Importance of Circulation Changes to Atlantic Heat Storage Rates on Seasonal and Interannual Time Scales

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ABSTRACT

Ocean heat budgets and transports are diagnosed to elucidate the importance of general circulation changes to Atlantic Ocean heat storage rates. The focus is on low- and midlatitude regions and on seasonal and interannual time scales. An estimate of the ocean state over 1993–2004, produced by a coarse-resolution general circulation model fit to observations via the method of Lagrange multipliers, is used. Meridional heat transports are first decomposed into contributions from time-mean and time-variable velocity and temperature and second from zonally symmetric baroclinic (overturning, including Ekman) and zonally asymmetric (gyre and other spatially correlated) circulations. Heat storage rates are then ascribed to ocean–atmosphere heat exchanges, diffusive mixing, and advective processes related to the various components of the meridional heat transport. Results show that seasonal heat storage changes generally represent a local response to surface heat inputs, but seasonal advective changes are also important near the equator. Interannual heat storage rate anomalies are mostly due to advection in tropical regions, whereas both surface heat fluxes and advection contribute at higher latitudes. Low-latitude advection can be primarily attributed to zonally symmetric baroclinic circulations, but temperature variations and zonally asymmetric flows can contribute elsewhere. A relationship between interannual heat storage rates in the equatorial Atlantic's top 100 m and meridional heat transport associated with the zonally symmetric baroclinic flow is observed; however, due in part to the role of shallow advective processes at these latitudes, any direct relationship between sea surface temperature variability and heat transport changes associated with intermediate or deep meridional overturning circulations is not clear.

1. Introduction

The thermal physics of the ocean is of great interest in climate studies. This is mainly due to the ocean’s role in redistributing heat in the earth system. Meridional transport of heat by ocean currents contributes as much as one-third to the global poleward heat transport around 14°–24°N (Trenberth and Caron 2001; Ganachaud and Wunsch 2003). Divergences in ocean heat transport either can be stored locally within the ocean or exchanged with the atmosphere via radiative and turbulent fluxes. Sea surface temperatures mediate such air–sea energy fluxes, thus exerting a direct influence on climate, for example in the Atlantic sector (Marshall et al. 2001). Understanding and attributing the storage and transport of heat in the ocean is thus integral to fuller comprehension of climate system dynamics.

Thermal variability in the ocean can be due to direct forcing by the atmosphere, truly coupled ocean–atmosphere interactions, or intrinsic oceanic dynamics. Sea surface temperature variability has been studied extensively and related to various statistical indices of climate, as reviewed by Deser et al. (2010): interannual-to-decadal temperature variability in the extratropical ocean is related primarily to modes of atmospheric circulation (e.g., North Atlantic Oscillation); year-to-year tropical ocean temperature changes correspond to coupled atmosphere–ocean interactions (e.g., El Niño–Southern Oscillation); and on multidecadal time scales, surface temperature variability is supposed to reflect intrinsic modes of the ocean's thermohaline circulation (e.g., Atlantic multidecadal oscillation). However, correlation with these statistical indices does not amount to causal explanation, and the dynamical relationship of sea surface temperature variability patterns to the storage and transport of heat in the ocean interior is not clear.

One time and repeat hydrographic occupations have allowed for the estimation of ocean meridional heat transports at a number of latitudes (MacDonald 1998; Bryden and Imawaki, 2001; Ganachaud and Wunsch 2003), and
some estimates of the relative importance of various components of the circulation to the mean transports have been made from observations (Talley 1999, 2003) and models (Boccaletti et al. 2005). Shallow, intermediate, and deep overturning components can contribute importantly to the total meridional heat transport, for example, at 24°N in the Atlantic (Talley 1999, 2003). Regarding the temporal variability of heat transports, direct measurements are sparse (excepting 26°N in the Atlantic; see Johns et al. 2011), but coupled models have been used to estimate the seasonal-to-decadal variability and diagnose the underlying causes (e.g., Dong and Sutton 2001; Jayne and Marotzke 2001; Shaffrey and Sutton 2004). Variable Ekman fluxes generally make important contributions to seasonal and interannual heat transports, but other mechanisms (e.g., temperature variability, deeper overturning circulation changes) can also play a role depending on region and time scale (Dong and Sutton 2001; Jayne and Marotzke 2001). Subsequent attribution of ocean heat storage rates in terms of the factors known to underlie heat transport changes in large part has not been made, however.

Links between ocean circulation and heat transport variability, sea surface temperature changes, and ocean heat storage must be understood before the potential impacts of overturning circulation changes on Atlantic sector climate can be anticipated (Lozier 2010). The goal of this study is to elucidate how ocean heat storage rates are tied to changes in ocean circulation across various regions and time scales. To this end, variability in meridional heat transport and heat storage rates are examined across the Atlantic using an ocean state estimate produced by the Estimating the Circulation and Climate of the Ocean (ECCO) consortium (Wunsch et al. 2009). The ECCO estimates—generated by a general circulation model along with its adjoint and constrained to a multiplicity of observations through least squares—represent model–observation syntheses. The ECCO approach reduces some of the shortcomings characterizing model-only or observation-only approaches (e.g., ill-known errors in the former case and data paucity in the latter) and thus, in a statistical sense, ECCO solutions arguably represent “best estimates” of the ocean state. The remainder of this paper is structured as follows: in section 2 the characteristics of both the state estimate and numerical model are discussed; for context, meridional heat transports and transport decompositions are presented in section 3; heat storage rate budgets are analyzed in section 4 to determine the importance of heat transport to heat storage rates as a function of region, depth, and time scale; and finally, a summary and discussion are offered in section 5.

