On the time variability of the net ocean to atmosphere heat flux in midlatitudes, with application to the North Atlantic basin.

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*Submitted to Q. J. R. Meteorol. Soc. on October 4th 2002*

*Revised on March 24th 2003*

**Keywords:** Air-sea interactions, climate variability, North Atlantic Oscillation
Abstract

A new diagnostic to investigate the role of ocean dynamics on midlatitude air - sea interactions is presented and tested against observations. It is based on the analysis of the time variability of the net ocean to atmosphere heat flux \( F_S \).

A hierarchy of air - sea interaction models indicate that, in absence of ocean dynamics, the power spectrum of \( F_S \) should be blue at timescales longer than a threshold set by the late winter ocean mixed layer thickness and the sensitivity of \( F_S \) to sea surface temperature anomaly. Comparison of the predicted \( F_S \) spectrum with observations over the North Atlantic shows a good agreement over the subpolar gyre where the deep ocean mixed layer combined with strong stochastic forcing allows large fluctuations in \( F_S \) at decadal and longer timescales. Discrepancies however arise over the Gulf Stream extension region. Here it is suggested that the observed variability of \( F_S \) at timescales longer than a decade is controlled by geostrophic ocean dynamics rather than local atmospheric forcing.

The diagnostic appears as a useful and simple tool to investigate the role of ocean dynamics in the upper ocean heat budget. It is particularly well suited to the analysis of long simulation of coupled ocean - atmosphere models.

One implication of the study for ocean - only numerical simulations is that one can not specify externally the low - frequency variability of \( F_S \). The latter should only arise as a consequence of ocean dynamics.
1 Introduction

The net ocean to atmosphere heat flux $F_S$ is the sum of turbulent (latent and sensible) and radiative components

$$F_S = F_{\text{turb}} + F_{\text{rad}}$$  \hspace{1cm} (1)

In midlatitudes, atmospheric variability creates large-scale patterns of anomalous heat flux through its impact on both $F_{\text{turb}}$ and $F_{\text{rad}}$. During the cold season, changes in surface windspeed and anomalous cold/warm advection associated with the dominant patterns of atmospheric variability lead to a strong modulation of the turbulent heat flux $F_{\text{turb}}$ over the Northern hemisphere (e.g., Cayan, 1992; Alexander and Scott, 1997). During the warm season, changes in cloudiness associated with the anomalous displacement of the storm-tracks impact the radiative component of the heat flux $F_{\text{rad}}$ (e.g., Frankignoul, 1985; Norris, 2000). Thus, the variability of $F_S$ is controlled to a large extent by the variability of atmospheric circulation.

It should be remembered however, that the net ocean to atmosphere heat flux arises solely as a response to a thermodynamic imbalance between the ocean and the atmosphere. In a purely one-dimensional (vertical) framework, without a role for ocean dynamics, the adjustment of the upper ocean to atmospheric variability will lead to an equilibrium state with vanishing net heating / cooling at the surface. Thus, it is expected that at some timescale the variability of $F_S$ is significantly reduced (blue power spectrum)
even though atmospheric state variables (like temperature, pressure) have an essentially white spectrum, or might even be ‘reddened’ by the interaction with the ocean. This has been seen in a variety of coupled models that do not incorporate ocean currents: energy balance models (Barsugli and Battisti, 1998), atmospheric general circulation models (GCMs) coupled to a slab ocean mixed layer (Bladé, 1997).

Ocean dynamics could conceivably sustain a thermodynamic imbalance between the upper ocean and the atmosphere and introduce power in $F_S$ spectra at long timescales. If, for example, cooling of the upper ocean driven by atmospheric variability through air - sea heat flux, is balanced by a warming due to oceanic advection, the variability in net surface heat flux $F_S$ could be enhanced by ocean advection which ‘pulls away’ the SST anomaly from its thermodynamic equilibrium value. This is seen for instance in the simple coupled model of Saravanan and Mc Williams (1998) when oceanic advection becomes sufficiently large compared to damping effects.

