CORRIGENDUM

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Because of a production error, Buckley et al. (2014) contains a number of instances where the variable $T$ was mistakenly replaced by $\theta$. These errors occur in section 5a and in the caption to Fig. 14. To present the article as correctly as possible, it is reproduced in its entirety in this corrigendum as it was meant to appear originally.

The staff of the Journal of Climate regrets any inconvenience these errors may have caused.

REFERENCE


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Low-Frequency SST and Upper-Ocean Heat Content Variability in the North Atlantic

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ABSTRACT

A recent state estimate covering the period 1992–2010 from the Estimating the Circulation and Climate of the Ocean (ECCO) project is utilized to quantify the upper-ocean heat budget in the North Atlantic on monthly to interannual time scales (seasonal cycle removed). Three novel techniques are introduced: 1) the heat budget is integrated over the maximum climatological mixed layer depth (integral denoted as $H$), which gives results that are relevant for explaining SST while avoiding strong contributions from vertical diffusion and entrainment; 2) advective convergences are separated into Ekman and geostrophic parts, a technique that is successful away from ocean boundaries; and 3) air–sea heat fluxes and Ekman advection are combined into one local forcing term. The central results of our analysis are as follows: 1) In the interior of subtropical gyre, local forcing explains the majority of $H$ variance on all time scales resolved by the ECCO estimate. 2) In the Gulf Stream region, low-frequency $H$ anomalies are forced by geostrophic convergences and damped by air–sea heat fluxes. 3) In the interior of the subpolar gyre, diffusion and bolus transports play a leading order role in $H$ variability, and these transports are correlated with low-frequency variability in wintertime mixed layer depths.

1. Introduction

Observations (instrumental records and proxy data) indicate that Atlantic sea surface temperatures (SSTs) exhibit significant low-frequency (intraannual–decadal) variability. Atlantic SST variability impacts regional and global climate, including temperatures across North America and Europe (Sutton and Hodson 2005; Pohlmann et al. 2006), rainfall in the Sahel region (Zhang and Delworth 2006), and frequency and intensity of Atlantic hurricanes (Knight et al. 2006; Zhang and Delworth 2006). Furthermore, knowledge of low-frequency SST variability is essential for efforts aimed at decadal climate predictions (Smith 2007; Keenlyside et al. 2008). However, the underlying causes of low-frequency SST variability are poorly understood. In particular the relative contributions of local atmospheric forcing and ocean dynamics in creating low-frequency SST anomalies remain to be quantified. In this work, we utilize a recent state estimate from the Estimating the Circulation and Climate of the Ocean (ECCO) project to examine the relative roles of local atmospheric (wind and buoyancy) forcing and ocean dynamics in creating intraannual–interannual SST and upper-ocean heat content (UOHC) anomalies.

The “null hypothesis” for the origin of midlatitude SST anomalies is that they reflect the passive response of the ocean to stochastic atmospheric forcing. Based on the theory of stochastic climate models (Hasselmann 1976), Frankignoul and Hasselmann (1977) demonstrate that the statistical properties of observed midlatitude SST anomalies are well reproduced by the response of the ocean to stochastic air–sea heat fluxes and to a lesser extent Ekman transport anomalies (Frankignoul 1985). For example, the tripole SST anomalies seen in the North Atlantic are primarily forced by anomalous air–sea heat flux and Ekman transport anomalies associated with the North Atlantic Oscillation (NAO) (Cayan 1992a,b), and these fluxes are primarily attributable to weather noise (Fan and Schneider 2012). Frankignoul and Reynolds (1983) utilize SST and air–sea heat flux...
fields over the North Atlantic to estimate the damping parameter for SST anomalies to be $\alpha = 20 \text{W m}^{-2} \text{K}^{-1}$, which corresponds to an $e$-folding time scale of 2–6 months for mixed layer depths (MLDs) of 30–75 m. Barsugli and Battisti (1998) extend the stochastic climate model of Frankignoul and Hasselmann (1977) to include thermodynamic coupling between the atmosphere and ocean and show that in midlatitudes coupling decreases atmosphere–ocean heat fluxes and increases the variance of both SST and atmospheric temperatures at low frequencies.

In simple stochastic climate models (e.g., Frankignoul and Hasselmann 1977), the MLD must be accurately specified (as a parameter) in order to reproduce the observed magnitude of SST anomalies. More sophisticated models determine the MLD by invoking a turbulent kinetic energy budget (Niiler and Kraus 1977). Numerous studies have shown that in many regions, SST anomalies are well reproduced by forcing a one-dimensional mixed layer model with the observed atmospheric air–sea heat fluxes (e.g., Seager et al. 2000).

Despite the success of the null hypothesis in replicating SST variability in many regions, there is ample evidence that more complex dynamics play a role. Including reemergence either implicitly by reformulating the model in terms of an effective ocean thermal capacity given by the depth of the winter mixed layer or explicitly by considering entrainment through a time-variable mixed layer. de Coëtlogon and Frankignoul (2003) demonstrate that the persistence of the tripole SST anomalies associated with the NAO may be explained through the reemergence mechanism.

While the one-dimensional processes of local atmospheric forcing and reemergence may explain SST anomalies over much of the ocean, evidence suggests that ocean advection may play a role, particularly in regions of strong currents and on decadal and longer time scales. Bjerknes (1964) and Kushnir (1994) examine relationships between Atlantic SST and sea level pressure anomalies and conclude that, while interannual SST variability is primarily driven by air–sea heat fluxes forced by atmospheric variability, ocean dynamics plays a role in setting SST on decadal time scales. Gulev et al. (2013) utilize SST and turbulent air–sea heat flux estimates to show that in the midlatitude North Atlantic surface turbulent heat fluxes are driven by the ocean on time scales longer than 10 years. The spatial pattern of decadal SST anomalies (generally a single polarity over the entire North Atlantic with maximal anomalies in subpolar regions) has led numerous authors to conclude that these anomalies are due to changes in the strength of the Atlantic meridional overturning circulation (AMOC; e.g., Kushnir 1994). Dong and Kelly (2004) and Dong et al. (2007) argue that interannual UOHC anomalies in the Gulf Stream region are forced by changes in ocean geostrophic advection and damped by air–sea heat fluxes. Marshall et al. (2001a) and Czaja and Marshall (2001) develop simple models to show that SST anomalies across the Gulf Stream path result from the delayed adjustment of the gyre circulation and/or the AMOC to stochastic wind forcing. de Coëtlogon and Frankignoul (2003) find that allowing for reemergence to occur nonlocally in accordance with estimated advection pathways leads to increased wintertime persistence, particularly in the northern portion of the North Atlantic.

The goal of this paper is to extend previous work through a refined examination of the relative contributions of local processes (atmospheric forcing and reemergence) and ocean dynamics in setting SST and UOHC on intraannual–interannual time scales. The refinement is afforded through the use of a dynamically and kinematically consistent ocean state estimate produced by the ECCO project. Two major requirements, in particular, are satisfied: 1) compared to free-running coupled or ocean-only general circulation models (GCMs), ECCO estimates are consistent (within derived uncertainty estimates) with existing ocean observations; 2) in contrast to filter-based (sequential) reanalysis products, which incur artificial heat sources/sinks during the analysis steps, ECCO estimates are the result of a smoother-based method. As such they are
free of artificial internal heat and freshwater sources/sinks and fulfill known conservation laws exactly. This enables accurate (closed budget) term-by-term diagnostics of the heat equation.

