Isohaline Salinity Budget of the North Atlantic Salinity Maximum

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(Manuscript received 21 August 2014, in final form 4 November 2014)

ABSTRACT
In this study, the salinity budget of the North Atlantic subtropical salinity maximum region for control volumes bounded by isohaline surfaces is analyzed. The authors provide closed budgets based on output from a high-resolution numerical simulation and partial budgets based on analyses of observational climatologies of hydrography and surface fluxes. With this choice of control volume, advection is eliminated from the instantaneous volume-integrated salt budget, and time-mean advection is eliminated from the budget evaluated from time-averaged data. In this way, the role of irreversible mixing processes in the maintenance and variability of the salinity maximum are more readily revealed. By carrying out the analysis with both near-instantaneous and time-averaged model output, the role of mesoscale eddies in stirring and mixing for this water mass is determined. This study finds that the small-scale mixing acting on enhanced gradients generated by the mesoscale eddies is approximately equal to that acting on the large-scale gradients estimated from climatological-mean conditions. The isohaline salinity budget can be related to water mass transformation rates associated with surface forcing and mixing processes in a straightforward manner. The authors find that the surface net evaporation in the North Atlantic salinity maximum region accounts for a transformation of 7 Sverdrups (Sv; 1 Sv = 10^6 m^3 s^-1) of water across the 37-psu isohaline outcrop into the salinity maximum in the simulation, whereas the estimate based on climatological observations is 9 to 10 Sv.

1. Introduction
In each of the subtropical gyres of the global ocean there exists a distinct surface salinity maximum, with closed isohaline contours. They are an expression of the coupling of the ocean and atmosphere through the hydrologic cycle. The surface salinity maxima are located in the vicinity of regional extrema in net evaporation but are not exactly collocated with them. These surface features connect to equatorward- and westward-extending subsurface salinity maxima (typical core depths of 100–150 m) sometimes identified in the literature as Subtropical Underwater (STUW) (O’Connor et al. 2005). Interest in the role of these features in the global climate system has been elevated because of several factors. The surface waters subducted in these regions make their way to the equatorial thermocline via the shallow subtropical cells, as well as the higher-latitude North Atlantic via the Atlantic meridional overturning circulation (AMOC) (Qu et al. 2013), and thus have the potential to contribute to large-scale climate variability. With indications of an intensification of the hydrologic cycle in a warming world (Held and Soden 2006), the signature of changes in surface salinity provide a useful additional gauge on the magnitude of these global-scale changes in the climate system. Recent studies have shown that the surface salinity in the salinity maximum regions is increasing (Durack et al. 2012). Studies with coupled climate models (Latif et al. 2000) have suggested that low-latitude salinity increases in the Atlantic in response to global warming may act to stabilize the AMOC. The launch of the Soil Moisture

* The National Center for Atmospheric Research is sponsored by the National Science Foundation.

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DOI: 10.1175/JPO-D-14-0172.1

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The Budget of the Salinity Maximum

The Salinity Processes in the Upper Ocean Regional Study (SPURS; http://spurs.jpl.nasa.gov) field experiment was designed as a multiscale investigation of the processes that give rise to the North Atlantic subtropical salinity maximum. A series of cruises, moorings, autonomous instrumentation, and an augmentation to the extant global observing system were deployed in the region surrounding a central mooring at 25°N, 38°W. The experimental design was developed around the concept of attempting to close the surface mixed layer salinity budget. Several previous investigations have used observations collected during SPURS, hydrographic climatologies, and ocean models to examine aspects of the near-surface salinity budget (Qu et al. 2011; Busecke et al. 2014; Gordon and Giulivi 2014). In this study, we take a complimentary approach by formulating the salinity budget for a control volume bounded by an isohaline surface. This is a natural choice for the salinity maximum region because the isohaline surface defines a closed volume that, by definition, bounds the feature of interest. As discussed further below, this choice of control volume has the unique property that advection is eliminated from the budget, allowing one to focus on mixing processes. This property has been exploited in previous investigations of the heat budget of the tropical warm pools (Niiler and Stevenson 1982; Toole et al. 2004; Enfield and Lee 2005; Song and Yu 2013) as well as salinity budgets of estuary systems (Walin 1977; MacCready 2011). McWilliams et al. (1996) investigated the heat and salinity budgets of the upper ocean in a coarse-resolution global ocean model using this methodology, and Schneider and Bhatt (2000) inferred mixing rates required from salinity and temperature variance budgets evaluated within this framework using the World Ocean Atlas (Levitus 1982) hydrographic climatology.