2. General circulation model framework

An ocean state estimate produced by the ECCO group forms the basis of this study. The ECCO state estimates represent evolved, optimized solutions of the Massachusetts Institute of Technology general circulation model (MITgcm) (Marshall et al. 1997a,b; Adcroft et al. 2010) fit to a majority of the World Ocean Circulation Experiment (WOCE) era hydrography, Argo floats, satellite altimetry, and most other available ocean data in a weighted nonlinear least squares sense. Using the model’s adjoint (Heimbach et al. 2005; Heimbach 2008), a cost function weighing the model–data misfit is minimized via iterative adjustment of a high-dimensional control vector containing initial conditions and boundary conditions (Wunsch and Heimbach 2007; Wunsch et al. 2009). The optimization achieves both consonance with observations (to within specified uncertainties) as well as dynamical consistency with all model equations. The solution’s consistency allows for quantitative analysis of property budgets that close exactly. The various ECCO products, optimization procedures, as well as the data and data weights (i.e., uncertainties) used in the optimization are described in much more detail in several publications. Interested readers are referred to Wunsch and Heimbach (2007) as well as Wunsch et al. (2009) and the references therein. Refer to Adcroft et al. (2010) for a detailed description of parameterizations and numerics in the circulation model proper.

A brief description suffices for the present purposes. The solution is defined on a (nearly) global range, from 80°S to 80°N, with a 1° horizontal grid resolution and 23 vertical layers ranging in thickness from 10 m at the sea surface to 500 m in the abyssal ocean. The model uses the Boussinesq approximation with an implicit linear free surface and a virtual salt flux boundary condition. Hydrographic initial conditions are first taken from a blend of Gouretski and Koltermann (2004) and Boyer and Levitus (1998) climatologies; six-hourly products from the National Centers for Environmental Prediction (NCEP)–National Center for Atmospheric Research (NCAR) re-analysis described by Kalnay et al. (1996) are used over the duration 1992–2004 as “first guess” atmospheric forcing fields; these initial conditions and boundary conditions constitute the ECCO control vector and are iteratively adjusted and optimized. Mixing fields generally have Laplacian, Gent–McWilliams (Gent and McWilliams 1990; Griffies 1998), Redi (1982), and nonlocal K-profile parameterization (Large et al. 1994) components. Surface heat fluxes comprise longwave, shortwave, sensible, and latent contributions; shortwave radiation is allowed to
TABLE 1. Notation used in the text.

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Description</th>
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<th>Description</th>
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<tbody>
<tr>
<td>( T )</td>
<td>Meridional heat transport</td>
<td>( (\cdot)_{m} )</td>
<td>Zonally symmetric barotropic component</td>
</tr>
<tr>
<td>( D )</td>
<td>Heat transport divergence</td>
<td>( (\cdot)_{e} )</td>
<td>Zonally symmetric baroclinic component</td>
</tr>
<tr>
<td>( H )</td>
<td>Heat storage rate</td>
<td>( (\cdot)_{h} )</td>
<td>Zonally asymmetric component</td>
</tr>
<tr>
<td>( Q )</td>
<td>Surface heat flux</td>
<td>( (\cdot)_{Q} )</td>
<td>Mean component</td>
</tr>
<tr>
<td>( M )</td>
<td>Diffusive/turbulent mixing</td>
<td>( (\cdot)_{M} )</td>
<td>Anomalous temperature component</td>
</tr>
<tr>
<td>( V )</td>
<td>Variance explained</td>
<td>( (\cdot)_{V} )</td>
<td>Anomalous transport component</td>
</tr>
<tr>
<td>( \sigma )</td>
<td>Magnitude (standard deviation)</td>
<td>( (\cdot)_{\sigma} )</td>
<td>Covariable (eddy) component</td>
</tr>
<tr>
<td>( (\cdot) )</td>
<td>Temporal mean</td>
<td>( (\cdot)_{m} )</td>
<td>Near-surface (0–100 m) integral</td>
</tr>
<tr>
<td>( (\cdot)^{S} )</td>
<td>Mean seasonal cycle</td>
<td>( (\cdot)_{\infty} )</td>
<td>Upper-ocean (0–1000 m) integral</td>
</tr>
<tr>
<td>( (\cdot)^{I} )</td>
<td>Interannual anomaly</td>
<td></td>
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penetrate vertically as Jerlov Type IA water (Jerlov 1968; Paulson and Simpson 1977).

The present analysis makes use of ECCO solution v2.216 (version 2, iteration 216), which covers the years 1992–2004 and was produced at the Massachusetts Institute of Technology and Atmospheric and Environmental Research, Inc. (MIT–AER). This solution is described in detail in Wunsch et al. (2007) in the context of an analysis of decadal sea level trends. Though more recent 1° ECCO state estimates are now available, comparisons to the solution used here reveal similar qualitative results and suggest that contemporary versions are approaching a stable estimate (Wunsch and Heimbach 2009). Given the nonlinearity of the problem, however, one cannot categorically rule out the possibility of qualitatively distinct results in future ECCO estimates.