This effort is complementary to the study of Piecuch and Ponte (2012b), which utilizes a previous ECCO estimate to demonstrate the importance of advection in setting UOHC in several broad latitude bands. Their results are extended in three important ways: 1) Advection transports resulting from local Ekman transports are isolated, allowing a separation between locally forced UOHC anomalies and ones which require active (geostrophic) ocean dynamics. 2) The dominant modes of SST and UOHC variability are considered, rather than restricting the analysis to latitude bands. 3) Rather than considering budgets over somewhat arbitrarily defined fixed-depth layers, we consider UOHC budgets integrated over the maximum climatological mixed layer depth, which reflects the portion of the ocean that comes in contact with the atmosphere.

In section 2 we introduce the current ECCO estimate, its fit to observations, and the suitability of ECCO for understanding SST and UOHC variability. In section 3 we describe the estimate’s SST and UOHC variability. The roles of local atmospheric forcing (air–sea heat fluxes and Ekman transports) and ocean dynamics (primarily geostrophic advection) in setting SST and UOHC are determined in section 4. Furthermore, in section 5 the North Atlantic is divided into several dynamically distinct regions, and the important terms in the UOHC budget in each of these regions are examined. Finally, the main conclusions of our work and how they relate to previous studies are discussed in section 6.

2. The ECCO state estimate

a. Overview

In the ECCO ocean state estimation project, the Massachusetts Institute of Technology General Circulation Model (MITgcm) is fit in a least squares sense to several hundred million (satellite and in situ) ocean observations spanning the last two decades (Wunsch and Heimbach 2007; Wunsch et al. 2009). Each data point is weighted by a best estimate of its observational and representation error variance, and the least squares problem is solved by the method of Lagrange multipliers, using an iterative process relying upon a gradient search. The fit is achieved by adjusting uncertain variables (surface forcing, initial conditions, and interior mixing coefficients). The model is then run forward in time using the adjusted parameters, free of any constraints. The resulting model outputs are dynamically and kinematically consistent: they satisfy the equations of motion and preserve property budgets exactly (Wunsch and Heimbach 2013a). Thus, ECCO estimates are particularly well suited for exploring mechanisms of SST and UOHC variability.

Through the optimization process, atmospheric forcings are adjusted (within error bars) in order to make the ECCO estimates consistent with ocean observations (also within error bars). As there are substantial uncertainties in various atmospheric data products, making adjustments within those error bars is a valid and consistent formal estimation approach. However, it is possible that the adjustments to atmospheric forcings degrade the atmospheric forcing by, for example, compensating for model errors.

As ECCO estimates are produced using an ocean-only GCM with specified (albeit adjusted) atmospheric forcing, they cannot be utilized to address the origin of atmospheric forcing. Variability in atmospheric forcing may result from internal atmospheric dynamics (Frankignoul 1985; Fan and Schneider 2012), local thermodynamic atmosphere–ocean coupling (Barsugli and Battisti 1998; Frankignoul et al. 1998), and/or dynamic atmosphere–ocean coupling (see review by Kushnir et al. 2002).

A potential disadvantage of the ECCO estimate is its relatively coarse resolution (nominally 1°). In the real ocean there may be low-frequency mesoscale variability that is not resolved by ECCO (e.g., the Gulf Stream path modulation), and parameterizations utilized to capture the effects of unresolved variability are certainly imperfect.

The new-generation ECCO version 4 (ECCO v4) estimate released by the ECCO-Production project (G. Forget et al. 2014, unpublished manuscript) covers the period 1992–2010 (with periodic updates to extend the estimation period to near present). Like in past ECCO estimates, 1992 is chosen as the beginning of the estimation period to coincide with the beginning of the satellite altimetry period for which continuous global sea level observations are available. Unlike previous ECCO estimates, the domain is global: that is, it encompasses the Arctic. A new global grid was produced to this end with slightly higher resolution, isotropic meridional scaling, refined resolution in the tropics, and grid topologies adapted to the adjoint modeling infrastructure (G. Forget et al. 2014, unpublished manuscript). Other improvements over past ECCO solutions include a state-of-the-art dynamic–thermodynamic sea ice model (Losch et al. 2010) and the use of real freshwater fluxes in conjunction with a nonlinear free surface (Campin et al. 2004). See Wunsch and Heimbach (2013a) for a summary of the observational
constraints included in this estimate and Speer and Forget (2013) for a discussion of the estimate’s hydrographic properties. The specific ECCO v4 estimate discussed in this study is revision 3, iteration 3.

b. Comparison to observations

Figures 1a–d show potential temperature (θ) misfits in the upper ocean compared to in situ data for the “first guess” solution (no optimization; i.e., no adjustment of initial conditions, atmospheric forcing, or interior mixing coefficients) and the optimized ECCO v4 solution. The first-guess atmospheric forcing is given by the European Centre for Medium-Range Weather Forecasts (ECMWF) Interim Re-Analysis (ERA-Interim) atmospheric state (surface air temperature, specific humidity, precipitation, and downwelling radiation) and wind stress vector fields. See Wunsch and Heimbach (2013a,b) and G. Forget et al. (2014, unpublished manuscript) for details regarding the data used in ECCO v4. It is clear that the optimization process has improved the fit to observations substantially,
but errors on the order of 1°C are still present along the equator, along the Gulf Stream path, and in regions of the subpolar gyre. Normalized misfits (see Figs. 1e,f) are near one in most regions, indicating that the optimization procedure has succeeded in producing a model solution that matches observations within the observational uncertainties. Exceptions include the Greenland Sea, where normalized misfits remain significantly larger than one despite optimization, indicating a not fully converged solution in limited regions.

Figure 2 shows a map of the root-mean-square difference between Reynolds et al. (2002) mapped SST data and SST in (a) first-guess solution and (b) optimized solution (ECCO v4). Note the logarithmic color scale.

Figure 3 shows the first two empirical orthogonal functions (EOFs) and corresponding principle component (PC) time series of anomalies of monthly (seasonal cycle removed) North Atlantic SST (1992–2010) from Reynolds data and ECCO v4. In both Reynolds data and ECCO v4, the leading two modes of variability explain ~25% and ~15% of the spatially integrated variance, respectively. The correlation between the PC time series of Reynolds data and ECCO v4 is 0.95 for both PC1 and PC2. EOF1 resembles the classic SST tripole, which has been demonstrated to be associated with the NAO (Cayan 1992a,b), and both EOF1 and EOF2 strongly resemble the EOF patterns found by de Coëtlogon and Frankignoul (2003) (see their Fig. 5). Significant lagged correlations between PC1 and PC2 (maximum correlation of ~0.5 when PC1 leads by ~6 yr) indicate the presence of a propagating mode, although PC2 has the largest variance during the summertime (not shown), as was previously found by de Coëtlogon and Frankignoul (2003). The power spectra of the PC time series (see Fig. 3f) are red at high frequencies with slopes ranging from −1.6 to −1.8 and flatten out at a time scales between 2 and 5 years.