In this study, we carry out an investigation of the isohaline salinity budget of an eddy-resolving ocean simulation, with particular focus on the role of mesoscale eddies in the maintenance of the salinity maximum region. We compare selected terms in the resulting budget to calculations based on recent hydrographic and air–sea flux climatologies. As we show below, the isohaline salinity budget can be related to the volume budget of the same control volume through a water mass transformation function (Walin 1977, 1982; Marshall et al. 1999). This allows us to associate the volume flux of water into and out of the salinity maximum region with particular mixing processes.

2. Methods

a. Model description and experimental design

The model used in this study is the ocean component of the Community Earth System Model (Smith et al. 2010). The model is configured on a global tripole grid (Murray 1996) with zonal resolution of 0.1° and meridional resolution of 0.1° cos(latitude), yielding a grid spacing of approximately 10 km in the North Atlantic salinity maximum region. There are 62 levels in the vertical, with 10-m resolution above 160 m, increasing to 250 m in the abyss. The model uses a partial cell representation of topography (Adcroft et al. 1997) with depths derived from ETOPO2v2 (http://www.ngdc.noaa.gov/mgg/fliers/06m gg01.html). Surface mixed layer and interior diapycnal mixing processes are parameterized using the K-profile parameterization (KPP) (Large et al. 1994). The closure for lateral mixing by subgrid-scale processes is a biharmonic operator with hyperviscosity and diffusivity scaled by the cube of the local grid spacing. The equatorial value of hyperviscosity is $2.7 \times 10^{10} \text{m}^4 \text{s}^{-1}$, while that for hyperdiffusivity is $3 \times 10^9 \text{m}^4 \text{s}^{-1}$.

Surface forcing is provided by the Co-ordinated Ocean–Ice Reference Experiments (CORE) normal-year forcing (Large and Yeager 2004, 2009). This forcing dataset is derived from a combination of atmospheric reanalysis and remote sensing products, with adjustments made in each of the prescribed variables to bring the globally integrated heat and freshwater budgets into balance when fluxes are computed using observed sea surface temperature. The “normal-year” version of this forcing is derived from the full interannual product for the period 1948–2009 and provides one exactly repeating synthetic year of forcing data at 6-h intervals. During model integration, heat, freshwater, and momentum fluxes are updated every 6 h using the model’s predicted sea surface temperature and velocity and the CORE prescribed surface meteorological state, precipitation, and downwelling radiation. Note that in this experiment the freshwater forcing is imposed as a virtual salt flux.

The model is initialized at rest with the temperature and salinity distributions interpolated to the model grid from the World Ocean Circulation Experiment (WOCE) global hydrographic climatology (Gouretski and Koltermann 2004). The model is integrated forward...
for 15 yr saving monthly statistics of the solution, then extended for an additional 5 yr with mean state variables and all fluxes required to recover exact term balances of heat and salt, saved every 5 days. The analysis presented in this paper is based on this final 5-yr period.

b. Datasets

In addition to the datasets used for initializing and forcing the model, several observational products are used in the analysis. For evaluation of the simulated interior salinity distribution and mixed layer depth we use the Monthly Isopycnal/Mixed-Layer Ocean Climatology (MIMOC) (Schmidtko et al. 2013) and World Ocean Atlas (WOA09) (Antonov et al. 2009) climatologies. The surface salinity climatology (LEGOS) of Reverdin et al. (2007) is used as well. The same datasets are also used to define the surface area and volume of the salinity maximum region to calculate terms in the salinity budget. Several previous studies, for example, Toole et al. (2004), have decomposed the tendency term on the left-hand side of (3) into contributions due to changes in mean salinity within the volume and contributions due to changes in the volume contained within the isosurface:

\[
\int \int \int_{V(S>S_0)} \frac{\partial S}{\partial t} \, dv = \int \int _{\partial V(S=S_0)} \mathbf{u} S \cdot da + \int \int _{\partial V(S=S_0)} A_S \mathbf{V} \cdot (\mathbf{S} \cdot da, \quad (3)
\]