Monthly model output is the target of investigation here. Because of the observational data paucity and transients evident in the solution during 1992 (cf. Wunsch et al. 2007), the solution is analyzed only over 1993–2004. Our focus is mostly on transport and storage of heat in the subtropical and tropical Atlantic Ocean, though some higher-latitude regions in the North Atlantic are also considered. This study expands upon the work of Wunsch and Heimbach (2006), who computed heat fluxes at 26°N using an earlier ECCO integration.

3. Meridional heat transport

To motivate the heat storage rate budgets that follow, we begin by presenting estimates and decompositions of meridional heat transport in the Atlantic Ocean. The meridional transport of heat \( T \) across a zonal oceanic section can be defined as

\[
T = \int \rho_0 c_p \theta \, dx \, dz \quad \text{[Watts]}. \quad (1)
\]

where \( c_p \) is the specific heat capacity of seawater (J kg\(^{-1}\) °C\(^{-1}\)), \( \rho_0 \) is a reference density (kg m\(^{-3}\)), \( v \) is meridional velocity (m s\(^{-1}\)), and \( \theta \) is potential temperature (°C). (A list of frequently used notation is given in Table 1.) Meridional heat transports computed from the ECCO state are presented in Fig. 1 alongside previously published observational estimates (Table 2).\(^2\) Uncertainties in the literature values are standard errors while uncertainties in the ECCO estimates represent temporal variability. Although quantitatively ECCO values appear to underestimate the magnitude of the observationally derived heat transport at many latitudes, uncertainty ranges are mostly overlapping and the overall latitudinal pattern of the ECCO heat transports is qualitatively similar to the structure of the observation-based estimates.

Given our interest in temporal variability on seasonal and interannual time scales, heat transports are separated into time mean (\( \langle T \rangle \)), mean seasonal cycle (\( T^{S} \)), and interannual anomaly (\( T^{I} \)) time series at each latitude. The quantity \( T^{S} \) is computed by removing both the time mean and linear trend from \( T \), and then averaging fields corresponding to January, February, etc., into a 12-month time series, while \( T^{I} \) is computed by first subtracting the mean, linear trend, and seasonal cycle from \( T \), and then

\(^2\) Because the ECCO state estimate is constrained by historical data, Fig. 1 should not be taken to represent a comparison to independent data; here, we mean simply to provide a context for ECCO heat transport values.

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Table 2. Heat transport estimates from the literature used in Fig. 1. Latitudinal sections given are nominal; see references for exact section location. Uncertainties are standard errors as provided. Where more than one value was given, mean values are presented here, and uncertainties are propagated standard errors plus the standard deviation of estimates. In the case of Talley (2003), no formal errors were given. Talley (2003) instead gives a general uncertainty of 10%–20% of the transport value, up to 0.2 PW. The upper bound of this range is used here.

<table>
<thead>
<tr>
<th>Section</th>
<th>$T$ (PW)</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>59°N</td>
<td>0.70 ± 0.14</td>
<td>Talley (2003)</td>
</tr>
<tr>
<td>55°N</td>
<td>0.28 ± 0.06</td>
<td>Bacon (1997)</td>
</tr>
<tr>
<td>53°N</td>
<td>0.62 ± 0.12</td>
<td>Talley (2003)</td>
</tr>
<tr>
<td>48°N</td>
<td>0.65 ± 0.25</td>
<td>MacDonald (1998)</td>
</tr>
<tr>
<td>47°N</td>
<td>0.60 ± 0.09</td>
<td>Ganachaud and Wunsch (2003)</td>
</tr>
<tr>
<td>45°N</td>
<td>0.62 ± 0.12</td>
<td>Talley (2003)</td>
</tr>
<tr>
<td>36°N</td>
<td>1.01 ± 0.26</td>
<td>MacDonald (1998)</td>
</tr>
<tr>
<td>24°N</td>
<td>1.27 ± 0.15</td>
<td>Ganachaud and Wunsch (2003)</td>
</tr>
<tr>
<td>11°N</td>
<td>1.39 ± 0.25</td>
<td>MacDonald (1998)</td>
</tr>
<tr>
<td>8°N</td>
<td>0.73 ± 0.15</td>
<td>Talley (2003)</td>
</tr>
<tr>
<td>5°S</td>
<td>1.00 ± 0.55</td>
<td>Ganachaud and Wunsch (2003)</td>
</tr>
<tr>
<td>11°S</td>
<td>0.90 ± 0.40</td>
<td>Ganachaud and Wunsch (2003)</td>
</tr>
<tr>
<td>23°S</td>
<td>0.33 ± 0.25</td>
<td>MacDonald (1998)</td>
</tr>
<tr>
<td>24°S</td>
<td>0.33 ± 0.06</td>
<td>Holfert and Siedler (2001)</td>
</tr>
<tr>
<td>25°S</td>
<td>0.36 ± 0.06</td>
<td>Holfert and Siedler (2001)</td>
</tr>
<tr>
<td>27°S</td>
<td>0.49 ± 0.25</td>
<td>MacDonald (1998)</td>
</tr>
<tr>
<td>30°S</td>
<td>0.35 ± 0.15</td>
<td>Ganachaud and Wunsch (2003)</td>
</tr>
<tr>
<td>32°S</td>
<td>0.23 ± 0.05</td>
<td>Talley (2003)</td>
</tr>
</tbody>
</table>

We next examine what variables (mean and anomalous temperature and velocity) and components of the circulation (zonally symmetric barotropic, zonally symmetric baroclinic, and zonally asymmetric) contribute most to the behavior of $\overline{T}$, $T^\prime$, and $T^I$.