3. Upper-ocean heat content variability

In the remainder of this work, we examine the variability in ocean heat content between the free surface $\eta$ and a fixed (in time) depth $D$ defined by forming a monthly climatology (1992–2010) of MLD at each spatial location and choosing the maximum MLD from this monthly climatology. The heat content per unit area integrated over this layer is defined as

$$H = \rho_o C_p \int_{-D}^{\eta} \theta \, dz,$$

where $\theta$ is temperature, $\rho_o$ is the mean density, and $C_p$ is the heat capacity. More common metrics of UOHC include heat content integrated over fixed depth layers (Dong and Kelly 2004; Forget 2010; Piecuch and Ponte 2012b), isothermal layers (Dong et al. 2007; Forget et al. 2011), and the mixed layer (Foltz et al. 2003; Dong and Kelly 2004; Kim et al. 2006; Foltz et al. 2013). Our novel choice for defining $H$ is based on the following considerations:
(i) The layer from the ocean surface to \(-D\) reflects the portion of the ocean that comes in contact with the atmosphere.

(ii) Budgets from the surface to \(-D\) generally do not cut across the mixed layer. When a layer budget (e.g., fixed depth layers) cuts across the mixed layer, the UOHC budget will be dominated by air–sea heat fluxes and vertical mixing (diffusion), which redistributes heat within the mixed layer. Such a balance provides little insight into important dynamics.

(iii) Unlike budgets over the time-evolving mixed layer (e.g., Foltz et al. 2013), considering budgets over \(D\) eliminates the need to explicitly consider entrainment processes in UOHC since \(D\) is fixed in time (Deser et al. 2003).
output with the seasonal cycle removed by simply subtracting the mean monthly climatology from each month. Figure 5 shows the first two EOFs and PC time series of monthly anomalies of North Atlantic $H$. The spatial patterns of the EOFs of $H$ are similar to those of SST, but, as expected, the magnitude of $H$ variability is larger in the subpolar gyre where mixed layers are deep. As with SST (Fig. 3f), the power spectra of the PC time series of $H$ (Fig. 5b) are red at high frequencies and flatten out at low frequencies. $H$ variability is weaker at high frequencies and stronger at low frequencies compared to that of SST (i.e., power spectra of $H$ are steeper than those of SST) because of the integration over a deeper layer.

Figures 5e,f show SST anomalies projected\(^1\) on to the first two PC time series of $H$. The patterns of SST and $H$ are quite similar, indicating that at these low frequencies upper-ocean temperature anomalies are quite coherent with depth. Additionally, the spatial patterns of SST associated with the first two PCs of $H$ (Figs. 5e,f) and the first two EOFs of SST (Figs. 3c,d) are quite similar, indicating the PC time series of $H$ also capture a significant portion of low-frequency SST variance. In fact, the PC time series of SST and $H$ are highly coherent on interannual time scales (not shown).

4. Upper-ocean heat content budgets

a. Roles of advection, diffusion, and air–sea heat fluxes

In this section, we analyze the terms that are important in the $H$ budget. We integrate the conservation equation for heat from $\eta$ to $-D$,

$$\begin{align*}
\rho_o C_p \int_{-D}^{\eta} \frac{\partial \theta}{\partial t} dz &= -\rho_o C_p \int_{-D}^{\eta} \nabla \cdot (u \theta + u^* \theta) dz \\
&\quad - \rho_o C_p \int_{-D}^{\eta} \nabla \cdot K dz + Q_{net},
\end{align*}$$

where $u$ is the three-dimensional (explicit model) velocity; $u^*$ is the eddy-induced transport velocity parameterized by the Gent and McWilliams (1990) scheme; $K$ is the diffusive temperature flux resulting from diapycnal diffusion [convective adjustment, Gaspar et al. (1990) scheme, and background interior diffusion] and parameterized isopycnal diffusion (Redi 1982); and $Q_{net}$ is

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\(^1\)Projecting a data field onto a time series means computing the covariance between the time series and the data field at each spatial location.
the net (turbulent plus radiative) air–sea heat flux. The heat content tendency (term on the left-hand side, \( H_t \)) is given by the sum of the advective heat transport convergence (first term on right-hand side, \( C_{\text{adv}} \)), the diffusive heat transport convergence (second term on right-hand side, \( C_{\text{diff}} \)), and the air–sea heat flux less any shortwave radiation that passes through the bottom of the climatological mixed layer (last term on right-hand side, \( Q_{\text{net}} \)).

Figure 6 shows the variance of monthly anomalies of \( H_t \), as well as the contributions by \( C_{\text{adv}}, C_{\text{diff}}, \) and \( Q_{\text{net}} \). The term \( C_{\text{adv}} \) plays a significant role in creating variance of \( H_t \) in regions of strong currents/fronts, such as along the Gulf Stream path. The term \( Q_{\text{net}} \) is more
homogenous over the entire basin than $C_{\text{adv}}$, but it is also largest in regions of strong currents/fronts. Consistent with the fact that our chosen layer generally does not cut across the mixed layer, $C_{\text{diff}}$ only provides minor contributions, except over portions of the Gulf Stream (in particular the Mann eddy region centered at 45$^\circ$N, 40$^\circ$W) and along the boundaries in the subpolar gyre. The terms $C_{\text{adv}}$ and $Q_{\text{net}}$ are correlated over broad regions of the subtropical and subpolar gyres and anticorrelated over the tropics (see Fig. 7a), patterns that likely reflect the role of the winds in creating both air–sea heat flux and Ekman transport anomalies (Foltz and McPhaden 2006). The terms $C_{\text{adv}}$ and $C_{\text{diff}}$ are anticorrelated over the regions where $C_{\text{diff}}$ plays a sizable role in the $H$ budget (see Fig. 7b).

b. Separation of advective convergence into Ekman and geostrophic parts

A primary goal of this study is to determine the dynamical mechanisms that contribute to variability of $C_{\text{adv}}$, specifically whether advective heat transports are attributable to the local response to surface wind stress variability (Ekman transports) or ocean dynamics (likely primarily heat transport by geostrophic currents). We first note that monthly averaged $C_{\text{adv}}$ can be written as

$$C_{\text{adv}} = -\rho_o C_p \int_D \mathbf{v} \cdot (\mathbf{u} + \mathbf{u}^\prime) \, dz$$

$$= -\rho_o C_p \int_D \mathbf{v} \cdot (\mathbf{u} + \mathbf{u}^\prime) \, dz = C_{\text{lin}}$$

where overbars denote monthly mean variables and primes denote deviations from monthly means. The variance of the linear transport $C_{\text{lin}}$ and the eddy-driven (bolus) transport $C_{\text{bol}}$ are plotted in Figs. 8a,b. The variance of $C_{\text{lin}}$ is qualitatively quite similar to that of $C_{\text{adv}}$ (see Fig. 6b), and the correlation coefficient between monthly anomalies of $C_{\text{adv}}$ and $C_{\text{lin}}$ (see Fig. 9a) is greater than 0.8 almost everywhere. Figure 9c shows the fraction of the variance of $C_{\text{adv}}$ that is explained by $C_{\text{lin}}$. With the exception of shallow regions (particularly in

$$f = 1 - \sigma^2_{X,Y}/\sigma^2_X.$$
the subpolar gyre) and the region of the Mann eddy, the majority of the advective heat transport convergence is captured by the linear term.