Several previous studies, for example, Toole et al. (2004), have decomposed the tendency term on the left-hand side of (3) into contributions due to changes in mean salinity within the volume and contributions due to changes in the volume contained within the isosurface:

\[
\int \int \int_{V(S>S_0)} \frac{\partial S}{\partial t} \, dv = V \frac{\partial \langle S \rangle}{\partial t} + (\langle S \rangle - S_0) \frac{dV}{dt},
\]

where \( \langle S \rangle \) is the volume-mean salinity. For most of the analysis presented here, we keep the tendency term in the form of the left-hand side of (3) but provide an example of the decomposition on the right-hand side of (5) using climatological observations in section 3b.

The derivation to this point has considered the state variables to be instantaneous values. In any practical application of the method, we are more typically beginning with time-averaged or spatially smoothed data. For the evaluation of the simulation, we are

\[
\frac{\partial S}{\partial t} = -\mathbf{V} \cdot (\mathbf{u} S) + \frac{\partial}{\partial z} \left[ \kappa \left( \frac{\partial S}{\partial z} - \Gamma \right) \right] - A_S \nabla^2 h S, 
\]

with the surface boundary condition

\[
\frac{\partial S}{\partial z}_{z=0} = S_{ref}(E - P),
\]

where \( \kappa \) is the vertical diffusivity, \( \kappa \Gamma \) is the counter-gradient flux as predicted by KPP, \( A_S \) is the horizontal hyperdiffusivity, \( E \) is evaporation, \( P \) is precipitation, and \( S_{ref} = 34.7 \text{ psu} \) is a constant reference salinity used to convert freshwater flux to virtual salinity flux. No-flux boundary conditions are imposed on all solid boundaries.

Consider a control volume \( V \) bounded by an isohaline surface \( \partial V \), where \( S = S_0 \). Integrating (1) over \( V \) yields

\[
\int \int \int_{V(S>S_0)} \frac{\partial S}{\partial t} \, dv = -\int \int _{\partial V(S=S_0)} \mathbf{u} S \cdot da + \int \int _{\partial V(S=S_0)} A_S \mathbf{V} \cdot (\mathbf{S} \cdot da,
\]

where \( dv \) is an element of volume, \( da \) refers to an element of the isohaline surface within the ocean interior, \( dA \) refers to an element of the sea surface within the area bounded by the outercrop of \( \partial V \), and \( \mathbf{k} \) is the unit vector in the vertical direction. Figure 1 shows the control volumes defined by \( S = 37.0 \text{ psu} \) for an annual mean and a 5-day mean from the simulation. Note that the control volume can be multiply connected. The subducted tongue of high-salinity water is apparent extending to the southwest below the surface mixed layer, reaching depths of approximately 150 m at its westernmost extent.

Noting that since \( S \) is a constant on the surface \( \partial V \), the first term on the right-hand side of (3) vanishes by mass conservation under the Boussinesq approximation:

\[
\int \int _{\partial V(S=S_0)} \mathbf{u} \cdot da = S_0 \int \int _{\partial V(S=S_0)} \mathbf{u} \cdot da \quad = S_0 \int \int _{\partial V(S=S_0)} \mathbf{V} \cdot \mathbf{u} \, dv = 0. \quad (4)
\]
beginning with 5-day averages. The observational climatologies being used have been spatially smoothed through objective mapping procedures and time averaged into climatological monthly means. Beginning with the time-averaged form of (1), and defining the bounding surface by the average salinity $\bar{S} = S_0$, the budget integrated over the isohaline control volume becomes
\[
\iint_{V(S=S_0)} \frac{\partial S}{\partial t} \, dv = - \iiint_{\partial V(S=S_0)} \overline{uS} \cdot da + \iint_{\partial V(S=S_0)} \left( \kappa \left( \frac{\partial S}{\partial z} - \Gamma \right) \right) \mathbf{k} \cdot da + \iint_{A(S>S_0)} S_{ref}(E-P) \, dA - \iint_{\partial V(S=S_0)} A_j (V^2S) \cdot da
\]

TEND = EDADV + VDIFF + SRFLX + HDIFF.

