Ventilation of the Subtropical North Atlantic: Locations and Times of Last Ventilation Estimated Using Tracer Constraints From GEOTRACES Section GA03

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Abstract The ventilation of the subtropical North Atlantic along GEOTRACES section GA03 is quantified in terms of where and how long ago water was last in the mixed layer. Measurements of T, S, PO4, CFC-11, CFC-12, SF6, and estimates of prebomb 14C are deconvolved for the boundary propagator G using a maximum-entropy approach. From G, we calculate the fractions of water last ventilated in specified surface regions Ωw. We estimate that (56 ± 13)% of the water deeper than 1,000 m was ventilated in northern high latitudes, (15 ± 5)% in the Mediterranean, and (27 ± 12)% in the Southern Ocean. Below the thermocline and outside the deep western boundary current, mean ages of Ωw-ventilated water exceed a century. Consequently, memory of where last ventilation occurred tends to get lost and the deep mean-age patterns of Ωw-ventilated water are broadly similar for all Ωw. The mean ventilation ages, averaged over the section with Ωw-fraction weights, are roughly 200 years for all deep water masses except for water last ventilated south of the Antarctic divergence, which is about twice as old. The uncertainties in the section-mean profiles of the Ωw fractions and their mean ages are ~50% and ~20%, respectively. The Ωw fractions have vertically diffuse overlapping patterns suggesting significant diapycnal mixing, consistent with century-scale mean ages. We quantify the seasonal cycle of ventilation and find that in both hemispheres peak ventilation occurs during late winter and early spring, but Northern Hemisphere ventilated deep waters have a more pronounced seasonal cycle with nearly zero summertime ventilation.

1. Introduction

The times and locations since water in the ocean interior was last ventilated are of first-order importance for understanding and predicting the ocean’s uptake of heat and carbon and the distributions of trace elements that are biologically important in the euphotic zone. A complete characterization of ventilation is provided by the ocean’s boundary propagator, which is a fundamental tracer-independent integral representation of the ocean’s surface-to-interior transport: the boundary propagator of a given water parcel is simply the joint distribution of the times and places since the fluid elements of the parcel were last in the mixed layer (e.g., Hall et al., 2007; Holzer & Hall, 2000; Primeau & Holzer, 2006). Measurements of tracers along hydrographic lines encode important information on the boundary propagator. Here we use a maximum-entropy approach (e.g., Holzer et al., 2010; Holzer & Primeau, 2010; Khatiwala et al., 2009, 2012) to unlock this information and construct estimates of the boundary propagator, ̂G. We then use ̂G to calculate the water mass fractions that were last ventilated in key surface regions, the mean transit time from these regions, and the fraction that is ventilated each month of the year in these regions.

We focus on data from US North Atlantic GEOTRACES section GA03, which was occupied between 2010 and 2011. Section GA03 traverses the subtropical North Atlantic from the North American continental shelf off Cape Cod to the African continental shelf near 17°N, with a meridional extension from Africa to the Portuguese coast near Lisbon (Figure 1). Section GA03 samples all of the major mode, intermediate, deep, and abyssal water types formed in the North Atlantic and crosses the Deep Western Boundary Current (DWBC), which is an important southward conduit for deep waters newly formed in the subpolar North Atlantic and in arctic seas. Intermediate and bottom waters formed in the Southern Ocean penetrate into the subtropical North Atlantic and are also sampled by section GA03.
In addition to being geographically well positioned for transecting key water masses, the data set from section GA03 is of high quality and exceptionally comprehensive. The GEOTRACES mission is to survey the global oceans for the distribution of a suite of trace elements and isotopes with the objective of understanding the marine biogeochemical and physical processes that control their distributions (Boyle et al., 2015). High quality temperature, salinity, nutrient, oxygen, CFC, and SF6 measurements were also made to provide information on these processes. These are the data we use to constrain the surface origin and age of deep waters, which in turn influence the distribution and cycling of trace elements.

Our study complements that of Jenkins et al. (2015), who estimated the admixture of predefined water types using a modified multiparameter inversion of quasi-stationary tracers. By contrast, here we determine water masses in terms of where water was last ventilated and additionally extract transit-time information by utilizing the constraints provided by transient tracers and radiocarbon. Our approach is more general than typical transit-time distribution (TTD) analyses (e.g., Hall et al., 2002; Waugh et al., 2003, 2004) in that we make no assumptions about the functional form of the TTD and estimate the full boundary propagator, not just the distribution of times since last ventilation anywhere.

We begin by outlining our methodology in section 2 and then discuss the data used for constraining the boundary propagator in section 3. In section 4, we present our results for the water mass fractions last ventilated in key surface regions, the mean transit times from these regions, and the seasonality of ventilation. We contrast our results with the work of Jenkins et al. (2015) in section 5 and present conclusions in section 6.

2. Method

The concentrations of tracers that have no interior sources or sinks other than radioactive decay are set by the ocean’s advective-eddy-diffusive circulation. The interior concentration of the $j$th tracer at location $r$ and time $t$ measured during hydrographic surveys, $\chi_j(r, t)$, is determined by the time history of its mixed-layer (or simply surface) concentrations $\chi_j^S(r_s, t_s)$ for surface points $r_s$ and all past times $t_s \leq t$. Formally, interior and surface concentrations are related through convolution with the ocean’s boundary propagator $\mathcal{G}$, that is,

$$\chi_j(r, t) = \int_{-\infty}^{t} dt_s \int d^2r_s \mathcal{G}(r, t| r_s, t_s) \chi_j^S(r_s, t_s) e^{-\frac{(t-t_s)}{\tau_j}},$$  

where the radioactive $e$-folding time $\tau_j$ is infinite for nonradioactive species. Given knowledge of the surface history $\chi_j^S(r_s, t_s)$ and measured interior concentrations $\chi_j(r, t)$, equation (1) represents a constraint from each tracer species on $\mathcal{G}$. The boundary propagator $\mathcal{G}$ encapsulates complete information on the advective-diffusive transport from the surface (Holzer & Hall, 2000).

Here we deconvolve seven tracer constraints of the form (1) to estimate $\mathcal{G}$. With only seven constraints on the function $\mathcal{G}$, the deconvolution is highly underconstrained. We follow Holzer et al. (2010) and regularize the deconvolution using a maximum-entropy approach, which demands that the information entropy of the boundary propagator relative to measure (or “prior”) $\mu$ be maximized subject to the tracer constraints. We do not formally account for instrumental measurement uncertainty (except where it leads to
inconsistencies near the detection limit—see below) because the uncertainty in $G$ is dominated by the underdeterminedness of the deconvolution (Holzer et al., 2010; Holzer & Primeau, 2010). The maximum-entropy estimate of the boundary propagator is then given by

$$G(r, t| r_s, t-s) = \frac{\mu(r, r_s, t-s)}{Z} \exp \left( -\sum_{j=1}^J \lambda_j(r, t) C_j^T(r, t-s)e^{-t/s} \right),$$

(2)

where $Z$ ensures the normalization $\int_0^t dt |d^2 r_s G(r, t| r_s, t-s)|=1$, and $s$ is the $r_s$-to-$r$ transit time. The $J$ Lagrange multipliers $\lambda_j(r, t)$ are determined by substituting (2) into the $J$ tracer constraints (1) and solving the resulting nonlinear system. The deconvolution is performed independently for every interior bottle location $r$. Surface locations $r_s$ are discretized into a $3^\circ \times 3^\circ$ grid, and transit time $s$ is discretized into a nonuniform grid of years $\tau_n$, each of which is further discretized into 12 months. We used the following year grid: $\tau_n \in \{0, 1, 2, 3, 6, \ldots, 75, 80, 90, 120, 160, 200, 250, 300, 400, \ldots, 1, 000, 1, 200, 1, 500, 2, 000, \ldots, 4, 000, 5, 000, 7, 000, \ldots, 11, 000, 15, 000, 20, 000\}$ years.

In the absence of tracer constraints, (2) simply reduces to $G=\mu$. We therefore assess uncertainty by performing the deconvolution for an ensemble of widely different priors, $\mu$, constructed as described below. If the ensemble variance of a given quantity is small, the estimate of the quantity is insensitive to the choice of $\mu$, and hence robust and driven by the tracer observations.