We now develop a decomposition of $C_{\text{lin}}$ into Ekman and geostrophic parts. First, we decompose the horizontal velocity $u_H$ into Ekman and geostrophic parts, $u_H = u_{ek} + u_g$. Assuming that the Ekman mass transport is uniformly distributed within the Ekman layer ($-D_{ek} < z < 0$) and zero elsewhere, the Ekman velocity is given by

$$u_{ek} = \frac{M_{ek}}{D_{ek}} = \tau \times \mathbf{z} / \rho_o f D_{ek}, \tag{4}$$

where $M_{ek}$ is the Ekman mass transport, $f$ is the Coriolis parameter, and $\tau$ is the wind stress. Here, we assume that the Ekman layer is the shallower of $D$ and a depth of 100 m [a choice motivated by the assumption that $D_{ek} = 100$ m in the Rapid Climate Change (RAPID) AMOC estimates at 26°N]. The geostrophic velocity is given by

$$u_g = \frac{1}{f \rho_o} \mathbf{z} \times \nabla p, \tag{5}$$

where $p$ is the hydrostatic pressure. We then integrate the continuity equation from $\eta$ to $-D$ to approximate the vertical velocity at $z = -D$,

$$w(-D) = \frac{d\eta}{dt} + (E - P - R) + \int_{-D}^{\eta} \mathbf{V}_H \cdot \mathbf{u} \, dz = w_{ek}(-D) + w_g(-D), \tag{6}$$

where $E$, $P$, and $R$ are the evaporation, precipitation, and runoff fields, respectively. Here, the terms $E - P - R$ and $d\eta/dt$ are found to be two orders of magnitude smaller than the horizontal divergence and are thus neglected. The Ekman and geostrophic components of $C_{\text{lin}}$ are approximated as

$$C_{ek} = -\rho_o C_p \int_{-D_{ek}}^{\eta} \mathbf{V} \cdot (\mathbf{u}_{ek} \mathbf{\bar{u}}) \, dz$$

$$+ \rho_o C_p \bar{w}_{ek}(-D) \bar{\mathbf{u}}(-D), \tag{7}$$

and

$$C_g = -\rho_o C_p \int_{-D}^{\eta} \mathbf{V} \cdot (\mathbf{u}_g \mathbf{\bar{u}}) \, dz + \rho_o C_p \bar{w}_g(-D) \bar{\mathbf{u}}(-D). \tag{8}$$

where $\mathbf{\bar{u}}_{ek}$ and $\mathbf{\bar{u}}_g$ are given by Eqs. (4) and (5), respectively, and $\bar{w}_{ek}(-D)$ and $\bar{w}_g(-D)$ are given by Eq. (6).
Figure 8c shows the variance of our estimate of the linear heat transport convergence \( C_{ek} + C_g \), and Fig. 8d shows the variance of the error introduced by this estimation \( C_{err} = C_{lin} - (C_{ek} + C_g) \). The variance of \( C_{ek} + C_g \) is qualitatively quite similar to that of \( C_{lin} \) (see Fig. 8a), and \( C_{err} \approx C_{ek} + C_g \), except in a few shallow regions along the ocean boundaries. Physical reasons why \( C_{err} \) is large in shallow boundary regions include our neglect of inertia and viscous processes (lateral friction and bottom Ekman layers). The term \( C_{err} \) may also be large in these regions for numerical reasons; for example, errors are induced in the calculation of \( u_g = (u_g, v_g) \) near boundaries because on a C grid \( \partial \rho / \partial x \) is at \( u \) points and must be interpolated onto \( v \) points to calculate \( v_g \) (and similarly for the calculation of \( u_g \)). Correlations between \( C_{lin} \) and \( C_{ek} + C_g \) (see Fig. 9b) are greater than 0.8, and the fraction of the variance of \( C_{lin} \), explained by \( C_{ek} + C_g \) (see Fig. 9d) is greater than 0.7 everywhere except shallow boundary regions. In summary, with the exception of shallow coastal regions, \( C_{lin} \) is well reproduced by \( C_{ek} + C_g \) and our decomposition into Ekman and geostrophic components provides useful insight into the origin of the linear advective heat transport convergences observed in ECCO.

Variance maps of \( C_{ek} \) and \( C_g \) (see Figs. 8e,f) exhibit their largest values in regions of strong currents/fronts. Both strong geostrophic currents and large temperature gradients contribute to large values of \( C_g \) in these regions. While Ekman mass transports are more homogenous over the basin (not shown), \( C_{ek} \) is also large over these regions because of strong temperature gradients. Unlike \( C_g \), \( C_{ek} \) also exhibits significant variability over the interior of the gyres.

The calculation of \( C_{ek} \) is relatively insensitive to the assumed depth of the mixed layer \( D_{ek} \); taking \( D_{ek} = 50 \) m instead of 100 m results in negligible changes in \( C_{ek} \) (not shown). Since \( u_{ek} \) is constant over \( D_{ek} \), the first term in the equation for \( C_{ek} \) [Eq. (7)] can be written as (neglecting small horizontal variations in \( \eta \))
where $\bar{\vec{u}}$ is $\vec{u}$ averaged over the Ekman layer. The second term in the Equation for $C_{\text{ek}}$ is independent of $D_{\text{ek}}$. Hence, the value of $D_{\text{ek}}$ only determines the depth over which the temperature is averaged, which explains the relative insensitivity of the calculation to $D_{\text{ek}}$.

The correlation patterns between $C_{\text{ek}}$ and $Q_{\text{net}}$ (Fig. 7c) are quite similar to the correlation patterns between $C_{\text{adv}}$ and $Q_{\text{net}}$ (Fig. 7a). These correlation patterns are consistent with the hypothesis that changes in the wind stress lead to changes in both $Q_{\text{net}}$ and $C_{\text{ek}}$, a result that was previously found by Foltz and McPhaden (2006) in the tropics. For example, a positive phase of the NAO leads to negative $Q_{\text{net}}$ in the tropics and subpolar regions and positive $Q_{\text{net}}$ in the subtropics and northward Ekman transport anomalies (hence $C_{\text{ek}} > 0$) in the tropics/subtropics and southward Ekman transport anomalies (hence $C_{\text{ek}} < 0$) in the subpolar regions (Cayan 1992a, b; Marshall et al. 2001b). The resulting correlation patterns are negative in the tropics and positive in the subtropics and subpolar regions, as observed. If alternatively, as a reviewer suggested, changes in $C_{\text{ek}}$ forced SST anomalies that were then damped by air-sea heat fluxes, $C_{\text{ek}}$ and $Q_{\text{net}}$ would be anticorrelated everywhere. Correlations between $C_{\text{ek}}$ and $C_{g}$ (see Fig. 7d) and between $Q_{\text{net}}$ and $C_{g}$ (not shown) are generally small and without clear spatial structure.