Now the first term on the right-hand side does not vanish because the integrand involves a covariance rather than a simple product. Using a Reynolds decomposition, only the part of the advection associated with the time-mean velocity vanishes, leaving only the eddy advection

\[
\iint_{\partial V(S=S_0)} \overline{uS} \cdot da = - \iint_{\partial V(S=S_0)} \overline{uS} \cdot da + \iint_{\partial V(S=S_0)} \overline{uS}' \cdot da
\]

\[
= \iint_{\partial V(S=S_0)} \overline{uS}' \cdot da.
\]

The budget (6) is evaluated in two ways. First, each integral is evaluated with the fluxes in each integrand on the right-hand side and \(\overline{S}\) defining the control volume surface taken from the 5-day mean model output. The individual 5-day integral budgets are then averaged term by term into climatological monthly means. Second, each integral is evaluated with the fluxes in each integrand on the right-hand side and \(\overline{S}\) derived from climatological monthly means of the model output. In addition to the conventional eddy advection contribution to the budget, there are potential rectification effects [defined as the difference for each term of (6) between the estimates based on 5-day and climatological monthly-mean output] even for the terms with integrands that are linear in the state variables or fluxes. They arise from the covariance of the fluxes across the bounding surface and the position and orientation of the surface itself (McWilliams et al. 1996).

The seminal study of Walin (1982) showed how the budget of a scalar in a control volume bounded by isosurfaces of the same scalar is related to its volume budget. Most work utilizing this methodology has focused on partially open control volumes of temperature or density, with the main result being an estimate of the water mass transformation rate or subduction rate estimated from heat or buoyancy fluxes across the sea surface. The same methodology can be applied to the salinity maximum region to estimate rates of water mass transformation into or out of the isohaline control volume, but since the volume is completely bounded by the isohaline surface, there is no net subduction out of the volume. Conservation of volume for the salinity maximum region bounded by salinity \(S = S_0\) may be written (Marshall et al. 1999; Viúdez 2000) as

\[
\frac{\partial V(S_0, t)}{\partial t} = - \frac{\partial}{\partial S_0} \left[ \iint_{\partial V(S=S_0)} \overline{uS}' \cdot da + \iint_{\partial V(S=S_0)} \left( \kappa \left( \frac{\partial S}{\partial z} - \Gamma \right) \right) \mathbf{k} \cdot da + \iint_{A(S>S_0)} S_{ref}(E-P) \, dA - \iint_{\partial V(S=S_0)} A_j (V^2S) \cdot da \right].
\]

The change in volume of the salinity maximum region is determined by the divergence in salinity space of the nonadvective fluxes of salinity across the boundary. Note that the term within brackets on the right-hand side of (8) is just the same as the right-hand side of (6). We will evaluate (8) using both the 5-day mean and climatological monthly-mean fluxes in the integrands.

### 3. Results

#### a. Climatological description of the model solution

The climatological annual-mean distribution of surface salinity in the \(S_{\text{max}}\) region from the simulation is shown in Fig. 2. In this analysis, we will use \(S = 37.0\) as the outermost isohaline surface considered in the budgets. As shown in Fig. 2, the simulated outcrop area of this surface is considerably smaller than for any of the climatological observational estimates, with the northern boundary displaced approximately 2° latitude to the south of the WOCE climatology used to initialize the model. The highest salinity within the \(S_{\text{max}}\) region in the simulation is at the lower end of the observational climatologies, comparable to WOA09, but about 0.1 psu lower than the MIMOC or LEGOS climatologies. North of the peak salinity maximum, the simulated salinity is approximately 0.2 psu fresher than the WOCE...
climatology, the freshest of the four climatologies shown. The slight expansion of the simulated $S_{\text{max}}$ region to the west is insufficient to compensate for the reduction of area in the north. The south and east extents of the simulated $S_{\text{max}}$ region remain in quite good agreement with observations. The time evolution of the simulated annual-mean outcrop area and volume bounded by $S = 37.0$ are shown in Fig. 3. Both drop rapidly over the first 8 to 10 yr of the simulation, then begin to rebound. The outcrop area is relatively stationary over the analysis period years 16–20, while the volume has some positive trend and interannual variability.