### a. Variable decomposition

Time mean and anomalous components of the meridional velocity and potential temperature fields contribute to heat transport variability. To determine the relative contribution of each, the following decomposition is used:

\[
\overline{T} = \int \rho_0 c_p \overline{\theta'} dx \, dz + \int \rho_0 c_p \overline{\theta} dx \, dz + \int \rho_0 c_p \overline{u'} \overline{\theta'} dx \, dz + \int \rho_0 c_p \overline{u'} \overline{\theta} dx \, dz + \int \rho_0 c_p \overline{v'} \overline{\theta'} dx \, dz + \int \rho_0 c_p \overline{v'} \overline{\theta} dx \, dz, \tag{2}
\]

where the overbar ( ) denotes the local time mean, and the prime (') denotes deviation from the time mean, such that $u' = u - \overline{u}$, and likewise for potential temperature. The right-hand side terms can be interpreted physically as (from left to right) mean transport of mean temperatures ($\overline{T}$), mean transport of anomalous temperatures ($T^\prime$), anomalous transport of mean temperatures ($T^I$), and covariable transport ($T^\prime$).

Decompositions are presented in Fig. 2. Here, $\overline{T}$ is determined primarily by $\overline{T^\prime}$ while $T^I$ makes little contribution (Fig. 2a). On seasonal and interannual scales, the magnitude of $T$ variability (defined as the standard deviation $\sigma$) is comparable to $\sigma(T^I)$ at most latitudes (Figs. 2b–c). Although $\sigma(T^I)$ appears to be the main contributor to $\sigma(T)$, there are several regions where $\sigma(T^I)$ and $\sigma(T^\prime)$ are not negligible (Figs. 2b–c).
The $\sigma$ values in Figs. 2b–c do not contain information regarding signal phase. For a more thorough comparison, let the variance $\mathcal{V}$ of a time series $x(t)$ explained by another time series $\hat{x}(t)$ be defined as (cf. Fukumori et al. 1998)

$$\mathcal{V}(x,\hat{x}) = 1 - \frac{\sigma^2(x - \hat{x})}{\sigma^2(x)},$$

where $\sigma^2$ represents temporal variance. Thus defined $\mathcal{V}(x,\hat{x})$ can take on values from $-\infty$ to 1. Values near unity indicate that the timing and magnitude of $x$ are well explained by $\hat{x}$; large negative values indicate that $x$ and $\hat{x}$ are out of phase and/or that $\sigma(\hat{x})$ is larger than $\sigma(x)$. By way of example, a value of $\mathcal{V}(x,\hat{x}) = 0.5$ indicates that $\hat{x}$ explains 50% of $x$’s variance (i.e., removing $\hat{x}$ from $x$ amounts to a 50% decrease in variance relative to $x$), while a value of $\mathcal{V}(x,\hat{x}) = -0.5$ indicates that $\hat{x}$ in fact increases the variance of $x$ by 50% (i.e., removing $\hat{x}$ from $x$ amounts to a 50% variance increase). As shown in Figs. 2d–e, seasonal and interannual variance in $T$ is mostly explained by $T_{\nu\hat{T}}$ at tropical and subtropical latitudes. At nearly all latitudes, $\mathcal{V}(T, T_{\nu\hat{T}})$ deviates considerably from zero, suggesting that seasonal $\theta$ changes cannot be entirely neglected, however (Fig. 2d). At higher latitudes, variance in $T$ is generally not attributable to any one component alone and depends upon $\nu$ and $\theta$ variability as well as their covariability (Figs. 2d–e).

The dominant role of $T_{\nu\hat{T}}$ in setting $T$ variance at tropical and subtropical latitudes agrees qualitatively with previous modeling work (Jayne and Marotzke 2001; Dong and Sutton 2002). Although our results agree with earlier studies in that heat transport variations due to $v$–$\theta$ covariation are relatively small at tropical and subtropical latitudes (Dong and Sutton 2002), they are in contrast to work done at higher grid resolutions suggesting that covariable contributions associated with mesoscale eddy processes (that cannot be resolved by coarse-resolution models) can explain up to 20% of the seasonal-to-interannual meridional heat transport variance in the tropics and along latitudes of strong western boundary currents (Jayne and Marotzke 2002; Volkov et al. 2008).

b. Circulation component decomposition

To elucidate the components of the circulation through which the ocean transports heat, the following decomposition suggested by Bryden and Imawaki (2001) is used:

$$T = \int \int \rho_0 c_p v_m \theta_m \, dx \, dz + \int \int \rho_0 c_p v_\nu \theta_\nu \, dx \, dz + \int \int \rho_0 c_p v_h \theta_h \, dx \, dz,$$

where, for clarity,

$$v_\nu = \int \int v \, dx \, dz / \int \int dx \, dz$$

and likewise for potential temperature. Following the terminology of Peixoto and Oort (1992), the first term on the right-hand-side of Eq. (4) is temperature transport associated with the zonally symmetric barotropic flow ($T_m$), the second term heat transport carried by the zonally symmetric baroclinic flow ($T_\nu$), and the final term heat transport associated with the zonally asymmetric branch of the flow ($T_h$). To provide a physical intuition, $T_m$ can be understood as temperature transport owing to net mass flux across a section (and as such does not represent a meaningful heat transport), $T_\nu$ overturning-related heat transport (which includes transports related to Ekman circulation), and $T_h$ heat transport by gyres and other shorter-scale spatial correlations.