The fraction of the variance of $H_t$ explained by our decomposition is summarized in the top panels of Fig. 10. Figure 10a shows the fraction of the variance of $H_t$ explained by $C_{\text{lin}} + Q_{\text{net}}$. Regions where this quantity is not approximately 1, such as shallow subpolar regions and the region of the Mann eddy, highlight areas where diffusion and bolus transports are important in the heat budget. Figure 10b shows the fraction of the variance of $H_t$ explained by $C_{\text{ek}} + C_{g} + Q_{\text{net}}$. The similarity between Figs. 10a and 10b indicates that approximating the linear convergence as the sum of the Ekman and geostrophic parts does not lead to substantial errors in the budget for $H_t$. The areas where the fraction of the variance of $H_t$ explained by $C_{\text{ek}} + C_{g} + Q_{\text{net}}$ is approximately 1 are the regions where we can understand the majority of the variance of $H_t$ by utilizing our decomposition of the heat transport convergence into Ekman and geostrophic parts.

To our knowledge, no author has previously utilized a decomposition into Ekman and geostrophic heat transport convergences that includes the vertical terms, as we have done here. However, a number of other studies have utilized similar decompositions. Marshall et al.
express the Ekman heat transport convergence as a pseudo air–sea heat flux $H_{ek} = C_{ek} \rho c D_{ek} \mathbf{u}_{ek} \cdot \nabla S_{ST}$, which is equivalent to our definition of the Ekman heat transport convergence if 1) we integrate over the full ocean depth so that the vertical term disappears, 2) we assume that the mean temperature over the depth of the Ekman layer is the same as the SST, and 3) $\theta \mathbf{V} \cdot \mathbf{u}_{ek} \ll \mathbf{u}_{ek} \cdot \nabla \theta$ so that $\mathbf{V} \cdot (\mathbf{u}_{ek} \theta) \approx \mathbf{u}_{ek} \cdot \nabla \theta$. Additionally, a number of authors have calculated the (depth integrated) meridional ocean heat transport (OHT) convergence resulting from Ekman transports (Levitus 1987; Jayne and Marotzke 2001), and the RAPID program utilizes a decomposition of the AMOC and OHT into Ekman and geostrophic parts in order to estimate their strengths at 26.5°N (Cunningham et al. 2007; Johns et al. 2011).

c. Role of local forcing

As stated earlier, a primary goal of this work is to determine the relative roles of local atmospheric forcing and ocean dynamics in setting $H_t$. Here, local atmospheric forcing is defined simply as the sum of the air–sea heat fluxes and Ekman heat transport convergences ($C_{loc} = Q_{net} + C_{ek}$). Our definition of $C_{loc}$ is motivated by an attempt to determine in what regions and on what time scales geostrophic ocean currents can be neglected as a source of variability of $H_t$. Furthermore, if $C_{ek}$ is primarily driven by changes in Ekman mass transports attributable to local wind variability rather than changes in the temperature field (Doney et al. 2007), $C_{loc}$ can be estimated directly from local atmospheric forcing fields. Since ECCO is an ocean-only model, we cannot address the origin of the atmospheric forcing: specifically, whether the ocean plays a role in variability of air–sea fluxes of heat, momentum, and freshwater (here called the atmospheric forcing). However, previous work (Kushnir et al. 2002; Schneider and Fan 2012) demonstrates that the response of the extratropical atmosphere to midlatitude SST anomalies is relatively small compared to internal atmospheric variability, so much of the variability of $C_{loc}$ likely reflects stochastic atmospheric forcing.

Figure 10c shows the fraction of the variance of $H_t$ explained by $Q_{net}$. As expected, because of large advective heat transport convergences over the Gulf Stream and other boundary current regions, air–sea heat fluxes cannot explain $H_t$ over these regions. However, even in the interior of the subtropical and subpolar gyres, $Q_{net}$ only explains on the order of 50% of the variance of $H_t$. Figure 10d shows the percent of the variance of $H_t$ explained by $C_{loc}$. Over the interior of the subtropical and eastern subpolar gyres, over 70% of the variance of $H_t$ is explained by local forcing. Thus, when attempting to isolate the role of local atmospheric forcing in creating $H_t$ anomalies, Ekman heat transport convergences should be included in addition to air–sea heat fluxes, which are traditionally used as the measure of the role of local forcing.
5. Regional analysis of \( H \) variability

We now consider \( H \) budgets over various regions in more detail. Gyre-scale regions are chosen in accord with the large-scale nature of the dominant modes of SST and \( H \) variability (see EOFs of SST and \( H \) in Figs. 3 and 5) and the relatively homogenous values of the fraction of the variance explained by local forcing over broad regions (values near 0.7 in the interior of the subtropical and subpolar gyres and values on the order of 0.5 over the Gulf Stream region). The regions considered are 1) the interior of the subtropical gyre, 2) the Gulf Stream, and 3) the interior of the subpolar gyre. The boundary between the subtropical and subpolar gyres is determined by the zero contour of the mean barotropic streamfunction (gray contour). The interior of the subtropical and subpolar gyres are determined by requiring that \( C_{\text{loc}} \) explains at least 70% of the variance of \( H \), (inside black contour) and the MLD is less than 1000 m. The Gulf Stream region is defined where the mean current speed is greater than 7 cm s\(^{-1}\), the latitude is between 25\(^{\circ}\) and 45\(^{\circ}\) N, and \( C_{\text{loc}} \) explains less than 70% of the variance of \( H \) (outside black contour). Time series of SST and \( H \) averaged over (c) interior of the subtropical gyre, (e) Gulf Stream region, and (g) interior of the subpolar gyre. Power spectra (black lines with uncertainty indicated by gray shading) of \( H \) in (d) interior of the subtropical gyre, (f) Gulf Stream region, and (h) interior of the subpolar gyre. The dashed black lines show a fit of Eq. (10) to the spectra. (b) Coherence between time series of SST and \( H \) in each region.
heat budget. Time series of SST and $H$ in each region (see Figs. 11c,e,g) are highly coherent on time scales longer than about 2 years (see Fig. 11b). The power spectra of $H$ in each region are red at high frequencies and flatten out at low frequencies (see Figs. 11d,f,h).

Here, we focus on three regions where diffusion and bolus transports do not play a significant role in the variance of $H_t$. In these regions we can separate the roles of local atmospheric forcing (air–sea heat fluxes and Ekman transport convergences) and ocean dynamics (geostrophic transports) in creating $H$ anomalies. However, the regions where these simple ideas fail are also interesting. In particular, the Mann eddy region has been isolated as a region where diffusion and bolus transports play a significant role in the heat budget. This region certainly deserves more study, in particular because previous work has suggested that this region is important for controlling variability of the AMOC (Forget et al. 2008a,b; Buckley et al. 2012; Tulloch and Marshall 2012).