It thus appears that a signature of the absence of an impact of ocean dynamics on the upper ocean heat budget is a weak variability of $F_S$ above a certain timescale (blue spectrum), which we will refer as $\tau_o$. To our knowledge, this signature, if present in the literature (see the above discussion), has not been discussed and tested against observational datasets. It is the purpose of this paper to determine what sets the timescale $\tau_o$, and assess how far one can go in interpreting observations of $F_S$ variability with a model of
air-sea interaction which does not include a dynamical ocean. The paper is structured as follows. A model for the variability of $F_S$ is presented in section 2. It is then compared to observations in section 3. Conclusions are offered in section 4.

2 A model for the variability of the net ocean to atmosphere heat flux in absence of ocean dynamics

2.1 A zero order model

As a starting point, let us assume that the atmosphere interacts with a slab ocean of constant thickness $h_o$. In the absence of ocean dynamics, only cooling or warming of the slab can balance fluctuations in the net ocean to atmosphere heat flux $F'_S$ (positive upward)

$$\rho C_p h_o \frac{dT'_m}{dt} = -F'_S$$

where $\rho$, $C_p$ denote the density and specific heat of sea water respectively, primes indicate deviations from the mean climatological seasonal cycle, and $T'_m$ is the sea surface temperature (SST) anomaly. Let us further decompose $F'_S$ as the sum of a SST-dependent part $\gamma T'_m$ and a part $N'$, independent of the latter, thus

$$F'_S = -(N' - \gamma T'_m)$$
The $N'$ component reflects that, in midlatitudes, a large fraction of the variability in atmospheric temperature, specific humidity, cloud cover and wind-speed exists independently of SST anomaly. It is physically associated with intrinsic dynamical processes within the atmosphere, like baroclinic instability or wave mean flow interactions. In absence of further ocean–atmosphere interactions, a simple representation of these processes is to take $N'$ as a stochastic process with a short (a week or so) decorrelation timescale (Frankignoul and Hasselman, 1977; Frankignoul, 1985). On timescales of interest here (interannual and longer), $N'$ is thus assumed to be a white noise forcing.

It must be emphasized that despite the assumption of white noise $N'$, it is not assumed that atmospheric variability, once coupled with the ocean, is white. An atmospheric state variable like temperature or geopotential height anomaly $Z'$ is conceivably modelled according to $Z' = a N' + f T_m'$ where $a$ is a scaling factor and $f$ measures a dynamical SST-feedback on the atmosphere. Depending on the strength of $f$, atmospheric spectra will depart more or less strongly from that of $N$ (e.g., Barsugli and Battisti, 1998; Ferreira et al. 2001). For instance, the slight redness hinted at some observed indices of the North Atlantic Oscillation (NAO) is actually consistent with that expected from a weak interaction with the ocean (Czaja et al., 2003).

Combining (2) and (3), one sees that the heat flux sensitivity $\gamma$ controls the rate of damping of SST anomalies. The precise value of $\gamma$ depends upon
the degree of adjustment of the atmosphere to SST anomaly. In other words, the parameters $\gamma$ and $f$ are actually closely related: the larger the feedback $f$, the weaker the sensitivity $\gamma$, because the larger the adjustment of the atmosphere to a given SST anomaly. Were there no adjustment, bulk-formulae would lead to a strong sensitivity $\gamma \simeq 40 \ Wm^{-2}K^{-1}$ and strong damping of SST anomalies (Haney, 1971; Frankignoul et al., 1998). The atmospheric adjustment to SST anomalies reduces this value somewhat, typically by a factor of about 2 in midlatitudes (Frankignoul et al., 1998; Czaja et al., 2003).

Empirical estimates of $\gamma$ from observations can be found in Frankignoul et al. (1998) and Frankignoul and Kestenare (2002), with typical values around $20 \ Wm^{-2}K^{-1}$.