The mean annual cycle of the outcrop surface area is shown in Fig. 4. Each of the observational estimates and the model has a maximum surface outcrop area in August–October and a minimum area in March–May. The amplitude of the annual cycle is 20%–25% of the annual-mean value. The outcrop area migrates in a southwest–northeast direction through the year (not shown). The north and east edges of the outcrop region reach their most poleward and eastward extent in boreal late summer and fall, while the southwest edge is displaced most equatorward in winter. The expansion to the north and east exceeds the retraction in the southwest during the fall so that the maximum outcrop area occurs at that time. There is little movement of the outcrop in the southeast quadrant through the year. It is seen that the outcrop area bounded by the 37.0-psu contour in the simulation is roughly equivalent to that bounded by the 37.1-psu contour on the WOA09 climatology and still somewhat smaller than that of the other climatologies. However, the northern edge of the simulated outcrop area remains to the south of even the WOA09 climatology (Fig. 2).

The seasonal variation of the net surface heat and freshwater fluxes along the central longitude of the $S_{\text{max}}$ region are shown in Fig. 5. The heat flux has a dominant annual cycle at all latitudes with net cooling from October to March and heating from April to September. The maximum heating is reached earlier in the year to the north of the $S_{\text{max}}$ than to the south. On the other hand, the freshwater flux has a more complex semiannual variation for latitudes between 15° and 30°N. The net flux is negative (evaporative) all year long, with minimum magnitude in April and October. Variations in evaporation and precipitation are each dominated by the annual period but are phase shifted with respect to one another. The fall minimum magnitude in net freshwater loss results from the maximum in precipitation at that time, while the spring minimum is the result of a minimum in evaporation.
The annual cycle of mixed layer depth at the longitude of the center of the $S_{\text{max}}$ region from the simulation is compared to the estimate of Schmidtko et al. (2013) in Fig. 6. The mixed layer is deepest in February–March throughout the latitude range of the $S_{\text{max}}$. Minimum mixed layer depths occur in May–July within and to the north of $S_{\text{max}}$ and in August–September to the south. There is a band along the southern flank of the $S_{\text{max}}$ region centered on 17°N with a relative maximum in summer mixed layer depth. The minimum mixed layer depth south of 15°N is coincident with the maximum in precipitation during the northern extreme of the ITCZ annual migration. Overall, the model captures the phase of the annual cycle of mixed layer depth and the summer season minimum depths quite well, but the winter mixed layer depth is too deep, especially north of 20°N.

b. Isohaline salinity budget

The budgets for the simulation have been evaluated for isohaline surfaces from $S = 37.0$ to $S = 37.4$. The isohaline salinity budget of the simulation evaluated from 5-day model output for the control volume bounded by the $S = 37.0$ isohaline surface is shown in Fig. 7a. The surface forcing is the only term acting to increase the salinity within the volume. It has a dominant semiannual variation, reflecting the variation in $P - E$ shown in Fig. 5b. The variation of the catchment area of the surface outcrop, which has a predominantly annual period (Fig. 4), has a smaller effect. The net flux of salt out of the volume through mixing processes varies predominantly at the annual period, with the maximum loss of salt occurring in December–January and minimum in April–May. The annual variation in mixing is governed primarily through the vertical mixing term with the largest loss during the time of year when the mixed layer is rapidly deepening. The salt loss associated with horizontal hyperdiffusion is relatively uniform throughout the year, with magnitude comparable to that of vertical mixing during the boreal spring season when vertical mixing is weakest. There is a small eddy advection contribution to the budget associated with motions on time scales less than 5 days. The mean advection is indistinguishable from zero by construction, as described in the appendix. The net salt tendency is positive through the summer months, coincident with the maximum in net evaporation. The winter maximum in net evaporation is more than offset by the higher vertical mixing in that season so that the salt tendency is negative from boreal fall through spring.
The salt budget for the same $S = 37.0$ isohaline bounded control volume evaluated using climatological monthly-mean model output is shown in Fig. 7b. The surface forcing term is nearly identical to that evaluated using high-frequency output, so that there is little rectification due to covariation of the outcrop surface area and surface fluxes on nonseasonal time scales. Similarly, the net flux through the isohaline surface in the interior is very similar in Figs. 7a and 7b, as it must be for equilibrium conditions and the very similar surface fluxes. On the other hand, the contribution of each of the components of the salt flux through the isohaline surface does change. The rectification for each term is shown in Fig. 8. The contribution of eddy advection increases substantially when the budgets are computed from monthly-mean output. Its magnitude becomes comparable to that due to vertical mixing through much of the year. In contrast, the contribution of horizontal hyperdiffusion becomes vanishingly small. There is also a reduction in the flux associated with the vertical mixing of magnitude comparable to the reduction in horizontal mixing.

