smethie, W. M. J. 1993. Tracing the thermohaline circulation in the western North Atlantic using chlorofluorocarbons. *Progress In Oceanography* **31**(1), 51–99. DOI: 10.1016/0079-6611(93)90023-7.

- Smethie, W. M., Fine, R. A., Putzka, A. and Jones, E. P. 2000. Tracing the flow of North Atlantic deep water using chloro-fluorocarbons. *J. Geophys. Res.* **105**(C6), 14297–14323. DOI: 10.1029/1999JC900274.
- Smethie Jr, W. M., LeBel, D. A., Fine, R. A., Rhein, M. and Kieke, D.: 2007, Strength and variability of the deep limb of the North Atlantic meridional overturning circulation from chloro-fluorocarbon inventories. In: *Ocean Circulation: Mechanisms and Impacts Past and Future Changes of Meridional Overturning* (eds. A. Schmittner, J. Chiang and S. Hemming), *Geophys. Monogr. Ser.*, AGU, Washington DC, pp. 119–130. DOI: 10.1029/173GM09.
- Steele, M., Morley, R. and Ermold, W. 2001. PHC: a global ocean hydrography with a high-quality Arctic Ocean. *Journal of Climate* **14**(9), 2079–2087. DOI: 10.1175/1520-0442(2001)014<2079:PAGOHW>2.0.CO;2.
- Stramma, L. and Rhein, M. 2001. Variability in the deep western boundary current in the equatorial Atlantic at 44°W. *Geophys. Res. Lett.* **28**(8), 1623. DOI: 10.1029/2000GL011777.
- Stuiver, M., Quay, P. D. and Ostlund, H. G. 1983. Abyssal water carbon-14 distribution and the age of the world oceans. *Science*. **219**(4586), 849–51. DOI: 10.1126/science.219.4586.849.
- Sutton, R. T. and Hodson, D. L. R. 2005. Atlantic Ocean forcing of North American and European summer climate. *Science*. **309**(5731), 115–8. DOI: 10.1126/science.1109496.
- Talley, L. and McCartney, M. 1982. Distribution and circulation of Labrador Sea water. *J. Phys. Oceanogr.* **12**(11), 1189–1205. DOI: 10.1175/1520-0485(1982)012<1189:DACOLS>2.0.CO;2.
- Talley, L. and Raymer, M. 1982. Eighteen degree water variability. *Journal of Marine Research* **40**(suppl), 757–775.
- Thiele, G. and Sarmiento, J. L. 1990. Tracer dating and ocean ventilation. *J. Geophys. Res.* **95**(C6), 9377–9391. DOI: 10.1029/JC095iC06p09377.
- Thomson, C. W. 1877. *The Voyage of the 'Challenger' Vol. 1: The Atlantic*, Macmillan, Old Tappan NJ.
- van Sebille, E., Baringer, M. O., Johns, W. E., Meinen, C. S., Beal, L. M. and co-authors. 2011. Propagation pathways of classical Labrador Sea water from its source region to 26°N. *J. Geophys. Res.* **116**(C12), 1–18. DOI: 10.1029/2011JC007171.
- Walker, S. J., Weiss, R. F. and Salameh, P. K. 2000. Reconstructed histories of the annual mean atmospheric mole fractions for the halocarbons CFC-11 CFC-12, CFC-113, and carbon tetrachloride. *J. Geophys. Res.* **105**(C6), 14285–14296. DOI: 10.1029/1999JC900273.
- Waugh, D., Haine, T. and Hall, T. 2004. Transport times and anthropogenic carbon in the subpolar North Atlantic Ocean. *Deep Sea Research Part I: Oceanographic Research Papers* **51**(11), 1475–1491. DOI: 10.1016/j.dsr.2004.06.011.
- Waugh, D. W. and Hall, T. M. 2005. Propagation of tracer signals in boundary currents. *J. Phys. Oceanogr.* **35**(9), 1538. DOI: 10.1175/JPO2779.1.
- Williams, R. G., Marshall, J. C. and Spall, M. A. 1995. Does Stommel's mixed layer demon work? *J. Phys. Oceanogr.* **25**(12), 3089–3102. DOI: 10.1175/1520-0485(1995)025<3089:DSMLW>2.0.CO;2.
- Worthington, L. 1958. The 18° water in the Sargasso Sea. *Deep Sea Res.* **5**(2–4), 297–305. DOI: 10.1016/0146-6313(58)90026-1.