Equations (2)-(3) provide a prediction for the power spectrum of $F_s'$

$$|F_s|^2 = \frac{\omega^2}{\omega^2 + \tau_o^{-2}} |N|^2$$

(4)

where $|N|^2$ is the power spectrum of the stochastic forcing, $\omega$ denotes angular frequency and $\tau_o$ is a timescale for SST anomaly

$$\tau_o = \frac{\rho C_p h_o}{\gamma}$$

(5)

The separation timescale $\tau_o$ depends solely upon the thickness $h_o$ of the ocean layer and the heat flux sensitivity to SST anomaly $\gamma$. For a mixed layer depth $h_o = 100 \ m$ and sensitivity $\gamma = 20 \ Wm^{-2}K^{-1}$ this amounts to $\tau_o \simeq 8$ months.

Equation (4) indicates that the energy level of $|F_s|^2$ is set by the stochastic
forcing $|N|^2$ at high angular frequency ($\omega \gg \tau_o^{-1}$) but behaves like $(\omega \tau_o)^2$ at angular frequency much lower than $\tau_o^{-1}$. The power spectrum of the net ocean to atmosphere heat flux is thus expected to be blue at periods longer than $2\pi \tau_o$. Figure 1 plots power spectra for different timescales $\tau_o = 0.25$ yr, $\tau_o = 1$ yr, $\tau_o = 5$ yr and $\tau_o = 10$ yr. For a given sensitivity $\gamma$ and level of stochastic forcing $|N|^2$, if the mixed layer is shallow (small $\tau_o$) we see only a weak variability in $F_S$ at decadal timescales whereas if the mixed layer is deep (large $\tau_o$) we observe high amplitude fluctuations on monthly to decadal timescales. Conversely, for a given mixed layer depth $h_o$ and level of stochastic forcing $|N|^2$, the larger the adjustment of the atmosphere to SST (i.e. the smaller $\gamma$ and the larger $\tau_o$), the larger the amplitude of the power spectrum of $F_S$ at low - frequency.

2.2 Effects of entrainment and the seasonal cycle

Some clarifications are needed before comparing the prediction (4) with observations. The oceanic mixed layer depth undergoes a pronounced seasonal cycle in mid - to - high latitudes. For instance, over the Westerly belt region in the North Atlantic (see below), the late winter mixed layer reaches typically 500 m, while being only about 25 m in summer (Levitus and Boyer, 1994). We thus need to be more specific about the depth used in the above predictions. In addition, the mixed layer depth also varies from year to year and the associated change in entrainment could complicate the balance (2).
Figure 1: Model prediction for the power spectrum of the net ocean to atmosphere heat flux for various timescales $\tau_o$, as indicated on the plot, and a given level of stochastic forcing. The power is non dimensional and the frequency is expressed in cycle per year (cpy).
Consider first the effects of the climatological mean seasonal cycle. We assume that the ocean mixed layer thickness $h$ varies seasonally from its shallowest climatological value $h_s$ to its deepest $h_d$, usually reached in early spring and late winter respectively. At depths above $-h$, the temperature is $T_m$ (the SST), while in between $-h_d$ and $-h$ lies the seasonal thermocline. At depths below $-h_d$ (within the permanent thermocline), the temperature is assumed to be constant. The heat budget can then be written:

$$\frac{\partial}{\partial t} \int_{-h_d}^{0} \rho C_p T' dz = -F'_s$$  \hspace{1cm} (6)

where $T'$ is the temperature anomaly. Averaging in time from a given late winter (at time $t = t_d$ when $h = h_d$) to the next ($t_d + t_{yr}$ where $t_{yr}$ denotes a year), we have

$$\rho C_p h_d [T'_m(t_d + t_{yr}) - T'_m(t_d)] = - \int_{t_d}^{t_d+t_{yr}} F'_s dt$$ \hspace{1cm} (7)

since $T'(t_d) = T'_m(t_d)$ and $T'(t_d + t_{yr}) = T'_m(t_d + t_{yr})$. Eq. (7) thus suggests that the annually averaged anomaly in $F_s$ is related to the model (2) with the mixed layer thickness set to its deepest climatological value ($h_o = h_d$), and the annual average estimated from one late winter to the next.