2.1. Prior Ensemble

We construct the prior, $\mu$, similarly to what was done by Ting and Holzer (2017) for the southern oceans, but here we use a different tiling of the global ocean surface because our focus is on the North Atlantic. We divide the global ocean surface into 14 regions $\Omega_{\ell}$ with $\ell=1, 2, \ldots, 14$ as shown in Appendix A. For every interior location $r$, we estimate the fraction $f_0(r, \ell)$ last ventilated in $\Omega_{\ell}$ and the mean $\Omega_{\ell}$-to-$r$ transit time $\Gamma_0(r, \ell)$. As estimates of $f_0(r, \ell)$ and $\Gamma_0(r, \ell)$, we use their values computed with the steady coarse-resolution data-assimilated ocean model of Primeau et al. (2013). By only using the water mass fractions from large regions and the mean transit time from those regions, our prior captures only gross features of the model circulation such as the noninstant connectivity between different basins and the fact that the ideal mean age increases with depth. To fold as little model information into the prior as possible, we do not use the model’s $G$ but instead approximate the $\Omega_{\ell}$-to-$r$ transit-time dependence of $\mu$ with an inverse Gaussian (IG) $g^{ IG}(\tau; \Gamma, \Delta/\Gamma)$, with mean $\Gamma=\Gamma_0(r, \ell)$ and width-to-mean ratio $\Delta/\Gamma$. Our base-state prior is then constructed as

$$\mu(r, r_s, \tau) = f_0(r, \ell) s \Gamma_0(r, \ell), \Delta/\Gamma M(r| \ell),$$

(3)

where $M(r_s, \Omega_s) = 1$ if surface point $r_s$ lies in $\Omega_s$, and $M = 0$ otherwise.

Following Ting and Holzer (2017), we chose $\Delta/\Gamma = 1$ (e.g., Waugh et al., 2013) for the base-state prior and then perturb to generate an ensemble of priors. The $\Delta$ dependence is perturbed by using $\Delta/\Gamma = 0.5$ and 1.5, which covers the relevant oceanographic range of this shape parameter (e.g., Waugh et al., 2006). The ventilation-location dependence (index $\ell$) is perturbed by replacing $f_0(r, \ell)$ in (3) with $f_0(r, \ell)/Z_\ell$, where the $\beta_\ell$ are random numbers drawn independently from a uniform distribution over the interval [0.1, 1.9], and $Z_\ell = \sum_{\ell=1}^{14} f_0(r, \ell)\beta_\ell$ to ensure normalization. For each of the 3 choices of $\Delta/\Gamma$, we use 5 sets of random $\beta_\ell$ to generate an ensemble of 15 very different priors.

2.2. Uncertainty Estimation

For any given quantity $X$, we assign its uncertainty $\delta X$ as the ensemble standard deviation across the ensemble of inversions with different priors. We acknowledge that this is an ad hoc uncertainty assessment, but experiments with synthetic data (Holzer & Waugh, 2015) have shown that the spread of the deconvolutions due to a wide range of priors is a conservative indicator of robustness.

2.3. Water Mass Fractions Partitioned According to Transit Time and Ventilation Region

From our estimate of the boundary propagator $G$, we calculate the fraction of water that was last ventilated in region $\Omega_s$ less than a time $\tau^*$ ago (i.e., of age $\tau^*$ or younger), which is given by

$$f(r, t| \Omega_s; \tau^*) = \int_0^{\tau^*} dt \int_{\Omega_s} d^2 r_s G(r, t| r_s, t-s).$$

(4)
These fractions obey the normalization

$$\sum_{\Omega_w} f(r, t| \Omega_w; \infty) = 1,$$

where the sum extends over a set of (nonoverlapping) surface regions $\Omega_w$ that cover the global ocean surface $\Omega$. Equation (5) simply states that the water at $(r, t)$ must have been ventilated somewhere in the past (going back infinitely far in time; $t' = \infty$).

The mean age of the water mass fraction last ventilated in $\Omega_w$ is given by

$$\Gamma(r, t| \Omega_w) = \frac{1}{f(r, t| \Omega_w; \infty)} \int_{0}^{\infty} dt' \int_{\Omega_w} d^2r' G(r, t| r', t - t').$$

(6)

The usual ideal mean age, which is the mean transit time since being ventilated anywhere at the global surface $\Omega$, is given by $\Gamma(r, t) \equiv \Gamma(r, t| \Omega)$. Note that $f(r, t| \Omega_w; \infty) = 1$ and that

$$\Gamma(r, t) = \sum_{\Omega_w} f(r, t| \Omega_w; \infty) \Gamma(r, t| \Omega_w),$$

that is, the global ideal mean age is the water mass-fraction-weighted average of the regional mean ages.

### 2.4. Definition of Ventilation Regions $\Omega_w$

While our method provides the water mass fraction last ventilated in each 3° × 3° surface grid box, we integrate over larger regions $\Omega_w$ to reduce uncertainty and to summarize the information contained in the boundary propagator. (Note that the point of the higher 3° × 3° source-point resolution in $\mathcal{G}$ is to capture the effects of spatial gradients in the surface tracer values within the larger $\Omega_w$ regions.) We focus on the eight regions, or “patches,” $\Omega_w$ shown in Figure 2, which tile the global ocean surface without overlaps or gaps.

The northern-most patch $\Omega_{AONG}$ encompasses the Arctic Ocean, Greenland, and Norwegian Seas and is bounded to the south by the Greenland-Iceland-Scotland Ridge. Water last ventilated in $\Omega_{AONG}$ includes Denmark Strait and Iceland-Scotland overflow water (e.g., Dickson & Brown, 1994; Rudels et al., 2002; Saunders, 1990; Swift, 1984). The Labrador-Sea patch $\Omega_{LBIR}$ has been extended into the subpolar North Atlantic to include the Irminger Sea. Waters last ventilated in $\Omega_{LBIR}$ include the traditionally defined Upper and Lower Labrador Sea Waters and Irminger Sea Water (e.g., Lazier, 1973; Pickart et al., 2003; Yashayaev, 2007). The green patch $\Omega_{SPNA}$ in the subpolar North Atlantic ventilates subpolar mode waters and is bounded to the south by the $\sigma = 26.5$ kg m$^{-3}$ contour of the annual-mean surface density. The abutting blue patch $\Omega_{STNA}$ in the subtropical North Atlantic is bounded to its south by $\sigma = 25.5$ kg m$^{-3}$ and includes much of the formation region of subtropical mode waters, also referred to as 18°C Water (e.g., Talley & Raymer, 1982; Worthington, 1959). The Mediterranean is an important source of high-salinity waters to the Atlantic and is assigned its own surface patch $\Omega_{MED}$. The Southern Ocean is divided into two patches: $\Omega_{NAAD}$ lies south of the climatological maximum Ekman divergence (Antarctic Divergence, AAD) and extends to the Antarctic continent, while $\Omega_{SAAD}$ lies to the north of the maximum Ekman divergence with a northern boundary at $\sigma = 26.5$ kg m$^{-3}$. Water ventilated in $\Omega_{NAAD}$ includes Antarctic Intermediate Water (AAW), while water ventilated in $\Omega_{SAAD}$ includes Antarctic Bottom Water (AABW). The rest of the global ocean surface is assigned to the grey patch $\Omega_{OTHER}$. The definition of these patches remains fixed in time.
Our surface patches are defined to capture major water mass formation regions but are not intended to locate the exact source region of any given water type (which would also have seasonal time dependence). For example, the formation region of 18°C Water is known to extend further south and west (Worthington, 1959), although the majority of subtropical mode water is ventilated in Ω_{STNA} as shown below.