Decompositions are presented in Fig. 3. While $T$ is determined by $T_{\nu\hat{T}}$ over the range 0°–45°N, $T_{\nu\hat{T}}$ is not negligible elsewhere (Fig. 3a). Regarding seasonal and interannual variability, $T_{\nu\hat{T}}$ accounts for the magnitude and variance of $T$ in the tropics and subtropics (Fig. 3b–e). North of 45°N the balance can be more complex; in some instances, $\sigma(T)$ can be smaller than $\sigma(T_\nu)$ and $\sigma(T_h)$, and both $\mathcal{V}(T, T_\nu)$ and $\mathcal{V}(T, T_h)$ can be substantially different from zero (e.g., c. 50°N in the case of T$^S$; cf. Figs. 3b,d). This suggests that the combined effects of $T_h$ and $T_\nu$ are required to explain the heat transport at these latitudes. Contributions from $T_\nu$ are also not negligible in the southern subtropics, for example, near 20°S (Figs. 3c,e). That $T_h$ is more influential than $T_\nu$ at subpolar latitudes probably reflects stronger zonal temperature gradients, reduced vertical temperature gradients, and a more barotropic flow regime relative to lower latitudes.

These results are consistent with previous studies. In their inverse solution based on observations, Ganachaud and Wunsch (2003) find that $T_h$ is southward at 30°S and 8°N, $T_{\nu\hat{T}}$ and $T_\nu$ are of the same magnitude and sign at 5°S, and $T_{\nu\hat{T}}$ is the main contributor to $T$ at 19°S and 24°N (their Fig. 8). The modeling study of Dong and Sutton (2002) reveals qualitatively similar results for the time-mean circulation components, showing the importance of $T_h$ at boreal latitudes (their Fig. 3). Grist et al. (2010) present time-mean and interannual heat transport decompositions from a $1/4^\circ$-resolution simulation that are nearly indistinguishable from estimates presented here, though minor differences are apparent (their Fig. 6). For example, local divergence values (\(\sigma(T/\hat{T})\)) are larger in the Grist et al. solution than in the ECCO solution (Fig. 3).
This is likely due to mesoscale and other small-scale processes that are very much smoothed in the 1° ECCO solutions.

4. Heat storage rate budgets

Having established which processes contribute to variations in meridional heat transport, we now seek to determine how these factors influence variable rates of Atlantic Ocean heat storage. Variability in heat transport contributes to heat storage within an oceanic volume \( V \) by means of the heat transport divergence \( D \) over the volume

\[
D = -\iiint \nabla \cdot (c_p \rho_0 \mathbf{u} \theta) \, dV
= -\iiint (c_p \rho_0 \mathbf{u} \theta) \cdot \mathbf{n} \, dA \quad \text{[Watts]},
\]

(6)

Here, \( \mathbf{u} \) is the three-dimensional velocity field (m s\(^{-1}\)), \( A \) is the surface area bordering the volume (m\(^2\)), and \( \mathbf{n} \) is the normal unit vector. Here, \( D \) represents the difference between the heat storage rate and the combined action of surface heat fluxes and diffusive mixing. The heat storage rate budget within \( V \) can be defined as the volume integral of the forced advection–diffusion equation for local potential temperature conservation

\[
\iiint c_p \rho_0 \left[ \frac{\partial \theta}{\partial t} + \mathbf{v} \cdot \nabla \theta + \frac{1}{c_p \rho_0} \frac{\partial Q}{\partial z} \right] \, dV \quad \text{[Watts].}
\]

(7)

Here, \( \mathbf{v} \) is the diffusive flux of potential temperature (m \( \cdot \) C s\(^{-1}\)), and \( Q \) is the surface heat flux per unit area (W m\(^{-2}\)). Hereafter the (volume-integrated) four terms inside the brackets will be referred to as (from left to right) the heat storage rate \( (H) \), advection or heat transport divergence \( (D) \), diffusion or mixing \( (M) \), and surface heat flux \( (Q) \) (all in units of Watts). To investigate where \( D \) is important to \( H \) variability over large spatial scales in the Atlantic Ocean, Eq. (7) is studied in six geographic regions (Table 3; Fig. 4). These control volumes were devised with past hydrographic studies in mind. In these regions, the budget is integrated vertically to several depths at seasonal and interannual time scales. Budgets are presented for the