A similar dependence of annually averaged conditions upon late winter mixed layer depth was recently demonstrated by Deser et al. (2003; see also de Coëtlogon and Frankignoul, 2003) when considering extratropical SST anomaly. Note however that their argument about re-emergence (memory of SST anomaly from late winter to the following early winter due to en-
trainment at the base of the seasonally varying ocean mixed layer) is not part of the argument presented here, although re-emergence is present in the above conceptual model (temperature anomaly in the seasonal thermocline are those generated the previous winter). Equation (7) is merely an annual average of the mixed layer heat budget, not an annual average of the various processes governing the damping of SST anomaly through the course of a year.

To test that the model (2) should indeed be used with \( h_0 = h_d \) and the annual average estimated from one late winter to the next, and further assess how entrainment effects impact the simple prediction (4), we now turn to a long simulation (500 years) of a state of the art ocean mixed layer model driven by stochastic atmospheric forcing. The simulation is described in detail in Alexander and Penland (1996). It is sufficient to note here that anomalous fields of windspeed, air temperature and specific humidity were designed from different Markov models and added to a seasonally varying climatology to force the ocean mixed layer model of Gaspar (1988). The latter includes a sophisticated parameterization of entrainment at the bottom of the mixed layer and a diffusive thermocline below the mixed layer.

We compare the simulated and predicted spectrum in Figure 2. Annually averaged \( F_s \) anomalies were first computed for the simulation by starting from the time of deepest mixed layer \( (t_d) \), which, for the simulation coincides with January (i.e., the annual average is from January to January). The
power spectrum of the annual anomalies was then estimated (black curve). The predicted spectra (grey curve) was constructed from (4), using \( h_a = h_d = 120 \, m \) and \( \gamma = 15 \, Wm^{-2}K^{-1} \), which was estimated from the monthly cross covariance function of SST and net surface heat flux anomaly, as described in Frankignoul et al. (1998). The stochastic forcing energy level \( |N|^2 \) was set to that of the simulated spectrum at high frequency (using an average of the monthly spectrum over the frequency band \( 1 - 3 \, cpy \), i.e. periods between 4 months and one year, outside of the frequency range shown in Fig. 2). One observes a good agreement between the two spectra on interannual to interdecadal timescale. The simulated spectrum is blue, with a spectral slope close to \(+2\) for timescales ranging from about 5 to 30 years. This suggests a short timescale \( \tau_0 \) (see Fig. 1), consistent with our estimate \( \tau_0 \simeq 1 \, yr \) (using \( h_d = 120 \, m \) and \( \gamma = 15 \, Wm^{-2}K^{-1} \) - see Eq. (5)). The model (4) however fails to predict the flattening of the spectrum on longer timescales (centuries). This probably reflects the effects of vertical diffusion which, on these long timescales, act to increase the thickness of the oceanic layer interacting with the atmosphere.

In summary, despite the tendency for ocean-atmosphere interactions to slightly redden the spectra of atmospheric state variables like geopotential height (e.g., Barsugli and Battisti, 1998), the absence of ocean dynamics prevents low - frequency (decadal and longer) fluctuations in \( F_s \), unless, (i) the atmosphere interacts with a very deep ocean mixed layer (ii) the
Figure 2: Power spectrum of annually averaged net surface heat flux anomaly simulated by Alexander and Penland (1996) at $50^\circ N - 145^\circ W$ (black curve) and that predicted by the simple model of section 2.1 (grey curve). The frequency is expressed in cycle per year, and the power in $W^2 m^{-4}/cpy$, as computed using the multitaper method (Percival and Walden, 1993). The vertical line indicates a 95% confidence level.
atmosphere adjusts so strongly to SST anomaly that it reduces the sensitivity of the net heat flux, $\gamma$, to values close to zero. We now discuss the form of observed spectra of $F_S$ and compare them with our simple model.