3. Hydrographic Data

3.1. Interior Concentrations Along GA03

GEOTRACES section GA03 consists of two segments, a dominantly zonal trans-Atlantic section extending from Cape Cod southeast to the coast of Africa at ~17°N and a dominantly meridional section extending northward from just off the coast of Africa to the coast of Portugal (Figure 1). The section was occupied in two phases; the quasi-meridional section and the eastern end of the trans-Atlantic section in 2010 and the remaining trans-Atlantic section in 2011 (Boyle et al., 2015). We use measurements from these cruises for temperature $T$, salinity $S$, the concentrations of oxygen plus phosphate combined to form $\hat{\text{PO}}_4 = [\text{PO}_4] + [\text{O}_2]/175$, and the concentrations of the transient tracers CFC-11, CFC-12, and SF$_6$. Additionally, we use estimates of prebomb radiocarbon ($^{14}$C) from the gridded GLODAP data base (Key et al., 2004) interpolated (and if necessary extrapolated) to the GA03 bottle locations. The in situ $T$, $S$, and pressure values are used to calculate conservative temperature $\Theta$ and absolute salinity $S_A$ (McDougall & Barker, 2011), which are the temperature and salinity variables used in the deconvolutions.

Figure 3 shows vertical sections of these tracers. Their distributions reflect the various water masses, classically defined in terms of in situ hydrographic properties. We now briefly describe the most prominent features; a more detailed description of the temperature, salinity, oxygen, and nutrient distributions was reported by Jenkins et al. (2015).

Near-surface isotherms extend downward in the northwestern half of the zonal section forming a layer of relatively warm water with a low temperature gradient several hundred meters thick in the upper 500 m. The salinity of this layer is relatively high. This is 18°C Water, a subtropical mode water that forms in this region of the Atlantic by mixed-layer deepening and subduction during winter, and is found throughout the North Atlantic subtropical gyre. Along the meridional section, a similar structure is observed in the upper 250 m reflecting Madeira Mode Water (Hanawa & Talley, 2001), which forms in the eastern subtropical North Atlantic, but in much smaller quantities. The intermediate-depth isotherms deepen toward the Portuguese coast and this feature is accompanied by a salinity maximum. This marks the presence of warm salty Mediterranean Overflow Water (MOW; Candela, 2001) that enters the Atlantic through the Strait of Gibraltar and spreads northward and eastward throughout the North Atlantic as can be seen by the relatively high salinity along the center of the zonal section between about 1,200 and 1,800 m depth. The lower-salinity water in this depth range on the western side of the zonal section is from Labrador Sea Water (LSW) formed by deep convection during winter in the Labrador and Irminger Seas. The lower-salinity water on the eastern side is Antarctic Intermediate Water (AAIW) that forms north of the Antarctic Convergence in the Atlantic and Pacific oceans (Rintoul et al., 2001). In the deeper water along both sections, temperature decreases with depth indicating the presence of colder dense water that originates in the high latitudes of both hemispheres. Salinity also decreases with depth reflecting the increased presence of dense Antarctic Bottom Water (AABW).

The tracer PO$_4$ is a quasi conservative property that is relatively low for waters that form in the North Atlantic and relatively high for waters originating in the Southern Ocean (Broecker et al., 1998). The highest values of PO$_4$ occur in the bottom water along section GA03 indicating the presence of AABW. High values of PO$_4$ between 1,000 and 2,000 m at the eastern side of the trans-Atlantic section indicate the presence of AAIW.

The structure of the CFC and SF$_6$ distributions reveals the most recently ventilated water masses. CFC-11 and CFC-12 have been entering the surface ocean in significant quantities for only the past 60 years, and SF$_6$ for just the past 40 years. Thus, the CFC and SF$_6$ distributions are similar, but the different input histories provide unique temporal information. Highest concentrations are observed in the subtropical mode waters. Beneath the mode waters, high CFC and SF$_6$ concentrations extend from surface to bottom at the western end of the trans-Atlantic section. The highest concentrations lie between about 1,000 and 1,500 m in LSW. Relatively high concentrations also extend down to ~2,000 m at the northern end of the meridional section.
indicating the presence of LSW as well as MOW. LSW is colder and has a lower salinity than MOW. The underlying water is a mixture of overflow waters from the Nordic Seas, namely Denmark Strait Overflow Water (DSOW) and Iceland-Scotland overflow water (ISOW). The tracer concentrations are western intensified because these water masses are transported equatorward in the DWBC. High concentrations extending into the interior are the result of lateral exchange with interior circulation pathways. The concentrations are near zero in the deep interior indicating that deep water carrying CFCs and SF6 from southern sources takes longer than 5–6 decades to reach the subtropical North Atlantic.

The deliberate release of 320 kg of SF6 in the Greenland Sea during 1996 was shown to result in excess SF6 concentrations in DSOW as it winds its way through the Labrador Sea, although in 2003 there was still no detectable excess SF6 signal at the Grand Banks (Tanhua et al., 2005). While the excess SF6 was still clearly detectable in 2012 in water flowing into the Arctic Ocean through Fram Strait (Stóvén et al., 2016), it likely takes ~15 years for the excess SF6 signal entering the North Atlantic through Denmark Strait to reach section GA03, by which time the excess SF6 will have undergone substantial dilution. Indeed, we estimate the fraction of water last ventilated in the Arctic Ocean, Greenland, and Norwegian Seas with transit times to section GA03 of 15 years or less to be only about 3–9% in the DWBC below 3,000 m. Using 129I and CFC-11,

![Figure 3](image-url)

Figure 3. The seven tracer fields used in the ME inversions. The horizontal axis is distance, eastward along section GA03 from the western-most station off Cape Cod to the African shelf at ~17°N, then northward to Portugal. The light-grey vertical line marks to the station closest to Africa. For conservative temperature \(\Theta\), the contour interval below 5°C is 0.5°C. For absolute salinity \(S_a\), the contour interval below 35.5 g/kg is 0.1 g/kg. (In shelf regions where GLODAP data is not available, \(\Delta^{14}C\) was extrapolated onto the GA03 bottle locations.)
Smith et al. (2016) estimated the transit time for DSOW to flow from the Labrador Sea to Line W (the western portion of section GA03) to be 5–6 years. Hence, it is possible that the excess SF$_6$ signal Tanhua et al. (2005) observed in 2003 could just have been reaching section GA03 in 2010, but only in very small quantities. We therefore did not consider excess SF$_6$ for the GEOTRACES data from 2010/2011, other than to screen for an anomalously high SF$_6$-to-CFC-12 ratio for SF$_6$ concentrations near the detection limit as described in the following subsection.

Prebomb $\Delta^{14}$C also provides temporal information, but on a much longer time scale than the CFCs and SF$_6$. Radiocarbon is cosmogenic, enters the surface ocean, and decays with a half-life of 5,730 years. Its distribution resembles a smoothed version of the CFC and SF$_6$ distributions, with highest subsurface concentrations in mode waters, lower concentrations in intermediate waters, western intensification along the zonal section at intermediate and deep depths, and lowest concentrations in deep and bottom water.

### 3.2. Global Surface Concentration Histories

The surface boundary conditions for $\Theta$, $S_N$, $P_O_4$, and $O_2$ were approximated as cyclostationary and prescribed from the World Ocean Atlas 09 (WOA09) monthly-mean surface climatology (Antonov et al., 2010; Garcia et al., 2010; Locarnini et al., 2010). The surface boundary conditions for the CFCs were constructed from the Global Ocean Data Analysis Project (GLODAP, Key et al., 2004) CFC data set as follows. Where GLODAP data was available at surface locations $r_s$, we simply scaled the GLODAP surface concentration $C_A(r_s, t_0)$ with the atmospheric time history $C_A(t)$ (Bullister, 2015) according to

$$C(r_s, t) = C(r_s, t_0) \frac{C_A(t)}{C_A(t_0)},$$

(8)

where $t_0 = 1994$ is the nominal year of the GLODAP data set, and the function $s(r_s, t)$ in (8) is a simple parameterization of time-dependent saturation. Note that the scaling of the surface concentration $C_A$ with the atmospheric time history sidesteps the need to explicitly know the solubility and saturation, and for $s(r_s, t) = 1$, equation (8) implies a saturation that is constant in time. However, surface waters have become more saturated over time because the CFC-11 and CFC-12 atmospheric concentrations have been levelling off and decreasing since the 1990s. Here we assumed that the saturation prior to $t_0$ did not change, i.e., that $s(r_s, t) = 1$ for $t \leq t_0$. For $t > t_0$, we assumed that for locations where the $t_0$ saturation is less than 90%, the saturation increases linearly to 90% by the year 2011. For such locations, we therefore increased $s(r_s, t)$ linearly from unity such that the 2011 saturation at $r_s$ is 90%, where the saturation is calculated by taking the ratio of $C(r_s, t)$ and the equilibrium concentration calculated from $C_A(t)$ and the known solubilities (see below). At other locations, we left the saturations constant in time ($s = 1$).