The general seasonal variation of the terms and relative magnitudes are very similar for all isohaline surfaces ($S = 37.0$ to $S = 37.4$). The absolute magnitudes of the fluxes decrease as the threshold salinity increases, as will...
be discussed further below in connection with the water mass transformation analysis.

While there is insufficient data to compute the full salt budget from observations, we can estimate the surface forcing and salinity tendency terms and infer the flux due to mixing across the isohaline surface as a residual. These are shown in Fig. 9 for the control volume bounded by the $S = 37.1$ isohaline surface. Estimates are shown for the tendency terms evaluated using the MIMOC and WOA09 hydrographic climatologies, along with the OAFlux plus GPCP or CORE surface forcing data. We chose to show results for slightly higher threshold salinity than was presented for the model analysis above to more closely match the volume and surface area of the control volume in the simulation as shown in Fig. 4. The seasonal variation of the surface forcing term closely matches that from the simulation, though the annual-mean magnitude is larger, even for the surface outcrop defined by a higher salinity. The estimates for the MIMOC climatology are larger, in correspondence to the larger outcrop area compared to WOA09. There is little difference between the estimates based on OAFlux versus CORE surface flux climatologies. The similarity is to be expected because many of the same observational datasets go into both the CORE and OAFlux flux analyses. The salinity tendency evaluated from the MIMOC climatology varies with a dominant annual period, with salinity increasing in spring and summer similar to the simulation, though the peak is shifted earlier in the year by about 1 month. The tendency term estimate based on the WOA09 climatology is noisier but still suggests a minimum at the turn of the year and maximum in spring. The contribution of the two terms on the right-hand side of (5), attributed to changes in the mean salinity within the control volume and the volume itself, respectively, are generally comparable in magnitude, though there is rather poor agreement between the estimates from the two hydrographic climatologies. The net salt flux through the interior isohaline surface, computed as a residual from (6), has a maximum magnitude in December–January for all observational estimates, as in the simulation. However, the shift in the tendency term produces a more apparent semiannual variation for the MIMOC-based estimates than was evident in the simulation. The positive interior flux in April for the WOA09-based estimates would require unphysical upgradient mixing, an indication that this combination of climatologies is less reliable.

c. Isohaline volume budget

As discussed in section 2c, there is a simple connection between the salinity and volume budgets for an isohaline bounded control volume, allowing us to reinterpret the results from section 3b in terms of water mass transformation rates. In Fig. 10a, we show the value of the climatological annual mean of each term in the salinity budget based on 5-day model output for isohaline values between $S = 37.0$ and $S = 37.4$. From (8), the water mass transformation rate associated with each term is given by the negative slope of the corresponding curve and is shown in Fig. 10b. For the climatological annual mean, the left-hand side of the equation vanishes, so the individual contributions to the net flux across the isohaline surface must balance. Surface forcing is responsible for a transformation of about 7 Sverdrups (Sv; $1\text{ Sv} = 10^6\text{ m}^3\text{ s}^{-1}$) of fresher water to the higher-salinity water of the salinity maximum across the $S = 37.05$ isohaline contour. This is balanced by a 5-Sv transformation of higher-salinity $S_{\text{max}}$ water to surrounding lower-salinity water by vertical mixing, 2 Sv by horizontal hyperdiffusion, and 1 Sv by eddy advection due to high-frequency (time scales less than 5 days) motions. The transformation by surface fluxes decreases to about 5 Sv
at $S = 37.35$. The decrease in surface transformation is balanced by 0.5–1-Sv decrease in each of the mixing terms. Considering the budget based on climatological monthly output (Fig. 10c), the surface transformation is nearly the same as for high-frequency output except in the highest-salinity class. We consider this last estimate to be less reliable because there are months when the volume of water with $S > 37.4$ becomes very small in the climatological mean, making the evaluation of the integrals less robust. At $S = 37.05$, the transformation due to vertical mixing and that due to eddy advection are approximately equal at 3.5 Sv.