![FIG. 4. Atlantic Ocean study regions. Regions are named according to bounding WHP transect lines.](Unauthenticated Downloaded 07/14/22 01:47 AM UTC)
Fig. 5. (left) Twelve-month seasonal (month 1 represents January) and (right) 144-month interannual time series of near-surface (0–100 m) budgets. All dependent axes are in Petawatts. Budget terms are denoted as heat storage rate $H$ (red solid); heat flux divergence $D$ (black solid); diffusive mixing $M$ (green dash-dot); and surface heat exchange $Q$ (blue dashed).
case of the “near surface” (the vertical integral in the budget equation taken from the sea surface to a depth of 100 m and denoted by subscript ns) as well as for the case of the “upper ocean” (from the sea surface to a depth of ~1000 m and denoted by subscript uo). Near-surface and upper-ocean depths were chosen to correspond approximately to the vertical layers within which most of the regional heat storage rate variability is confined on seasonal and interannual time scales, respectively; in a rough sense, near-surface and upper-ocean budgets can be thought of as representing the surface mixed layer and depths above the base of the main thermocline, respectively, though we note that mixed layer and thermocline depths vary considerably in space and time and that the formulation of truly representative budgets would be distinct and more involved (e.g., Vialard and Delecluse 1998); such is beyond the scope of our present purpose, however. Full-depth budgets were also analyzed but are not presented because they are nearly identical to the upper-ocean budgets; we discuss full-depth budgets only in the context of our diagnosis of heat transport divergences (see below).

a. Near-surface budgets

Full time series of seasonal and interannual near-surface budget terms are shown in Fig. 5 and summarized statistically in Fig. 6. In most areas, $H_{ns}^S$ changes are related to $Q_{ns}^S$ variations, but the influence of $D_{ns}^S$ is evident in region A0608 and perhaps A0506: small $D_{ns}^S$ fluctuations in region A0506 have a sizable effect on $H_{ns}^S$ in August and September and are out-of-phase with $Q_{ns}^S$ and $H_{ns}^S$, resulting in a negative $V(H_{ns}^S, D_{ns}^S)$ value; $Q_{ns}^S$ and $D_{ns}^S$ make comparable contributions to the magnitude and timing of $H_{ns}^S$ in region A0608 (Figs. 5 and 6). That $H_{ns}^S$ away from the equator is mostly due to the ocean’s local response to $Q_{ns}^S$ agrees with previous theoretical (Gill and Niiler 1973) and observation-based (Hsiung et al. 1989) work; the important action of $D_{ns}^S$ at low latitudes likely reflects vertical motions associated with seasonal winds (e.g., Merle 1980; Böning and Herrmann 1994). Further, our closed budget approach provides details regarding the influence of mixing. Here, $M_{ns}^S$ is small, but this integral-view masks the role of mixing in the near-surface ocean: $M^S$ nearly balances $Q_{ns}^S$ in the upper 10 m but is compensated by $M^S$ deeper within the near-surface ocean (not shown), evidencing that while mixing can be important at particular depths, the net effect of $M$ integrated vertically to sufficient depth is small.

Regarding interannual time scales, $Q_{uo}^S$ is the most important contributor, yet it does not always account for the variance in $H_{uo}^S$ (Fig. 6d). Advection is important in some areas, most notably in region A0608, which is probably indicative of coupled atmosphere–ocean interactions known to underlie interannual sea surface temperature variability along the equatorial Atlantic (e.g., Ding et al. 2010). Mixing is compensated within the top 100 m (not shown) and hence the net effect of $M_{uo}^S$ is small (Figs. 5 and 6). A major exception is the highest latitude region A0203, characterized by a complex balance in which all budget terms appear important (Figs. 6b,d); the apparent influence of mixing here is simply due to the fact that mixed layer depths in this region typically exceed 100 m (not shown).

b. Upper-ocean budgets

Time series of upper-ocean budget terms are shown in Fig. 7 and summarized statistically in Fig. 8. In areas A0203 and A0305, $H_{uo}^S$ mostly reflects a local response to $Q_{uo}^S$, yet elsewhere the influence of advective processes occurring below the near surface is evident. In regions A0506, A0809, and A0910, $Q_{uo}^S$ is the main contributor to $H_{uo}^S$; however, $D_{uo}^S$ is out-of-phase with $Q_{uo}^S$, reflected in negative variances explained and resulting in $H_{uo}^S$ magnitudes that are somewhat smaller than corresponding $H_{uo}^S$ values. In equatorial region A0608, $D_{uo}^S$ plays the dominant role, evidencing the importance of seasonal processes occurring below the mixed layer and likely pointing to the ocean’s dynamical response to variable winds (e.g., Merle 1980). Wang and Carton (2002), who make use of a data assimilation product to study the top 500 m of the Atlantic, find similar qualitative results—at midlatitudes surface heat fluxes balance the seasonal heat storage, while closer to the equator heat transport divergences match the seasonal storage.
FIG. 7. As in Fig. 5, but for the upper ocean (0–1000 m).
correlated to sections nominally bounding the study regions (Table 3; Fig. 4). In most regions \( V \) (Interannual) \( 0.82 \) \( 0.96 \) \( 0.99 \) \( 0.99 \) \( 0.96 \) \( 0.98 \) \( \text{Seasonal} \) \( 0.95 \) \( 0.96 \) \( 1.00 \) \( 0.99 \) \( 0.88 \) \( 1.00 \)

in regions where northern/southern boundaries deviate considerably from the nominal bounding latitudes (e.g., A0203; see Fig. 4).

magnitudes and both contribute to \(\text{ern regions A0203 and A0305,} \)