GLODAP did not estimate CFC concentrations north of about 62°N. In this GLODAP void, we estimated the surface CFC concentrations by calculating the saturated concentrations in equilibrium with the atmosphere during the GLODAP year ($t_0$) using the solubilities from the work of Warner and Weiss (1985). The solubilities were multiplied by prescribed monthly saturations $S_{CFC}$ that were then ramped up over time to 90% by multiplying $S_{CFC}$ with a factor $s(r_s, t)$ as in equation (8). Specifically, for the parts of the Labrador and Irminger Seas not part of the GLODAP domain, we used $S_{CFC} = 0.6$ for January–April and $S_{CFC} = 0.85$ for the rest of the year; for the Greenland and Norwegian Seas (west of ~25°E and south of 80°N), we used $S_{CFC} = 0.75$ for January–April and $S_{CFC} = 0.95$ for the rest of the year; for the Arctic Ocean, we used $S_{CFC} = 0.75$ for the entire year. Our results are not sensitive to these details. We find that changes in water mass fractions inferred below due to different choices of CFC saturations are much smaller than the estimated uncertainties.

Measurement of SF$_6$ in the ocean became routine only in the early 2000s and thus there are far fewer observations for SF$_6$ than for CFCs. We could therefore not apply the procedure used for CFCs. For SF$_6$, the surface time history was constructed from the atmospheric SF$_6$ concentration (Bullister, 2015) using the equilibrium solubility (Bullister et al., 2002) based on the climatological mean surface $T$ and $S$, and a prescribed field of SF$_6$ saturations as detailed in Appendix B. Given the nearly linearly increasing atmospheric time history of SF$_6$, we assumed that the SF$_6$ saturation has no trends, only an annual cycle. We checked sensitivity to a globally uniform SF$_6$ saturation of 100% and again found the results to be insensitive.
To allow for possible inconsistencies between the measured interior bottle values and our approximate surface histories, and to be able to detect such inconsistencies, we solved the system of tracer constraints for the Lagrange multipliers $\lambda_j$ of equation (2) only in a least squares sense. The resulting residuals (differences between propagated and measured values) revealed that there were issues with very low SF$_6$ concentrations near the detection limit where either the ratio of SF$_6$ to CFC-12 exceeded that of the surface history, or where the SF$_6$ concentration was reported as exactly zero. We therefore did not use the SF$_6$ constraint where a value of zero SF$_6$ was reported, or where SF$_6$ fell below 0.012 fmol/kg and the ratio of SF$_6$ to CFC-12 exceeded 1.2 fmol/pmol. With these adjustments, the residuals below 500 m depth were near zero (root-mean-square [rms] residuals nearly 2 orders smaller than the detection limit for CFCs and SF$_6$). In the upper ocean, natural variability in the tracer concentrations associated with eddies may cause consistencies with our climatological surface histories. We therefore expect larger residuals in the upper ocean. However, even in the upper ocean the rms residuals were only about 1.5% or less, except for PO$_4^-$ which has an rms residual of ~5% above 500 m. The higher residuals for PO$_4^-$ underline the approximate nature of PO$_4^-$ conservation and the uncertainty and variability in the stoichiometric P:O ratio.

4. Results

4.1. Water Mass Fractions

4.1.1. $\Omega_{LBIR}$-Fraction Younger Than 4,000, 160, and 39 Years

We begin by examining the water mass fractions $f(r, t; \Omega_w; \tau^*)$ for the eight patches defined in Figure 1a. Because these patches tile the global sea surface without overlap or gaps, the eight water mass fractions account for 100% of the water and our deconvolution preserves the normalization (5) to numerical precision. We consider $\tau^* = \infty$ (approximated as 4,000 years), $\tau^* = 160$ years, and $\tau^* = 39$ years, that is, the water mass fraction regardless of age (time since ventilation), younger than 160 years, and younger than 39 years. While the precise values of $\tau^*$ are somewhat arbitrary, we chose these particular values in part because they lie on our $\tau$ grid and allow for comparison with the study of Holzer et al. (2010) who also used 39 and 160 year thresholds. Furthermore, 39 years is on the order of the time for which SF$_6$ and CFCs have been entering the ocean, while 160 years is a typical time scale of the Atlantic deep circulation, as identified from cumulative transit-time distributions (Holzer et al., 2010) and radiocarbon age below 2,000 m (e.g., Broecker et al., 1991, Figure 7). The fractions younger than $\tau^*$, as estimated with the unperturbed and likely most realistic prior, are plotted in Figures 4 and 5, organized geographically from north to south in terms of $\Omega_w$.

The first row of plots in Figure 4 shows that the fraction of water last ventilated in the Labrador/Irminger patch $\Omega_{LBIR}$ is found primarily between ~1,000 and 2,500 m depth, with the highest fractions of ~30% in the western part of the basin. This is consistent with the $\Omega_{LBIR}$ fraction being dominantly composed of Upper Labrador and Irminger Sea Water (1,000–1,500 m), and Classical Labrador Sea Water (CLSW) between 1,500 and 2,500 m depth. Some $\Omega_{LBIR}$-ventilated water lies below 2,500 m depth, especially in the western part of the basin, which is deeper than the reach of CLSW as defined by in situ hydrographic properties. However, the presence of CFCs below 2,500 m (Figure 3) shows that some of these deep waters are relatively young and our deconvolution identifies them as having been last ventilated in $\Omega_{LBIR}$. This may be a result of CLSW being entrained by, and mixed with, DSOW as it enters the Atlantic and flows downslope through the less dense CLSW in transit to greater depths.

When all ages are considered, $\Omega_{LBIR}$-ventilated water is spread horizontally across the entire section with only weak lateral gradients. When water older than 160 years is excluded (middle plots of Figure 4), the fraction deeper than 2,500 m is reduced and the remaining $\Omega_{LBIR}$ fraction lies primarily in the DWBC, with a maximum of ~27% around 1,500 m depth. There is also a secondary maximum of ~19% in the meridional part of the section in the eastern basin, showing that on century and longer time scales there is significant recirculation of $\Omega_{LBIR}$-ventilated water throughout the basin. For water younger than 39 years, only the core of the DWBC carries a significant $\Omega_{LBIR}$ fraction of ~21%.

Water last ventilated in $\Omega_{AONG}$ (second row of Figure 4) has its peak fractions (40–44%) around 3,000 m depth in the western part of the basin, with comparable fractions also found a few hundred meters higher in the meridional section of the eastern basin. The $\Omega_{AONG}$-ventilated fraction is presumably comprised dominantly of DSOW and ISOW. However, there is also a substantial $\Omega_{AONG}$ fraction above ~2,500 m, which would normally be associated with LSW and not overflow waters. This suggests that some of the water
ventilated north of the Greenland-Iceland-Scotland Ridge lies in the same density class as CLSW, and that mixing of \( \Omega_{\text{AONG}} \)-ventilated water and CLSW occurs as the more dense overflow water cascades downslope through the less dense CLSW.

Removing water older than 160 years reduces the peak \( f(\Omega_{\text{AONG}}; \tau) \) in the western basin to \( \sim 35\% \), and in the eastern basin to \( \sim 24\% \). Removing water older than 39 years leaves only a small fraction (\( \sim 16\% \)) in the DWBC. The suggestion that not all \( \Omega_{\text{AONG}} \)-ventilated water becomes overflow water is supported by the fact that the horizontal distribution of the young \( f(\Omega_{\text{AONG}}) \) fraction around 2,000 m depth is remarkably similar to that of the \( f(\Omega_{\text{LBIR}}) \) fraction. This shows that \( f(\Omega_{\text{LBIR}}) \)-ventilated water and the upper component of the \( f(\Omega_{\text{AONG}}) \) fraction share similar density classes and are dispersed by similar circulation pathways.