\textit{a. Heat budgets over regions}

The left panels of Fig. 12 show the power spectra of the terms of the $H_t$ budget in each region. In all regions the spectra of $Q_{\text{net}}$ and $C_{\text{Ek}}$ (and hence $C_{\text{loc}}$) are essentially white, consistent with the hypothesis that they are driven by local wind forcing, whereas $C_g$ and $C_{\text{diff}} + C_{\text{bol}}$ have red spectra. Everywhere, $C_{\text{err}}$ is small, although in the Gulf Stream region $C_{\text{err}}$ exhibits increased variance at low frequencies. The right panels of Fig. 12 show the magnitude of the coherence between $H_t$ and $Q_{\text{net}}$, $C_{\text{loc}}$, and $C_{\text{loc}} + C_g$ in each region. The salient message is that in the interior of the subtropical gyre local atmospheric forcing explains the majority of the variability of $H_t$ on all time scales. In contrast, in the Gulf Stream region, $C_g$ plays a role in $H_t$ variability for periods longer than about 1 yr. In the subpolar gyre both $C_g$ and $C_{\text{diff}} + C_{\text{bol}}$ are important in setting $H_t$ for periods longer than about 2 years.

To better understand the significant low-frequency variability of $C_{\text{diff}} + C_{\text{bol}}$ in the subpolar gyre, yearly average time series of $C_{\text{diff}}$, $C_{\text{bol}}$, and their sum are compared to a time series of the March MLD averaged over the subpolar gyre (see Fig. 13). March MLD is several hundred meters deeper than $D$ (dashed red line in Fig. 13) for 1992–95 and generally somewhat shallower than $D$ after 1996. The time series of $C_{\text{diff}}$ and $C_{\text{bol}}$ are highly correlated with March MLD: during years with deeper winter mixed layers, warming of the layer by $C_{\text{diff}}$ and $C_{\text{bol}}$ is larger. The horizontal contribution to $C_{\text{diff}}$ dominates over the vertical contribution (not shown), suggesting that the correlation between $C_{\text{diff}}$ and March MLD is attributable to lateral diffusion re-stratifying a newly formed mixed layer rather than vertical mixing resulting from convective instability (although this term does play a role when MLD variances are large). The correlation with $C_{\text{bol}}$ is likely the result of increased stirring by the Gent and McWilliams (1990) scheme when isopycnal slopes increase.

To determine which terms play the largest role in creating $H$ anomalies, we now consider time-integrated budgets. As the volume of each region is quite different, instead of considering the heat content, which will depend on the volume of the box, we choose to consider budgets of the average temperature over the layer from the surface to $-D$, $T = H(p_o C_p V)^{-1}$, where $V$ is the volume of the box. Dividing Eq. (2) for the $H$ budget by $p_o C_p V$ and integrating in time yields

$$\int_0^t H_t \, dt = \int_0^t C_{\text{adv}} + C_{\text{diff}} + C_{\text{bol}} \, dt.$$

Similarly, we divide the equations for $C_{\text{lin}}$, $C_{\text{bol}}$, $C_g$, $C_{\text{Ek}}$, $C_{\text{err}}$, and $C_{\text{loc}}$ by $p_o C_p V$ and integrate in time to yield $T_{\text{lin}}$, $T_{\text{bol}}$, $T_g$, $T_{\text{Ek}}$, $T_{\text{err}}$, and $T_{\text{loc}}$ respectively.

The left panels of Fig. 14 show time series of the terms in the $T$ budget in each of the three regions. As the quantities plotted are cumulative sums of fluxes, they start at zero. Since the mean monthly climatology has been subtracted from the fluxes, the cumulative sums tend to come back to near zero at the end of the time series (although this is not a constraint). Since budgets for $T - T_o$ are computed by cumulatively summing fluxes, there is the potential that any errors introduced by approximating $T_{\text{lin}} \approx T_{\text{Ek}} + T_g$ will accumulate with time. However, $T_{\text{err}} = T_{\text{lin}} - T_{\text{Ek}} - T_g$ is entirely negligible in the subtropical and subpolar gyres. In the Gulf Stream region, $T_{\text{err}}$ is small but not entirely negligible, particularly on longer time scales. The terms $T_{\text{err}}$ and $T_{\text{bol}} + T_{\text{diff}}$ are highly anticorrelated (correlation is $-0.90$), which seems to suggest that $T_{\text{err}}$ reflects physical processes, such as ageostrophic transports resulting from inertial effects, rather than numerical errors.

The right panels of Fig. 14 show the magnitude of the coherence between $T - T_o$ and $T_{\text{loc}}$ and $T_{\text{loc}} + T_g$ in each region. The time series of $T - T_o$ and $T_{\text{loc}}$ are highly coherent on all time scales in the subtropical gyre. In contrast, the coherence between $T - T_o$ and $T_{\text{loc}}$ abruptly decreases for periods longer than 1–2 yr in the Gulf Stream region and subpolar gyre, indicating the importance of ocean dynamics (primarily $T_g$ in the Gulf Stream region and $T_{\text{diff}} + T_{\text{bol}}$ in the subpolar gyre).
In summary, the main results from the time series and coherences plotted in Fig. 14:

(i) In the interior of the subtropical gyre, \( T_{\text{loc}} \) explains the majority of the variability of \( T - T_o \) on all time scales (91% of the total variance of \( T - T_o \) is explained by \( T_{\text{loc}} \)).

(ii) In the Gulf Stream region, \( T_g \) becomes important in setting \( T - T_o \) for periods longer than about 1 yr and \( T_g \) and \( T_Q \) are highly anticorrelated (correlation is \(-0.90\)). These results suggest that on interannual time scales \( T \) anomalies are formed by geostrophic heat transport convergence and damped by air–sea
heat fluxes, but we acknowledge the difficulty of using correlations to imply causation.

(iii) In the subpolar gyre, $T_{\text{diff}}$ and $T_{\text{bol}}$ also contribute to variability of $T$, particularly on the longest time scales resolved by the ECCO estimate. Here, $T_Q$ and $T_{\text{bol}} + T_{\text{diff}}$ are highly anticorrelated, particular at low frequencies (correlation is $-0.82$).

Computing budgets over $D^*$ instead of $D$ leads to no significant changes over the subtropical gyre and Gulf Stream region and quantitative rather than qualitative changes over the subpolar gyre. As expected, the contribution of $C_{\text{diff}}$ to $H^*_t$ is smaller but not entirely negligible (see Fig. 12, bottom panels). Correlations between $C_{\text{diff}}$ and March MLD are reduced from 0.82 (for $D$) to 0.57 (for $D^*$). Despite the smaller role of diffusive and bolus fluxes, $T_{\text{diff}} + T_{\text{bol}}$ still plays a role in the budget of $H^*$ at the lowest frequencies resolved in the ECCO estimate (see Fig. 14, bottom right panel).

b. Time scales of variability

In this section we determine the dominant time scale of variability in each of the three regions defined in Fig. 11a. Furthermore, we discuss how our analysis of the terms that contribute to $H$ variability (section 5a) provides insight into the reason for the different dominant time scales of variability in each region.