The transformation due to vertical mixing decreases slightly toward higher salinity, while that due to eddy advection is more uniform across salinity thresholds.

The annual-mean isohaline salinity budget surface forcing term and water mass transformation due to surface forcing derived from the observational hydrographic and freshwater flux climatologies are shown in Fig. 11. These observations suggest a higher transformation rate associated with surface forcing of about 10 Sv. For equilibrium conditions, the implied transformation rates due to interior fluxes are just the negative of the surface transformation shown. There is little difference in the estimates based on the CORE and OAFlux surface flux climatologies. The estimates based on the WOA09 have similar transformation rates as the MIMOC climatology at the fresher isohaline outcrops, but decrease more rapidly for the saltier outcrops near the center of the $S_{\text{max}}$ region.

4. Discussion and conclusions

The contrast between the estimates of the salinity budget using near-instantaneous (5-day mean) versus climatological monthly-mean model output provides several insights. The absence of mean advection in the estimate based on 5-day mean output makes it apparent that mesoscale eddies do not in and of themselves mix. The stirring they provide acts to increase the gradients of salinity locally so that yet smaller-scale processes may act to provide irreversible mixing. In the present simulation, the eddy stirring is accommodated in approximately equal parts by horizontal hyperdiffusion and an enhancement of vertical mixing over the level that would act on the climatological salinity distribution. The difference in the horizontal hyperdiffusion term is of special interest. Based on analyses of property budgets for Eulerian control volumes (e.g., Grodsky et al. 2014), it may appear that such parameterizations have
little impact on the large-scale properties of the solution in eddy-resolving simulations, though there are dissenting opinions (Roberts and Marshall 1998). The analysis here clearly illustrates the importance of this term in determining water mass properties. When evaluated on the time-varying isohaline surface, the small-scale variations of this term are coherent with the convoluted structure of the surface itself, with a non-trivial contribution to the mean budget. When evaluated from an Eulerian average of the model solution, then integrated over the mean position of the surface, there is little net contribution to the flux. This insight strongly motivates us to seek more physically based closures for the mesoscale-resolving regime (e.g., Fox-Kemper and Menenmenlis 2008; S. Bachman and B. Fox-Kemper, unpublished manuscript).

The small-scale mixing that is modulated by eddy stirring is approximately equal in magnitude to that provided by vertical diffusion alone acting on the climatological salinity distribution. This part of the mixing, both the enhancement of vertical mixing and the horizontal hyperdiffusion, is rather steady throughout the year. In contrast, the part of the vertical mixing acting on the large-scale salinity distribution has a strong seasonal cycle. Though we have not considered the variance budget directly, in rough terms, we can say that approximately half of the cascade of variance between the large-scale and microscale is accomplished via a pathway passing through the mesoscale eddy field.

A number of previous studies have used the isoscalar budget framework to estimate bulk diapycnal or diathermal diffusivities (Niiler and Stevenson 1982; Schneider and Bhatt 2000) by dividing the isopycnal or isothermal surface-averaged net interior flux obtained as a residual (as in Fig. 9) by the averaged normal gradient of that scalar. We have avoided that step here for several reasons. First, as is apparent from the full term balance, the net diffusive flux may be attributed to a mix of processes acting both laterally and vertically. It is not clear how a bulk scalar diffusivity should be interpreted in this case. Further, for the volumes under consideration, much of the vertical mixing is occurring within the surface mixed layer and is highly seasonal, again making interpretation in terms of diffusivity less obvious.