Diagnosing the heat transport divergence fluctuations in equatorial region A0608, whereas in northerm regions A0203 and A0305, \( Q_{uo} \) and \( D_{uo} \) have similar magnitudes and both contribute to \( H_{uo} \). These results appear to be at odds with the conclusions of Grist et al. (2010) that year-to-year surface heat flux variations are of minor importance in the subpolar and subtropical North Atlantic but contribute as much as advection in the tropical North Atlantic. However, the model integration studied by Grist and colleagues (1958–2001) is considerably longer than the period studied here, and their results could reflect the influence of decadal-scale variations not accessible to our analysis.

c. Diagnosing the heat transport divergence

Because we are ultimately interested in attributing heat storage rates to changes in the general circulation, we now examine what variables and circulation components most contribute to the heat transport divergence in the study regions. By Gauss’s Theorem (see Eq. 6), the heat transport divergence over a full-depth, basinwide control volume can be computed as the difference in heat transport across the section’s northern and southern boundaries. Because the control volumes studied here are not bounded on the north and south by strict parallels (Fig. 4), decompositions of heat transport divergences over the study regions are estimated as the difference in the respective component of the meridional heat transport evaluated at the nominal latitudes bounding the control volume (Table 3). This yields a very good approximation to heat transport divergences computed over the control volumes proper (Table 4). Statistical summaries of decompositions of full-column heat transport divergence time series are presented in Figs. 9 and 10.

The magnitude and timing of \( D \) is well explained by \( D_{uo} \) in a majority of regions on both seasonal and interannual time scales (Fig. 9). In regions A0203 and A0305, however, interannual fluctuations (and to a lesser extent seasonal variations) in \( D_{uo} \) are nonnegligible and \( D \) is not wholly accounted for by \( D_{uo} \) changes (Figs. 9b,d), revealing that temperature variations can be important at higher latitudes.

Decompositions of \( D \) into the components of Bryden and Imawaki (2001) (Fig. 10) are qualitatively similar to decompositions of the heat transport presented in Fig. 3. Here, \( D_s \) is important in attributing \( D \) in all regions and on both seasonal and interannual time scales; however, \( D_h \) can also be important in several instances (Fig. 10). In regions A0305, A0506, and A0608, variations in \( D_h \) are controlled mostly by \( D_{uo} \), but \( D_{uo} \) fluctuations can contribute to the magnitude of \( D_h \). Secondary contributions from \( D_h \) are also evident in regions A0809 and A0910, especially at interannual time scales. In the northernmost region A0203, \( D_s \) and \( D_h \) contribute equally to the timing and magnitude of \( D \).

5. Summary and discussion

In this study we sought to relate heat storage rates in the Atlantic Ocean to changes in the general circulation. For this purpose we have employed an ocean state estimate produced by the ECCO project (Wunsch and Heimbach 2007; Wunsch et al. 2009). The state estimate utilizes most available ocean observations, fitting them to a state-of-the-art general circulation model using advanced optimization methods. The solution retains dynamical consistency and is ideally suited for closed budget analysis.

<table>
<thead>
<tr>
<th>A0203</th>
<th>A0305</th>
<th>A0506</th>
<th>A0608</th>
<th>A0809</th>
<th>A0910</th>
</tr>
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<tbody>
<tr>
<td>( V ) (Seasonal)</td>
<td>0.95</td>
<td>0.96</td>
<td>1.00</td>
<td>0.99</td>
<td>0.88</td>
</tr>
<tr>
<td>( V ) (Interannual)</td>
<td>0.82</td>
<td>0.96</td>
<td>0.99</td>
<td>0.99</td>
<td>0.96</td>
</tr>
</tbody>
</table>

Fig. 8. As in Fig. 6, but for the upper ocean.

Substantial differences between interannual near-surface and upper-ocean budgets (cf. Figs. 5 and 7) indicate considerable interannual variability deeper than 100 m. Generally, \( D_{uo} \) is the dominant contributor to \( H_{uo} \) and \( Q_{uo} \) plays a somewhat lesser role, though the particular details of the balance vary depending on region. For instance, \( D_{uo} \) is responsible (to first order) for the size and timing of \( H_{uo} \) fluctuations in equatorial region A0608, whereas in northern regions A0203 and A0305, \( Q_{uo} \) and \( D_{uo} \) have similar magnitudes and both contribute to \( H_{uo} \). These results appear to be at odds with the conclusions of Grist et al. (2010) that year-to-year surface heat flux variations are of minor importance in the subpolar and subtropical North Atlantic but contribute as much as advection in the tropical North Atlantic. However, the model integration studied by Grist and colleagues (1958–2001) is considerably longer than the period studied here, and their results could reflect the influence of decadal-scale variations not accessible to our analysis.

Table 4. Variance in the heat divergence \( D \) explained by an approximation \( D_i \) computed as the difference in heat transport across zonal sections nominally bounding the study regions (Table 3; Fig. 4). In most regions \( D_i \) captures well \( D \)’s variance. In all cases, \( D_i \) is highly correlated to \( D (r = 0.9); \) not shown). However, in some instances \( \sigma(D) \) can differ from \( \sigma(D) \) by up to 20% (not shown). This is most evident in regions where northern/southern boundaries deviate considerably from the nominal bounding latitudes (e.g., A0203; see Fig. 4).
We contextualized our budget study by first presenting estimates of the time-mean and anomalous heat transports, decomposing the transports in terms of relevant variables and circulation components. Next, the role of heat transport divergence in determining heat storage rates as a function of region, depth, and time scale was studied. Last, it was determined what variables and components of the circulation contribute most to the behavior of the heat transport divergence. Viewed as a whole, these results provide insight on the role of circulation changes in determining Atlantic Ocean heat storage rates.