Since the power spectra of $H$ in each region are red at high frequencies and flatten out at low frequencies (see Figs. 11d,f,h), we fit a model of the form

$$P(f) = \frac{A}{f^n + \lambda^2}$$  \hspace{1cm} (10)

to the spectra $P(f)$, where $f$ is frequency, $n$ is the slope of the red portion of the spectrum, $A$ determines the power level of the spectrum, and $\lambda$ controls the frequency and period at which the spectrum flattens out. The dominant frequency [obtained by setting $\partial(fP)/\partial f = 0$; see Zangvil 1977] is

$$f^* = \left(\frac{\lambda^2}{n - 1}\right)^{1/n}.$$  \hspace{1cm} (11)

We solve for $n$ by finding the slope of the linear fit to $x = \ln f$ and $y = \ln P(f)$ for the red portion of the spectrum ($f > 0.5$ cpy) and then use a nonlinear fit to solve for $A$ and $\lambda$. Values of $n$, $\lambda$, and $\tau = 1/f^*$ in each region are given in Table 1. We find that $n = 2$ in the subtropical gyre and Gulf Stream region and $n > 2$ in the subpolar gyre, indicating that the subpolar gyre has relatively more variance at low frequencies and less variance at high frequencies than the other two regions. The dominant time scale $\tau$ is shortest in the sub tropical gyre and longest in the subpolar gyre. Values of $\tau$ found via our fit are quite similar to the time scale of the peak on a variance conserving power spectra [plot of $\{\ln f, P(f)\}$; not shown].

As discussed in section 5a, the majority of the $H$ variance in the subtropical interior can be explained by local (air–sea heat flux and Ekman) forcing. This result suggests that $H$ anomalies can be modeled by a first-order autoregressive process,

$$\frac{\partial H}{\partial t} = F_T - \lambda H,$$  \hspace{1cm} (12)

where $F_T$ is the atmospheric forcing anomaly (taken to be white noise with power level $A$) and $\lambda > 0$ parameterizes feedback and damping processes (see, e.g., Frankignoul and Hasselmann 1977; Frankignoul 1981). In Frankignoul (1981) $\lambda^{-1}$ is the damping time scale of SST anomalies, whereas in our case $\lambda^{-1}$ is the effective damping time scale, which takes into account the enhanced persistence attributable to reemergence of SST anomalies (Deser et al. 2003; de Coëtlogon and Frankignoul 2003), as discussed in the Introduction. The power spectrum of $H$ resulting from Eq. (12) is given by setting $n = 2$ in Eq. (10), and the time scale with maximum variance is $\tau = \lambda^{-1}$. Hence, the fact that we find $n = 2$ when we fit Eq. (10) to the power spectrum of $H$ confirms that Eq. (12) is a realistic model of $H$ variability in the subtropical gyre.

We can utilize our estimate of $\tau$ in the subtropical gyre to estimate the damping parameter (see Frankignoul et al. 1998),

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where $D_{st} = 203$ m is $D$ averaged over the subtropical gyre. In the classic Frankignoul and Hasselmann (1977) model, $\alpha$ quantifies the magnitude of the air–sea heat flux response to SST anomalies. In our model, since $H$ anomalies may be isolated beneath the seasonal thermocline during part of the year, $\alpha$ is a bulk measure of the damping of $H$ anomalies (averaged over the course of the year and over depth $D$). We find that $\alpha = 11 \text{ W m}^{-2} \text{ K}^{-1}$, which is somewhat smaller than the canonical damping parameter of 20 W m$^{-2}$ K$^{-1}$ estimated by Frankignoul et al. (1998). Our smaller value of $\alpha$ likely reflects the fact that $H$ anomalies are isolated beneath the seasonal thermocline during the summer months and are thus not damped by air–sea heat fluxes.

**Table 1.** Values of $n$ and $\tau$ in each region obtained by fitting Eq. (10) to spectra of $H$ in each region.

<table>
<thead>
<tr>
<th>Region</th>
<th>$n$</th>
<th>$\lambda$ (cpy)</th>
<th>$\tau$ (yr)</th>
<th>95% confidence</th>
</tr>
</thead>
<tbody>
<tr>
<td>Subtropical</td>
<td>2.0</td>
<td>0.43</td>
<td>2.4</td>
<td>2.2–2.6</td>
</tr>
<tr>
<td>Gulf Stream</td>
<td>2.0</td>
<td>0.30</td>
<td>3.3</td>
<td>3.0–3.8</td>
</tr>
<tr>
<td>Subpolar</td>
<td>2.4</td>
<td>0.17</td>
<td>5.2</td>
<td>4.9–5.7</td>
</tr>
</tbody>
</table>
[The impact of allowing α to vary seasonally is discussed in Deser et al. (2003).] However, as demonstrated by Frankignoul and Kestenare (2002), in the North Atlantic α ranges between 10 and 35 W m⁻² K⁻¹, so differences in α could also reflect averaging over a different geographical area (the subtropical gyre in our case and the entire North Atlantic in Frankignoul et al. 1998). It is also important to point out that Frankignoul et al. (1998) and Frankignoul and Kestenare (2002) estimate α via lagged covariances between SST and air–sea heat flux observations, whereas we have simply estimated the value to α that is consistent with the observed values of $D_\alpha$ and τ.

Variability of $H$ in the Gulf Stream region appears to be forced by geostrophic heat transport convergences and damped by air–sea heat fluxes. Thus, ocean dynamics play a role in setting the dominant time scale and hence the predictability of $H$ anomalies (and likely SST anomalies as well) over the Gulf Stream region. While we have not explored the dynamics responsible for these geostrophic heat transport convergence anomalies in detail, one possible hypothesis is that changes in the Gulf Stream path are the result of stochastic atmospheric forcing integrated over Rossby wave characteristics (Frankignoul et al. 1997; Marshall et al. 2001a; Picu and Ponte 2012a). The theoretically predicted baroclinic pressure spectrum for stochastically forced Rossby waves has a slope of −2 at high frequencies and flattens out at low frequencies (longer than the Rossby wave crossing time scale) to a level that increases quadratically with distance from the eastern boundary. [See Eq. (16) in Frankignoul et al. (1997) for the full equation for the spectrum.] The fact that we find $n = 2$ supports the hypothesis that wind forced baroclinic Rossby waves play a role in setting geostrophic transports and hence $H$ variability in the Gulf Stream region. The value for τ of 3–4 yr is a bit shorter than the expected (decadal) Rossby wave crossing time at a latitude of approximately 40°N (Sturges et al. 1998; Marshall et al. 2001b; Wunsch and Heimbach 2009). However, 1) mean flows and resulting potential vorticity gradients may act to increase the phase speed (Tulloch et al. 2009) and 2) much evidence suggests that, because of damping and instability (Isachsen et al. 2007), Rossby waves may reflect stochastic forcing over a more localized region (Osychny and Cornillon 2004). Addressing these questions is beyond the scope of this work, but in future work we plan to test if a linear Rossby wave model forced by observed wind (and perhaps buoyancy) forcing can explain the observed geostrophic variability in the Gulf Stream region.