3 Comparison with observations

We focus on two specific regions of the North Atlantic, the Westerly belt ($60^\circ - 15^\circ W/45^\circ - 60^\circ N$, hereafter WB) and the Gulf stream extension ($80^\circ - 40^\circ W/25^\circ - 45^\circ N$, hereafter GS). This choice is motivated by the relatively good data coverage over these areas, and their difference in oceanic conditions, such as late winter mixed layer depth. This should allow us to test the model of section 2 in different parameter regimes. In addition, as shown in various studies (e.g., Cayan, 1992; Marshall et al., 2001), changes in wintertime surface cooling of the ocean in these regions is known to be associated with changes in the North Atlantic Oscillation (NAO), the dominant signal of interannual variability of the Northern Hemisphere atmosphere. This makes these regions of particular relevance to climate variability over the North Atlantic.

A major issue in comparing the prediction (4) to observed $F_S$ spectra is the significant error expected in the estimation of $F_S$ from either atmospheric reanalyses or in-situ oceanographic observations. To take this explicitly into
account in the comparison, we rewrite (4) as

$$|F_S|^2 = \frac{\omega^2}{\omega^2 + \tau_0^{-2}} |N|^2 + n^2 |N|^2$$

(8)

where $n$ is a parameter measuring the uncertainty we allow on the surface heat flux. It is assumed that observational errors are uncorrelated with the true estimate of the surface heat flux and can be represented as a white noise with amplitude proportional to $|N|^2$. For instance, a value $n = 0.5$ postulates a 50% error on $F_S$. The impact of the error term in (8) will be small at high frequency but can significantly alter (and ‘whiten’) the low-frequency part of the blue shape spectrum as $n$ gets closer to unity (see below). In the following we will consider $n = 1$ and $n = 0$, the truth being somewhere between perfect data and a 100% error.

To apply the prediction (8), we further need to estimate the timescale $\tau_0$ and the energy level of stochastic forcing $|N|^2$. In accordance with the previous section, we set $h_o$ to the the late winter mixed layer depth $h_d$ ($h_d = 500 \text{ m}$ for WB and $h_d = 120 \text{ m}$ for GS) observed by Levitus and Boyer (1994). The sensitivity $\gamma$ is roughly known from observations (see section 2.1) but we decided to produce a prediction for weak and strong sensitivity, expecting typical $F_S$ spectra to fall in between those curves. In Figs. 3 (WB) and 4 (GS) these two predictions define an envelope shown as a dashed line, two envelopes being shown, one for $n = 1$ (100 % error on $F_S$, black) and the other for $n = 0$ (perfect data, gray). For each envelope, the lowest curve corresponds to the strongest sensitivity (set to 40 $Wm^{-2}K^{-1}$, with a shorter
$\tau_o$), while the upper most curve corresponds to the weakest sensitivity (set to 5 W m$^{-2}$ K$^{-1}$, with a longer $\tau_o$).

To estimate the level of stochastic forcing $|N|^2$ we have averaged the spectrum of observed monthly $F_S$ anomalies over the frequency band 1 – 3 cycle per year, as in Fig. 2. This was done separately using $F_S$ from the NCEP-NCAR reanalysis (Kalnay et al., 1996) and from da Silva et al. (1994). We then averaged the two estimates to estimate $|N|^2$. The spectrum of both the NCEP-NCAR and the da Silva surface heat flux are shown in Figs. 3 and 4 as continuous black curves$^1$.