The \( f(\Omega_{\text{SPNA}}) \) region ventilates subpolar mode water (third row of Figure 4). The \( f(\Omega_{\text{SPNA}}) \) fraction is largest in the far eastern basin along the Africa-to-Portugal part of GA03, with peak values of \( \sim 65\% \). This pattern is consistent with these waters being swept east and south with the subtropical gyre circulation. Along the eastern meridional transect, the bulk of these mode waters is younger than 39 years. The \( f(\Omega_{\text{SPNA}}) \) fraction has a local maximum (\( \sim 45\% \)) over the mid-Atlantic ridge (MAR), suggesting topographic control on the barotropic part of the thermocline circulation. Over the MAR, roughly a third of the \( f(\Omega_{\text{SPNA}}) \) fraction is between 39 and 160 years old.

Figure 4. The water mass fractions \( f(\Omega_{\text{w}}; \tau) \) last ventilated in the regions \( \Omega_{\text{w}} \) for transit times \( \tau \leq \tau^* \) as obtained with the base-state prior. The \( \Omega_{\text{w}} \) regions are defined in Figure 1.
The bottom row of Figure 4 shows the fraction of water last ventilated in the subtropical region \( \Omega_{STNA} \), which corresponds closely to 18°C Water. The \( \Omega_{STNA} \) fraction lies mostly in the upper thermocline above subpolar mode water and traces out the southward shoaling thermocline. Peak fractions reach as high as \( \sim 89\% \), and the \( \Omega_{STNA} \) fraction is completely dominated by young water; the fraction younger than 39 years captures essentially the entire \( \Omega_{STNA} \)-ventilated water mass.

The distribution of the Mediterranean-ventilated \( \Omega_{MED} \) fraction is plotted in the top row of Figure 5. Most of the \( \Omega_{MED} \) fraction lies between 1,000 and 2,000 m depth. \( \Omega_{MED} \)-Ventilated water can be found throughout the North Atlantic, with relatively weak lateral gradients across the trans-Atlantic transect where fractions lie between 15 and 20%. Though not visible with the 5% contour interval of Figure 5, the \( \Omega_{MED} \) fraction again has a small local maximum over the MAR. The highest fractions of \( \sim 45\% \) occur near the Portuguese coast in the high-salinity Mediterranean outflow plume (Figure 3). In the northern half of the meridional section, there is also a suggestion of a minor surface plume of \( \Omega_{MED} \) water, which might be associated with coastal upwelling.

The fraction of water last ventilated in the \( \Omega_{NAAD} \) patch of the Southern Ocean is plotted in the second row of Figure 5. A plume of \( \Omega_{NAAD} \) water, likely composed mostly of AAIW, spreads across most of the North
Atlantic between roughly 500 and 1,500 m depth. Peak $\Omega_{\text{NAAD}}$ fractions around 38% occur at ~900 m depth near the coast of Africa where these waters penetrate from the south. More than ~75% of the $\Omega_{\text{NAAD}}$ fraction is younger than 160 years, but nearly all of it is older than 39 years as expected for this remotely ventilated water mass.

Table 1

<table>
<thead>
<tr>
<th>$\Omega_w$</th>
<th>$\langle f(\Omega_w) \rangle$ (%)</th>
<th>$\langle f(\Omega_w) \rangle_{1\text{km}}$ (%)</th>
<th>$\langle \tau(\Omega_w) \rangle_f$ (years)</th>
<th>$\langle \tau(\Omega_w) \rangle_{20%}$ (years)</th>
</tr>
</thead>
<tbody>
<tr>
<td>LBIR</td>
<td>16 ± 9</td>
<td>20 ± 11</td>
<td>190 ± 19</td>
<td>120 ± 38</td>
</tr>
<tr>
<td>AONG</td>
<td>24 ± 12</td>
<td>30 ± 15</td>
<td>221 ± 12</td>
<td>195 ± 21</td>
</tr>
<tr>
<td>SPNA</td>
<td>9 ± 6</td>
<td>5 ± 4</td>
<td>100 ± 14</td>
<td>25 ± 5</td>
</tr>
<tr>
<td>STNA</td>
<td>8 ± 3</td>
<td>0.8 ± 0.7</td>
<td>25 ± 6</td>
<td>5 ± 1</td>
</tr>
<tr>
<td>MED</td>
<td>14 ± 4</td>
<td>15 ± 5</td>
<td>179 ± 21</td>
<td>94 ± 22</td>
</tr>
<tr>
<td>NAAD</td>
<td>9 ± 6</td>
<td>9 ± 6</td>
<td>182 ± 28</td>
<td>132 ± 8</td>
</tr>
<tr>
<td>SAAD</td>
<td>15 ± 9</td>
<td>18 ± 11</td>
<td>380 ± 84</td>
<td>403 ± 55</td>
</tr>
<tr>
<td>OTHER</td>
<td>5 ± 1</td>
<td>1.1 ± 0.4</td>
<td></td>
<td></td>
</tr>
<tr>
<td>LBIR + ANOG + SPNA</td>
<td>50 ± 12</td>
<td>56 ± 13</td>
<td></td>
<td></td>
</tr>
<tr>
<td>LBIR + ANOG + SPNA + MED</td>
<td>63 ± 12</td>
<td>71 ± 12</td>
<td></td>
<td></td>
</tr>
<tr>
<td>NAAD + SAAD</td>
<td>24 ± 10</td>
<td>27 ± 12</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Note. Uncertainties are the ensemble standard deviations.

Figure 6. Section-mean profiles (unperturbed state in black ± one ensemble standard deviation in red) of the water mass fraction $f(\Omega_w; \tau^*)$ last ventilated in region $\Omega_w$ for transit times $\tau \leq \tau^*$, as indicated. These profiles correspond to the vertical sections plotted in Figure 4.
Water last ventilated in the $\Omega_{SAAD}$ region of the Southern Ocean arrives as bottom water in the North Atlantic as can be seen in the third row of plots of Figure 5. The $\Omega_{SAAD}$ fraction lies mostly in abyssal waters below 3,000 m but has broad vertical gradients throughout the water column. Peak fractions of $\sim$40% occur at the bottom. In the western part of the basin, isolines sloping upward toward the MAR are consistent with northward flowing bottom water being pressed against topography by the Coriolis deflection. Roughly two thirds of the abyssal $\Omega_{SAAD}$ fraction is older than 160 years (60–70% in the western basin, 67–75% in the eastern basin). Essentially none of the $\Omega_{SAAD}$ fraction is younger than 39 years in the subtropical North Atlantic.

The fraction of water ventilated in the tropical Atlantic and elsewhere in the global ocean ($\Omega_{OTHER}$) is mostly confined to the top few hundred meters of the water column (last row of plots in Figure 5). While the $\Omega_{OTHER}$ region ventilates a negligible fraction of deep water, near the surface it exceeds 95%. Most of the $\Omega_{OTHER}$ fraction is younger than 39 years and was hence ventilated in the nearby tropical Atlantic. Between roughly 200 and 600 m depth in the western part of the basin where $18^\circ$C Water is found, the $\Omega_{OTHER}$ fraction ranges from about 20 to 50% and was likely ventilated in the western subtropical North Atlantic between the GA03 station line and the density boundary of $\Omega_{STNA}$.

Figures 4 and 5 show that the deep waters are dominantly comprised of waters last ventilated at high latitudes. Table 1 collects the section-mean $\Omega_{water}$-ventilated fractions calculated by interpolating to a uniformly spaced grid and averaging. (The grid was also used for plotting and covers both the trans-Atlantic and meridional parts of section GA03.) The Mediterranean contributes (14 ± 4)%; the Northern Hemisphere high-latitude regions AONG, LBIR, SPNA together contribute (50 ± 12)%; and the Southern Ocean regions NAAD and SAAD together contribute (24 ± 10)%.

4.1.2. Uncertainties in $\Omega_{water}$-Fraction Profiles

It should be firmly kept in mind that the deconvolution for the boundary propagator at our $3^\circ \times 3^\circ$ surface resolution is highly underdetermined. While integrating over the larger $\Omega_{water}$ regions reduces uncertainty, there is still substantial uncertainty in the inferred water mass fractions discussed above. To give an idea of

![Figure 7. Section-mean profiles as in Figure 6 but corresponding to the vertical sections plotted in Figure 5.](image-url)
the magnitude of these uncertainties, we plot in Figures 6 and 7 the section-mean profiles of the water masses shown in Figures 4 and 5, together with ±one ensemble standard deviation (red lines) of these mean profiles across our ensemble of deconvolutions.