The relationship of meridional heat transport changes to velocity field changes or diagnostics of the overturning circulation depends strongly on latitude and time scale (Figs. 2–3). Seasonal heat storage changes are confined mostly to the top 100 m of the ocean, although contributions are evident at greater depths in low-latitude areas (Figs. 5–8). More than 90% of the interannual ocean heat storage rate variance is found at depths above 1000 m (not shown). Advection of heat is important in low-latitude regions, contributing to the magnitude and timing of the heat storage (Figs. 5–8). On interannual scales, advection is the primary contributor to heat budgets in the upper ocean; it is larger than the anomalous surface heat flux in all regions and alone it explains more than half of the heat storage rate variance in most regions (Figs. 7 and 8).

The above results are generally consistent with prior studies, yet further insight is gained by decomposing the advection in terms of velocity and temperature variations as well as components of the circulation (Figs. 9 and 10). Variance in heat advection is mostly accounted for by changes in the meridional velocity field (Fig. 9). Temperature changes are not always irrelevant, however, making important contributions to the heat transport divergence at higher latitudes (Fig. 9). In equatorial regions and at northern tropical and subtropical latitudes, changes in advection are attributable to heat transport carried by the zonally symmetric baroclinic flow (Fig. 10). Outside of this range, consideration of heat transport changes associated with zonally symmetric baroclinic and zonally asymmetric branches of the circulation becomes necessary to attribute the advection (Fig. 10).

Shallow, intermediate, and deep components of the general circulation are thought to contribute to time-mean heat budgets in the subtropical Atlantic (Talley 1999, 2003), but little is known about how heat storage is affected by overturning changes. Our results suggest that seasonal and interannual changes in the meridional overturning circulation can be important to large-scale heat storage rates. For the latitudes studied, Atlantic heat storage rate changes are confined mainly to the upper ocean. The zonally symmetric baroclinic flow generally comprises contributions from intermediate and deep overturning circulations as well as surface Ekman and deeper return flows and other shallow overturning circulations (Bryden and Imawaki 2001; Talley 2003). In our analysis, a combination of surface heat fluxes, shallow overturning (including Ekman flows), and overturning of intermediate waters (below c. 500 m) accounts for ocean heat storage rates near the equator.
as well as for interannual storage at most tropical and subtropical latitudes; overturning of deep waters (below c. 2000 m) probably is not important on these time scales.

Although overturning changes are believed to impact Atlantic sector climate via direct influence on sea surface temperatures, the details of this influence have not been established (Lozier 2010). Results presented here reveal instances in which changes in the zonally symmetric baroclinic circulation can be readily related to upper-ocean heat storage rates (Figs. 7, 8, and 10). Furthermore, within approximately 10° of the equator (i.e., region A0608), the timing of interannual near-surface and upper-ocean heat storage rate and advection fluctuations are all very similar (cf. Figs. 5 and 7); in this region, interannual near-surface heat storage rate variations can be related to changes in the full column heat transport divergence (correlation between $H_{ns}^I$ and $D_k^I$ is 0.73) and (to a somewhat lesser extent) to the component associated with the zonally symmetric baroclinic circulation (correlation between $H_{ns}^I$ and $D_v^I$ is 0.65). At these latitudes, however, heat transport by the zonally symmetric baroclinic branch of the circulation is not directly equivalent to heat transport by intermediate and deep overturning but also includes a shallower wind-driven component (cf. Figs. 5 and 7). More generally, sea surface temperatures are not easily related to temperatures within the ocean (e.g., Donlon et al. 2002). Thus, establishing links between changes in sea surface temperature proper and the deeper meridional overturning circulation could be difficult in all the regions we have examined.

Some caveats of the present study should be acknowledged. In particular, the ECCO solution used in these analyses does not permit eddies, a sea ice model is lacking, and latitudes higher than 80° are not included. The sensitivity of our results to incorporation of any of these elements remains to be examined. As noted above, while our decomposition of meridional heat transports into circulation components (Fig. 3) agrees with a study based on a truly global, eddy-permitting model with a sea ice component (Grist et al. 2010), our assessment of the importance of covariable heat transports (Fig. 2) is lower than what one would expect based on eddy-permitting-model results (Volkov et al. 2008).³ The analysis period considered (1993–2004) also limits the range of time scales available for analysis. Once present shortcomings are resolved, future heat budget studies should allow us to explore, for example, how near-surface temperatures, sea ice cover, and circulation changes relate to one another in the high-latitude North Atlantic on decadal time scales (cf. Deser et al. 2002). Parallel efforts to incorporate elements mentioned above are being undertaken presently. (For a complete list of ECCO-related efforts and products, see http://www.ecco-group.org/).

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³ To be clear, our assessment of covariable heat transports does not constitute an assessment of eddy heat transports; in this solution, subgrid-scale eddy processes are parameterized and contained in the mixing terms, and thus are not contained in our heat transport decompositions; further analysis is needed to examine how the effects of parameterized subgrid-scale processes compare to higher-resolution solutions where eddies are permitted or resolved.

REFERENCES