Comparison of the observed spectra with the model prediction (either $n = 0$ or $n = 1$) over the WB region (Fig. 3) indicates no major discrepancy, the observed spectrum being within errorbars (indicated on the upper left corner of Fig. 3) of the prediction envelopes. The deep mixed layer in that region leads to a white spectrum for the theoretical prediction over a broad range of frequencies for high and low sensitivities $\gamma$ ($n = 0$, grey envelope), this being even more pronounced when including large errors on $F_S$ ($n = 1$, black envelope). Thus, for this region, the observed, essentially white net heat flux spectrum merely reflects atmospheric variability interacting with

$^1$To maintain consistency with section 2, annually averaged anomalies were first computed as averages from February (March) to the following February (March) for the WB (GS) region ($t = t_d$ of section 2.2), based on the seasonal climatology of mixed layer depth from Levitus and Boyer (1994). However, the results turned out to be insensitive to this choice and calendar annual averaging is used in the following analysis.
Figure 3: As in Fig. 2, but for the power spectrum of observed annually averaged (and linearly detrended) anomalies in $F_S$ over the westerly belt region (from NCEP-NCAR reanalysis over 1949-2001 and da Silva et al. over 1945-93, continuous curves) and the prediction of the simple model for different set of parameters (dashed curves). An approximate errorbar for the observed spectra is indicated on the upper left corner.
Figure 4: As in Fig. 3, but for the Gulf Stream extension region.
a deep ocean mixed layer. This interpretation is further reinforced by the comparison of the timeseries of $F_S$ over the WB region and Hurrell's NAO index (Hurrell, 1995), which vary in-phase from interannual to interdecadal timescales (not shown).

More severe departures from the model prediction however occur for the Gulf Stream extension region (Fig. 4). If observed and predicted spectra (perfect case, $n = 0$, grey envelope) agree within errorbars out to decadal periods, the low-frequency end of the observed spectra (either NCEP-NCAR or da Silva) is significantly higher than that predicted for both large or weak sensitivity (more than one order of magnitude departure at timescales longer than 25 years). Taking into account observational errors suggests that, in the worst case ($n = 1$, black envelope), we should expect a white spectrum even though the mixed layer is sufficiently shallow to in principle lead to a blue spectrum (grey envelope), were ocean dynamics being actually negligible. However, both NCEP-NCAR and da Silva estimates are still above this limit, suggesting a slightly red power spectrum for the observed $F_S$ over the Gulf Stream extension region. In other words, the late winter mixed layer depth is too shallow for local atmospheric forcing to account alone for the observed amplitude of $F_S$ anomalies at low-frequency.

As perhaps the simplest possible mechanism by which ocean dynamics might improve the model of section 2 over the GS region, we study the role
of anomalous Ekman advection $F'_{Ek}$ by modifying (2) as,

$$\rho C_p h_o \frac{dT'_m}{dt} = -F'_S - F'_{Ek}$$

(9)

with

$$F'_{Ek} = \rho C_p (U'_{Ek} \frac{\partial T'_m}{\partial x} + V'_{Ek} \frac{\partial T'_m}{\partial y})$$

(10)

and where $T'_m$ is the climatological mean SST and $(U'_{Ek}, V'_{Ek})$ is the anomalous Ekman mass transport (simply a function of latitude and the anomalous windstress vector). As far as the surface heat flux is concerned, the essential modification introduced by the inclusion of Ekman processes is a possible balance between Ekman advection and surface heating/cooling on timescales when the storage term becomes small, i.e. typically $\tau_o$ (a few years). This balance ($F'_{Ek} \simeq -F'_S$) might, if $F'_{Ek}$ is sufficiently large, result in larger amplitude $F'_S$ anomalies at decadal and longer timescales than accounted for by the model of section 2 and so enhance the power of $F'_S$ at low - frequency.

To check this, we have computed the spectral coherence between $F'_S$ and an estimate of $F'_{Ek}$, both constructed from the NCEP-NCAR reanalysis (Fig 5). One sees that it is only at the highest frequency (periods of about 2 years) that a significant coherence is found, with very small coherence at lower frequency (Fig. 5, top). This is a first indication that the balance $F'_{Ek} \simeq -F'_S$ does not occur at low - frequency. In addition, the phase between $F'_{Ek}$ and $F'_S$ is close to zero for most of the frequency bins rather than being close to $\pm 180$ degrees (Fig. 5, bottom). This reflects that anomalous surface easterlies reduce windspeed and hence turbulent heat flux over the GS region
while simulatenously creating anomalous northward Ekman current. Both act together to warm the upper ocean rather than oppose each other as the balance requires. Finally, $F_{Ek}'$ is typically found to be a factor of about 2 to 3 smaller than $F_S'$ at interannual timescales, with a power spectrum almost one order of magnitude lower than that of $F_S$ at timescales of decades or longer (not shown).