The standard deviations tend to be proportional to the corresponding $\Omega_w$ fractions and represent very roughly 50% of the base-state fractions, with the exception of $\Omega_{\text{MED}}$-ventilated water, where the uncertainty is roughly 25%. This is true not only for the mean profiles but also for the local standard deviations (not shown). Thus, while the absolute fractions have substantial uncertainties, the overall patterns tend to be robust regardless of the prior used. Nevertheless, given these substantial uncertainties one should refrain from overinterpreting smaller-scale features.

4.2. Regional and Global Mean Ventilation Ages
4.2.1. Regional Mean Age

Figure 8 shows the regional mean ages $\Gamma(r, t(\Omega_w))$ calculated from equation (6) for all ventilation regions $\Omega_w$ except $\Omega_{\text{OTHER}}$. Note that mean age is independent of the magnitude of the fraction itself. A large mean age of $\Omega_w$-ventilated water can be the mean age of just a few molecules or of a substantial $\Omega_w$ fraction. To guide the eye in associating the mean-age distribution with the corresponding $\Omega_w$ fraction, we have calculated the contours of the $\Omega_w$ fraction that contain $x$ percent of the section-integrated $\Omega_w$ fraction and overlaid these contours for $x = 20, 50, and 80%$.

The bands of large mean age within the thermocline for $\Omega_{\text{BIR}}$ and $\Omega_{\text{ONG}}$ occur where the corresponding water mass fractions are at most a few percent. While it is possible that the tiny fraction of these waters is very old in the thermocline, this fraction and its mean age are highly uncertain (the uncertainty of the regional mean ages is quantified below), and it is entirely possible that this feature is a spurious artifact.

Below about 1,000 m depth, the regional mean ages have broadly similar patterns, with the exception of the very old $\Omega_{\text{SAAD}}$-ventilated water. The similarity of these patterns shows that below the thermocline the first-order control on the regional ventilation age is advective-diffusive transport regardless of where the water was last ventilated at the surface. This is consistent with the fact that below ~1,000 m and outside the relatively young DWBC, the regional mean ages exceed a century, which is enough time to erase the memory of precisely where deep water was last ventilated. The $\Omega_{\text{SAAD}}$ regional ventilation age has a pattern below ~3,000 m that is similar to that of the other water masses, but with larger amplitude reflecting the slow abyssal transport. The pattern above ~3,000 m is different but represents only very small $\Omega_{\text{SAAD}}$ fractions and may therefore not be robust as indicated by the large uncertainty above ~3,000 m (see below). The different water mass fractions sample their regional mean-age fields depending on where the fraction has significant amplitude as indicated by the percentile contours.

The mean ages within the top 20 percentile contour (capturing the highest fractions that integrate to 20% of the section-wide total) can have a significant range for some of the water mass fractions. For example, for $\Omega_{\text{BIR}}$-ventilated water, the minimum mean age of ~50 years lies in the core of the DWBC, while the fraction-weighted spatial average within the 20 percentile contour is 120 ± 38 years. The overflow waters last ventilated in $\Omega_{\text{ONG}}$ have a minimum mean age within the top 20% contour of around 100 years, while their fraction-weighted average within that contour is 195 ± 21 years. For the subpolar and subtropical mode waters (ventilated in $\Omega_{\text{SPNA}}$ and $\Omega_{\text{STNA}}$), the fraction-weighted mean ages within the top 20% contour are 25 ± 5 and 5 ± 1 years, respectively, pointing to pathways and recirculations associated with the wind-driven gyres. Mediterranean
water within its top 20% contour has mean ages ranging from near zero at the outflow to ~170 years, the fraction-weighted average within the contour being 94 ± 22 years. The fraction-weighted mean ventilation ages of \( \Omega_{\text{NAAD}} \)-ventilated and \( \Omega_{\text{SAAD}} \)-ventilated waters (AAIW and AABW) within their top 20% contours are 132 ± 8 and 403 ± 55 years, respectively. The fraction-weighted ventilation ages averaged across the entire section are older because of the contribution of old waters with low fractions but large volumes. Broadly, the section-mean ventilation age of deep waters is around 200 years, except for \( \Omega_{\text{SAAD}} \)-ventilated water (mostly AABW), which is nearly twice as old because of the slow abyssal circulation and the remoteness of the source region. The section-wide averages are indicated in the plot titles of Figure 8 and collected in Table 1 together with the 20 percentile averaged regional mean ages.

### 4.2.2. Uncertainty Estimates of Regional Mean Age

The ideal mean age regardless of where last ventilation occurred is shown in Figure 10 together with its local ensemble standard deviation. As expected, thermocline waters have mean ages less than a few decades, and the mean age increases with depth until it reaches several hundred years in abyssal waters. The abyssal water in the western part of the basin is roughly 100 years younger than in the eastern part, which reflects the relatively rapid flushing with the DWBC in the western basin and the slow return flow in the eastern basin. The standard deviation is largest in a thin band at the base of the thermocline, the precise location of which (based on the tracer observations) is sensitive to the choice of prior. Elsewhere, the standard deviation in the mean age is roughly proportional to the mean age and increases with depth.

### 4.4. Seasonality of Ventilation

Our deconvolution formally provides the fraction of water last ventilated in each \( 3^\circ \times 3^\circ \) surface grid box, during every month of the year going back in time. Here we calculate the fraction of \( \Omega_w \)-ventilated water that was last ventilated during each month of the year but integrated over all past years. Figure 11 shows the percentage of the section-mean \( \Omega_w \) fraction last ventilated during each month of the year for our five Northern Hemisphere (NH) ventilation regions and separately for the two Southern Hemisphere (SH) regions (NAAD and SAAD). Because these percentages must sum to 100% over the 12 months of the year for each ensemble member, their seasonal cycles are relatively well constrained with small ensemble standard deviations.

The NH-ventilated water masses are primarily ventilated during late winter to early spring, roughly January–April, consistent with many studies of other ventilation metrics in the North Atlantic (e.g., de Boyer Montégut et al., 2004; Forget et al., 2011; Maze & Marshall, 2011). The seasonal cycle of ventilation is most pronounced for the deepwater fractions (\( \Omega_{\text{AABW}}, \Omega_{\text{LBIR}}, \text{and } \Omega_{\text{MED}} \)) for which virtually no ventilation occurs from June to October. This is consistent with deep convection that shuts off during summer playing a key role in ventilating these water masses. The seasonal cycle of North Atlantic subpolar mode water (ventilated in

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**Figure 9.** Section-mean profiles (base state in black, ± one ensemble standard deviation in red) of the regional mean ages of water less than 4,000 years old that was last ventilated in the regions indicated. The mean age in the title is the fraction-weighted spatial average \( \Gamma (\Omega_w) \) across the entire section as in Figure 8.
ensemble standard deviation. (bottom) Corresponding Figure 10. (top) The ideal mean age of water less than 4,000 years old regardless of origin as obtained with the base-state prior. (bottom) Corresponding ensemble standard deviation.

5. Contrast With Optimum Multiparameter Analysis

It is interesting to contrast our approach and results with the recent water mass analysis of the same hydrographic section by Jenkins et al. (2015). These authors did not consider the transient tracers (CFCs, SF6) or steady state radiocarbon, but used the five approximately steady state tracers T, S, silicic acid, PO4, and N8, the latter being a quasi-conservative combination of nitrate and phosphate. Jenkins et al. (2015) devised a modified optimum multiparameter analysis (MOMPA), where the modification was to limit the number of possible end-members by dividing the water column into predefined layers bounded by density surfaces. Specifically, they used a bottom layer (σ27 > 45.85 kg m⁻³), deep layer (σ25 > 34.68 kg m⁻³), intermediate layer (σ0 > 27.2 kg m⁻³), and a thermocline layer (26.2 < σ0 ≤ 27.2 kg m⁻³), the latter being further partitioned into thin layers with a width of 0.01 kg m⁻³ in density. In their approach, the tracer concentrations of a given hydrographic bottle is then considered to be a linear combination of the concentrations of a small number of end-members ("water types") predefined for the layer containing the bottle. The fraction that each end-member contributes to a given bottle is then typically overdetermined by one or two equations and found through a nonnegative least squares fit. Note that in this approach a given bottle is typically an admixture of only those water types contained within the same layer as the bottle, which implicitly assumes negligible diapycnal mixing between the layers.