The simulation-based estimates of the budget and transformation rate presented here are subject to a number of caveats. As we have shown, the simulated salinity is biased low in the $S_{\text{max}}$ region. The lower salinity reduces the surface outcrop area and hence the transformation rates. A full diagnosis of the source of this bias would require sensitivity experiments to isolate the contribution of various processes. However, at least two factors are likely to be contributing. First, the form of the virtual salt flux boundary condition (2) will tend to underestimate the effect of net evaporation on the surface salinity in this region by about 10% because the local salinity is about 10% higher than the constant $S_{\text{ref}}$ value used in the model. This is not likely to be the dominant factor, however, since the evaporation is stronger to the south of the $S_{\text{max}}$ while the bias is strongest to the north. A more likely candidate is the excessively deep winter mixed layer depth, particularly on the north side of $S_{\text{max}}$ (Fig. 6). As shown in Fig. 7, vertical mixing in winter contributes the largest negative tendency to the salinity budget of the control volume. It is interesting that the MIMOC climatology, which emphasizes the more recent, Argo era observations, shows the largest surface area outcrop and highest transformation rates. This is consistent with the observed increases in surface salinity in this region in recent decades (Durack et al. 2012). Another source of concern with respect to the robustness of the simulation-based budgets is the nontrivial role of the parameterized horizontal mixing. As indicated above, the biharmonic subgrid-scale closure employed by the model lacks a physical justification and may be distorting the partition of fluxes across the isohaline surface.

The formulation of the budget for the salinity maximum region in an isohaline framework provides a view of the processes acting to control the salinity distribution that is complementary to other approaches that have been applied to this region (Qu et al. 2013; Busecke et al. 2014; Gordon and Giulivi 2014). It has the advantage of providing a quantification of the role of scale interactions across all processes, not just advection. On the other hand, it is obtained using an integral over a surface that is not easily decomposed into subcomponents. In contrast, a budget over an Eulerian control volume defined by an observational array or set of model grid cells may be more easily subdivided into surfaces to assign relative importance to zonal, meridional, and vertical fluxes for example. Nevertheless, this study is consistent with these earlier studies in establishing that mesoscale eddies play a central role in determining the salt budget of the North Atlantic salinity maximum region.

As shown by Schneider and Bhatt (2000), the isoscalar budget framework can be applied not just to the scalar itself, but also to any differentiable function of that scalar. Thus, the same methodology may be useful in constraining budgets of salinity variance, providing a further link between microscale, mesoscale, and large-scale observations made during SPURS. We plan to pursue this approach in future work.

**Acknowledgments.** Support for this project was provided by NASA Award NNX10AC16G. The simulations described in this study were completed with resources...
provided by the Computational and Information System Laboratory at NCAR. We thank David Bailey for assistance in running the model integrations and data processing. We thank two anonymous reviewers for their suggestions for improving the manuscript.

APPENDIX

Evaluation of Flux Integrals

The numerical method used to compute the integral expressions in (6) is a refinement of that described in McWilliams et al. (1996). Rather than computing a surface integral of the normal component of the flux across the isohaline surface, each term is evaluated as a volume integral of the divergence of the flux over the time-varying control volume. The control volume may be multiply connected. The cells contained completely or partially within the selected salinity surface are identified for each sample by searching within a predefined range of latitude, longitude, and depth surrounding the feature of interest. A first guess of the fractional volume of those cells intersected by the isohaline surface is determined by three-dimensional linear interpolation. The contribution to the integral is simply the grid cell value of the flux divergence multiplied by the full cell volume for those cells completely contained within the isohaline surface or the fractional cell volume for those cells on the boundary. An iterative adjustment procedure is used to ensure that expression (4) is satisfied. We evaluate (4) using the first-guess cell volumes. If the volume integral is positive, the fractional volumes of those boundary cells that contribute positively to the integral are reduced slightly, while those that contribute negatively are increased slightly. The opposite adjustments are applied if the volume integral is negative. This is repeated until the estimate of the mean advection term is no more than 1% of the magnitude of the other terms in the budget. The process typically converges in $O(10)$ iterations. The fractional cell volumes so obtained are used to compute all the volume integrals in the salinity budget. The adjustments to the boundary cell volumes have little effect on the diffusive terms. Occasionally, a small blob of high-salinity water will pass through the bounding box over which the initial search for the control volume is done and there is an irreducible net-mean advective flux. These cases are infrequent enough that there is negligible influence on the climatological-mean monthly budgets shown in Fig. 7.

REFERENCES