Although local atmospheric forcing explains over 70% of the variance of $H_t$ in the subpolar gyre, diffusion and bolus transports exhibit significant low-frequency variability and thus play a large role in the $H$ budget in the subpolar gyre. Low-frequency variability of diffusive and bolus convergences is related to low-frequency variability of the wintertime MLD in the subpolar gyre. The steeper slope ($n > 2$) in the subpolar gyre is consistent with the dominance of slow processes, such as diffusion and mixing, which predominantly lead to variance at low frequencies.

6. Conclusions

This paper uses an ocean state estimate (ECCO) to quantify the upper-ocean heat budget in the North Atlantic on intraannual–interannual time scales. We introduce three novel techniques for viewing upper-ocean heat budgets:

(i) The heat budget is integrated over the maximum climatological mixed layer depth, which varies spatially. This method gives results that are relevant for explaining SST variability (on interannual time scales) while avoiding strong contributions from vertical diffusion and entrainment and thus simplifying the analysis.

(ii) Advecive heat transport convergences are separated into Ekman and geostrophic convergences, a technique that is successful away from boundary regions.

(iii) Air–sea heat fluxes and Ekman advection are combined into one local forcing term.

The central result of our analysis is that over large swaths of the North Atlantic, including the subtropics and the subpolar gyre, the tendency of $H$ is predominantly explained (>70% of variance) by local (air–sea heat flux and Ekman) forcing. In contrast, local air–sea heat fluxes (commonly used to determine the importance of local atmospheric forcing) alone explain only about 50% of the variance in these regions.

Based on the distinct dynamics of $H$ variability, we separate the North Atlantic into three regions (the interior of the subtropical gyre, the Gulf Stream, and the interior of the subpolar gyre) and present a detailed analysis of the terms that are important in the $H$ budget on various time scales. Results from our regional analysis are presented in two distinct and complementary ways: 1) consideration of fluxes contributing to $H_t$ and 2) temporally integrated budgets of $H$. We find the following:

(i) In the interior of the subtropical gyre, local forcing explains the majority of the variance of both $H_t$ and $H$ on all time scales resolved by the ECCO estimate ($1/6 \leq \tau \leq 9.5$ yr).
(ii) In the Gulf Stream region, geostrophic heat transport convergences play an increasingly important role in the $H_t$ budget on time scales longer than about $1/2$ yr. Analysis of temporally integrated budgets show that both changes in geostrophic convergences and local air–sea heat fluxes play a leading-order role in the $H$ budget. Geostrophic convergences and air–sea heat fluxes are strongly anticorrelated, which is consistent with the hypothesis that at low frequencies $H$ anomalies are forced by geostrophic convergences and damped by air–sea heat fluxes.

(iii) Geostrophic transports, diffusion, and bolus transports play a role in $H_t$ variability on time scales longer than about 1 yr in the interior of the subpolar gyre. Annual average diffusive and bolus transports are highly correlated with variability in wintertime mixed layer depths. Temporally integrated budgets highlight the importance of diffusion and bolus transports, since these terms exhibit substantial low-frequency variability.

Our analysis is complementary to previous studies attempting to determine the processes that are important in setting SST and UOHC variability in the North Atlantic. In some sense our study bridges the gap between theoretical studies, such as the null hypothesis (Hasselmann 1976; Frankignoul and Hasselmann 1977) or Rossby wave models (Frankignoul et al. 1997; Marshall et al. 2001a), and numerical studies utilizing ocean data or GCM output. Here, we quantitatively test the null hypothesis to determine in what regions and on what time scales local atmospheric forcing can explain observed UOHC variability. We find that atmospheric (air–sea heat flux and Ekman) forcing can explain 91% of the variance of UOHC in the interior of the subtropical gyre on time scales resolved by ECCO ($<9.5$ yr). Additionally, our finding that geostrophic heat transport convergences are important for UOHC variability along the Gulf Stream path is consistent with the idea that variability on the western boundary is in part determined by Rossby wave dynamics (Frankignoul et al. 1997; Marshall et al. 2001a), although here we do not explicitly determine the origin of the geostrophic variability.

A complementary technique for understanding the relative roles of atmospheric forcing and ocean dynamics in setting SST and UOHC is to compare the SST and UOHC variability in an ocean mixed layer model to that of a fully coupled GCM. Using this technique Seager et al. (2000) argue that the majority of wintertime SST variability observed during the last four decades can be explained as a (local) passive response to atmospheric forcing. Their assessment regarding the importance of local atmospheric forcing is quite similar to our results. Their results differ from ours in that they find Ekman transports to be important only in the subpolar North Atlantic, whereas we find them to be important in the subtropics as well, and they do not isolate the Gulf Stream as a region where geostrophic advection is important.

It is not easy to compare our results directly to the numerous studies that consider numerical UOHC budgets because of the different definitions of UOHC and regions studied. For example, Grist et al. (2010) use an eddy-permitting ocean model to show that air–sea heat fluxes play only a minor role in setting the depth integrated heat content in the subtropical and subpolar North Atlantic but are comparable in magnitude to the advective heat transport convergence in the tropical North Atlantic. The dominant role of advective heat transport convergences compared to air–sea heat fluxes is likely the result of taking an integral over the full ocean depth rather than the near surface ocean. Our results are consistent with the findings of Dong and Kelly (2004) and Dong et al. (2007) that interannual UOHC anomalies (defined as the integral over the top 400 m) in the Gulf Stream region are forced by changes in ocean geostrophic advection and damped by air–sea heat fluxes.

A common statistical technique for assessing the relative roles of atmospheric forcing and ocean dynamics is computing correlations between winds/air–sea heat fluxes and SST on various time scales. Kushnir (1994) utilizes SST and sea level pressure data to argue that, while the atmosphere forces SST anomalies on interannual time scales, ocean dynamics plays a role in setting SST on decadal time scales. In contrast, Deser and Blackmon (1993) do not find a difference between the SST–wind relationship on different time scales and suggest that SST is the passive response to atmospheric forcing on all time scales resolved in their study (90 years of data). Gulev et al. (2013) utilize SST and air–sea heat flux estimates to argue that in the midlatitude North Atlantic (defined as $35^\circ$–$50^\circ$N) surface turbulent heat fluxes are driven by the ocean and may force the atmosphere on time scales longer than 10 years. While our study does not resolve decadal time scales and thus cannot directly be compared to these results, our work highlights the fact that the time scale for which ocean dynamics becomes important in setting SST and UOHC depends strongly on the region considered. Our analysis also provides dynamical insight into the reason why dominant time scales of UOHC variability are different in each region. The dominant time scale of UOHC variability appears to be set by the damping time scale of UOHC anomalies in the subtropical gyre, geostrophic
variability (perhaps the time scale for Rossby waves to cross the basin) in the Gulf Stream region, and eddies and diffusive processes (likely related to deep convection and restratification) in the subpolar gyre.

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