The key differences between our approach and MOMPA are that MOMPA solves an overdetermined system of predefined end-members within specified density classes, while in our maximum-entropy approach each 3° × 3° surface grid box is a potential end-member, the deconvolution is highly underdetermined (but regularized by insisting that the information entropy of the local admixture is maximized), and there are no restrictions on the vertical distribution of the Ωw-ventilated water masses. In addition, our use of the transient tracers and 14C allows us to estimate the distribution of transit times from each 3° × 3° surface grid box. Aside from the maximum-entropy ansatz, which requires the boundary propagator to be as spread out as possible relative to the prior by maximizing the information entropy, the maximum-entropy approach makes no assumptions beyond the tracers being conservative or having simple transit-time-dependent loss. By contrast, MOMPA requires subjective choices in the definition of the end-members and the density layers in which these are allowed to be mixed.

Given the methodological differences, one cannot directly compare the MOMPA results with those obtained here. Because MOMPA estimates the fractions of end-members either already present in the density layer or proximal to it, the MOMPA fractions of most water types often exceed 85%, while our Ωw fractions are typically much smaller and capture the integrated effect of mixing with other water types during transit from the surface. Nevertheless, because our surface patches include the key formation regions of many of the in situ water types used by MOMPA, we briefly contrast MOMPA and maximum-entropy fractions.
Figure 11. The percentage of the section-mean fraction of water last ventilated during each month of the year as estimated with the base-state prior for the ventilation regions indicated. Summed over all 12 months of the year, these percentages add to 100%. The error bars indicate the ensemble mean standard deviation.

Our $\Omega_{SAAD}$ fraction is very similar to the MOMPA AABW fraction both in pattern and magnitude. This suggests that the SAAD surface properties closely match the AABW water type defined by Jenkins et al. (2015) and that the $\Omega_{SAAD}$ fraction above the corresponding MOMPA density bound for AABW is small. Our $\Omega_{SAAD}$ fraction has a similar horizontal distribution as MOMPA AAIW, but the $\Omega_{NAAD}$ fraction’s peak amplitude is only $(38 \pm 20)\%$ compared to $\sim65\%$ for MOMPA. The cores of the $\Omega_{NAAD}$ and AAIW fractions both lie between roughly 500 and 1,200 m depth, although the $\Omega_{NAAD}$ fraction is vertically more diffuse compared to MOMPA AAIW. The $\Omega_{NAAD}$-ventilated water likely includes some Upper Circumpolar Deep Water (UCDW), which is denser than AAIW. MOMPA defines AAIW and UCDW for the same density layer, but the fractions of the two water types are vertically well separated with UCDW lying immediately below AAIW and extending to the layer’s lower density horizon at $\sim2,000$ m depth. The peak fraction of MOMPA UCDW is $\sim70\%$. This is much larger than the roughly $15\%$ $\Omega_{NAAD}$ fraction at the same depth, which means that the UCDW water type is not well represented in the NAAD surface waters that define the $\Omega_{NAAD}$ fraction. While the MOMPA-derived AAIW and UCDW fractions have similar horizontal distributions, the sum of the AAIW and UCDW fractions does not well approximate our $\Omega_{NAAD}$ fraction because of the very different peak amplitudes and vertical profiles, the MOMPA profile being dominated by UCDW. These differences are methodological and likely due to the fact that the AAIW and UCDW water types defined by Jenkins et al. (2015) are based on section A07 at $\sim5^\circ$S, where water from the Southern Ocean has already been mixed during its transit with water ventilated elsewhere. That such mixing likely played a big role in setting the end-member properties defined by Jenkins et al. (2015) is underscored by an $\Omega_{NAAD}$ mean ventilation age of $\sim200$ years (Figure 8).

Our $\Omega_{AONG}$ fraction has a pattern broadly similar to the overflow waters defined by Jenkins et al. (2015), but with a peak fraction that is $(46 \pm 21)\%$ compared to a peak MOMAP fraction of $\sim88\%$. The $\Omega_{AONG}$ fraction is also vertically more diffuse and at the western boundary extends into the density class of Labrador Sea Water as discussed above. The $\Omega_{LBIR}$ fraction has a pattern broadly similar to the sum of the Upper Labrador Sea, Classical Labrador Sea, and Irminger Sea fractions of the MOMPA analysis, but again $\Omega_{LBIR}$-ventilated water is vertically more diffuse with peak fractions of $(30 \pm 18)\%$ compared to $\sim90\%$ for MOMPA. The $\Omega_{MED}$ fraction has a pattern similar to the MOMPA MOW fraction but is vertically much more diffuse with a maximum in the trans-Atlantic section that is roughly 500 m deeper than for MOMPA MOW. The maximum $\Omega_{MED}$ fraction near Portugal is $(46 \pm 14)\%$, significantly less that the MOMPA MOW maximum of $\sim95\%$.

It is worth noting that the two methods have very different types of uncertainty. For the maximum-entropy method, the uncertainty is substantial and fundamentally due to the fact that only a handful of tracer constraints leaves the determination of the boundary propagator highly underconstrained. Jenkins et al. (2015) do not enforce that their end-members account for 100% of the water in each layer, and they consider the 10–20% deviation from a 100% sum to be a measure of uncertainty. However, as Jenkins et al. (2015) acknowledge, there is also unquantified uncertainty inherent in the choice of the density layers, the composition of the end-members, and the weighting of the different tracer constraints.

6. Conclusions

We have used measurements of $T$, $S$, PO$_4$, CFC-11, CFC-12, and SF$_6$ from section GA03 together with GLO-DAP prebomb $^{14}$C to constrain the transport from the mixed layer to the measurement locations. A maximum-entropy deconvolution method was employed to estimate the boundary propagator with a $3^\circ \times 3^\circ$ resolution in surface origin. We used this estimate to compute the transit-time partitioned water mass...
fractions last ventilated in surface regions $\Omega_w$ and the mean transit time from each of these regions. Our key findings are as follows:

1. Our analysis confirms that most of the North Atlantic was last ventilated at high latitudes, with a substantial fraction of deep water also being ventilated in the Mediterranean. Along section GA03, $(50 \pm 12)\%$ of the water was ventilated in the northern high-latitude regions AONG, LBIR, and SPNA, $(24 \pm 10)\%$ in the southern high-latitude regions NAAD and SAAD, and $(14 \pm 4)\%$ in the Mediterranean. Below 1,000 m depth, these fractions are even higher at $(71 \pm 12)\%$ for AONG, LBIR, SPNA, plus MED-ventilated water and $(27 \pm 12)\%$ for NAAD plus SAAD-ventilated water.

2. We estimated the regional mean ventilation age, that is, the mean time since water was last ventilated in a given $\Omega_w$ region. The spatial pattern of these regional mean ages is largely decoupled from the patterns of the $\Omega_w$ fractions themselves. Below the thermocline and outside the DWBC, mean ventilation ages tend to exceed a century, which is sufficiently old for water to have lost memory of where it was last ventilated. Consequently, the spatial pattern of the regional mean ages is set by the advective-diffusive circulation to a pattern that is broadly similar for all $\Omega_w$. For deepwater masses, the $\Omega_w$-fraction weighted mean ages across the section are roughly 200 years, except for SAAD-ventilated bottom waters for which this average age is $\sim$400 years. Mode waters are younger, with weighted section means of $\sim$100 years for subpolar and $\sim$25 years for subtropical mode waters.

3. The ideal mean age of water regardless of where it was last ventilated has a relatively simple, large-scale structure along section GA03. Vertical mean-age gradients are larger in the eastern part of the basin than in the more rapidly ventilated western part. The mean age of abyssal water ranges from 200 to 400 years, older in the eastern part of the basin, where old bottom waters from the Southern Ocean contribute roughly a third of the water. Overflow waters last ventilated in the Arctic, Norwegian, and Greenland Seas make roughly a 40% contribution to the deep western basin where $\sim$60% of these waters are younger than 160 years.

4. The water masses last ventilated in each of our $\Omega_w$ regions broadly coincide with the traditional water masses that are known to form in these regions. However, because the traditional water masses are defined in terms of their in situ properties, the correspondence is at best qualitative and there are some striking differences. For example, we find that some of the water last ventilated in the Labrador/Irminger region can be found well below where one would expect Classical Labrador Sea Water. Compared to the water mass fractions estimated by Jenkins et al. (2015) using MOMPA, the fractions last ventilated in the corresponding $\Omega_w$ patches are vertically much more diffuse. Given that the deep mean ages are typically older than a century and represent the mean of distributions with long tails skewed toward even older ages, one can expect significant diapycnal transport. For example, a typical background vertical diffusivity of $10^{-5} \text{ m}^2 \text{s}^{-1}$ and a mean age of 200 years correspond already to a vertical diffusive scale of $\sim$250 m. The sharp density bounds of the MOMPA impose unrealistic vertical structure on the composition of deep water masses.

5. We estimated the seasonal cycle of ventilation in terms of the percentage of the section-integrated $\Omega_w$ fraction that was last ventilated during each month of the year. We found that water mass fractions last ventilated in key NH formation regions (AONG, LBIR, and MED) have a large seasonal cycle, with peak ventilation in late winter to early spring and nearly no ventilation in summer. By contrast, the ventilation in the NAAD and SAAD regions of the SH has a less pronounced seasonal cycle. Peak ventilation in the SH again occurs in late winter and early spring, but there is also significant ventilation during summer. While these qualitative features are not surprising and consistent with other studies of ventilation and mixed-layer depth, it is remarkable that our estimates of the boundary propagator from the tracer measurements alone allow reasonably well-constrained quantitative estimates of the seasonal cycle of ventilation.

6. Because our analysis is based on an underdetermined deconvolution, there are substantial uncertainties in all estimated quantities. Uncertainties were estimated in an ad hoc fashion by repeatedly performing the deconvolution for an ensemble of strongly perturbed priors. We find that the local ensemble standard deviation of the $\Omega_w$ fractions has a pattern broadly similar to the fractions themselves, pointing to robustly estimated large-scale patterns. However, the amplitude of the patterns carry up to 50% (relative) uncertainty. The uncertainties in the section-mean profiles of the regional mean ages are about 20% or less except for possibly spurious features where the corresponding water mass fractions are small and the uncertainties can exceed 50%.
The relatively simple deconvolution procedure used here can be applied to any hydrographic section for which transient tracers are available. Although a more rigorous bound on the uncertainties would be desirable, Ting and Holzer (2017) demonstrated for the Southern Ocean that even with the simple approach used here, one can identify decadal changes in ventilation from one occupation to the next. In future work, we plan to extend these analyses to other sections and to repeat hydrographies as they become available.

**Appendix A: Surface Patches for Construction of the Base-State Prior**

Figure A1 shows the 14 surface patches used in the construction of the prior guess at the boundary propagator, $\mu$. These patches are mostly based on the $\sigma = 23.5, 24.5, 25.5,$ and $26.5 \text{ kg m}^{-3}$ isolines of the annual-mean climatological surface density, with the following exceptions: The line separating the two Southern Ocean patches is based on the maximum Ekman divergence, the southern boundary of the northern-most patch (Norwegian, Greenland, and Arctic Seas) was chosen to follow the Greenland-Iceland-Scotland Ridge, and the Labrador-Sea patch was extended by hand to include the Irminger Sea. The ocean other than the Atlantic and the two Southern Ocean regions is assigned to a single patch.

**Appendix B: Specified SF$_6$ Saturation**

The SF$_6$ saturation was defined to be patch-wise uniform with a seasonal cycle as shown in Figure B1. This saturation field was constructed by (i) examining the available SF$_6$ measurements and (ii) by using the observed CFC-12 saturations for the early to mid 1990s. The rationale for (ii) is as follows. Currently and for the past 25 years, the atmospheric SF$_6$ concentration has been increasing nearly linearly and the surface saturation is assumed to be in quasi steady state. This was also the case for CFC-12 from 1970 to 1992. Thus, the saturation of CFC-12 from the mid 1980s until 1992 would be expected to be very similar to saturations of SF$_6$ observed from the early 2000s until the present, assuming the physical processes (such as gas exchange rate and winter convection) have not changed.

For the Arctic Ocean there are very few SF$_6$ observations. We examined observations from the U.S. and German GEOTRACES Arctic cruises in 2015 made by one of us (Smethie) and the mode of the surface saturation was 85%. These cruises extended across the entire Arctic Ocean providing the value we assign to the Arctic. (The range and mode for CFC-12 saturation for cruises reported in the GLODAPv2 (Olsen et al., 2016) data base taken in 1991 and 1994 were 60–90% and approximately 80%).

For the Greenland/Norwegian Sea during 1982–1992, the CFC-12 saturation ranged from 80 to 105% with a mode of 85–90%. Although there are SF$_6$ measurements in the Greenland/Norwegian Sea from the late 1990s until present, the cruises in the GLODAPv2 data base for this period were focused on following a large purposeful injection of SF$_6$ into the Greenland Gyre in 1996 (Messias et al., 2008) and the near-surface water was supersaturated by up to 200% at some locations, so these data were not used to estimate saturation. For the Greenland/Norwegian Sea, we chose a winter value of 85% based on the Arctic SF$_6$ measurements and assigned a value of 90% to the rest of the year based on the 1982–1992 CFC-12 measurements.

For the Labrador and Irminger Seas, two cruises with SF$_6$ measurements taken in 2003 (Met 59, Steinfeldt et al., 2009) and 2012 (MBL2012, https://cchdo.ucsd.edu/cruise/18MF20120601) were examined. Both cruises were nonwinter cruises and had SF$_6$ saturations ranging from 80 to 110%. Deep convection occurs in winter months and saturations in the underlying winter mixed layer ranged from 75 to 80%.
SF₆ surface saturations were thus assigned a value of 75% for January–April and 90% for the remainder of the year.

An analysis by Tanhua et al. (2008) of North Atlantic CFC-12 surface saturation data reported in the GLODAP and CARINA data bases reveals an average CFC-12 surface saturation of 86% prior to 1990. These authors also demonstrate that the SF₆ saturation after 1990 should be about 86% because its atmospheric concentration has been increasing at about the same rate as the CFC-12 atmospheric concentration prior to 1990 as discussed above. Based on this, we take the SF₆ saturation to be 85% for the January–April winter period for both the subpolar and subtropical North Atlantic patches. For the July–October period, we assign SF₆ saturations of 95%, 100%, and 105% to the subpolar, northern subtropical, and southern subtropical patches based on SF₆ measurements from the 2012 occupations of the A20 (https://cchdo.ucsd.edu/cruise/33AT20120419) and A22 (https://cchdo.ucsd.edu/cruise/33AT20120324) Repeat Hydrography lines. For the transition period between winter and summer, we assign 95% to the subpolar and subtropical patches.

For the Mediterranean Sea, there is one cruise with SF₆ measurements (Meteor 84/3, Tanhua et al., 2013). The SF₆ saturations range from 78 to 103%, with an average and mode of 90%, which we assign to be the year-round saturation for the Mediterranean.

For the tropical Atlantic, we assume 100% saturation year round. For the subtropical South Atlantic, we prescribe the same seasonal pattern as for the North Atlantic but with seasons reversed for the Southern Hemisphere.

The SF₆ saturations in the NAAD and SAAD Southern Ocean patches are based on CFC-12 and SF₆ observations in the Atlantic and Pacific sectors collected on CLIVAR Repeat Hydrographic cruises (http://cchdo.ucsd.edu). The ocean outside the Atlantic and the two Southern Ocean regions has a wide range of surface saturations but contributes little to the water in the North Atlantic. We simply assign an average saturation of 90% for \( \Omega_{\text{OTHER}} \).

**References**


**Figure B1.** The patch-wise uniform SF₆ saturation used and its seasonal variation.