It thus appears very unlikely that anomalous Ekman currents alone could be responsible for the discrepancy between predicted and observed slow changes in $F_S'$ over the GS region, which points to a role for geostrophic ocean dynamics in that region. Overall, our conclusions are consistent with modeling studies of SST variability (e.g., Battisti et al., 1995; Halliwell, 1998; Seager et al., 2000). Halliwell (1998) for example showed that over the Westerly belt region geostrophic ocean dynamics is not needed to simulate the basic structure of interdecadal SST variability, whereas it is crucial in the Gulf Stream extension region.

4 Conclusions

A prediction for the power spectrum of the net ocean to atmosphere heat flux was derived from a simple stochastic model of air-sea interaction, and tested in more elaborate models and observations over the North Atlantic ocean. In absence of ocean dynamics, the spectrum is predicted to show a characteristic timescale $\tau_o$, above which the spectrum is blue (power decreas-
Figure 5: Squared coherence (top) and phase (bottom, in degrees) between annually averaged anomalies of net surface heat flux ($F'_S$) and Ekman heat advection ($F'_E$), averaged over the Gulf Stream extension region. $F'_E$ was first computed from (10) for each month using monthly windstress anomaly and the climatological mean SST seasonal cycle from the NCEP - NCAR reanalysis, then annually averaged. The co-spectrum was averaged over 5 frequency bins (expressed in cycle per year), with a 95% confidence level indicated for the squared coherence as the horizontal line.
ing with timescales) and below which it is white (nearly constant power with timescale). The timescale $\tau_o$ is set by the late winter mixed layer depth $h_d$ and the sensitivity of the net heat flux to SST change $\gamma$.

Predicted and observed surface heat flux spectra are in agreement over the subpolar gyre of the North Atlantic ($60^\circ - 15^\circ W/45^\circ - 60^\circ N$, or Westerly belt region WB). In that region of deep late winter mixed layer depth ($h_d = 500 \text{ m}$), the timescale $\tau_o$ is sufficiently large so that, as observed, the predicted spectrum is white from interannual to interdecadal timescales. Thus over that region, one dimensional air - sea interactions alone are able to create pronounced decadal and longer timescales fluctuations in surface heat flux.

The situation is more complicated over the Gulf Stream extension region ($80^\circ - 40^\circ W/25^\circ - 45^\circ N$, or GS), however, where the late winter mixed layer depth reaches only 120 m. For both large or weak sensitivity $\gamma$, the associated timescale $\tau_o$ is too short to account for the slight redness displayed by the observed spectrum. Thus, over the Gulf Stream extension region, one dimensional air - sea interaction is not sufficient to explain low - frequency fluctuations in surface heat flux. Geostrophic dynamics is thus postulated to be involved in driving SST away from a thermodynamic equilibrium with the atmosphere, thereby allowing the observed changes in surface heat flux at decadal and longer timescales.

It is suggested that analysis of the low - frequency variability of $F_S$ provides a robust way to diagnose the impact of ocean dynamics on the upper
ocean heat budget, particularly in regions of relatively shallow ocean mixed layer. The impact of ocean currents on SST typically represents a modulation of about a factor 2 or 3 of its canonical red power spectrum in certain frequency bands (e.g., Czaja and Marshall, 2001). It is of about one order of magnitude for $F_S$ (going from a blue to a white or red spectrum), which makes the ocean dynamics signature easier to detect.

Although the diagnostic presented here could easily be applied to other oceanic regions than the North Atlantic basin, uncertainties in observed $F_S$ datasets make the diagnostic even more relevant to coupled ocean- atmosphere numerical models. It allows a simple and straightforward alternative ($h_d$ is easily diagnosed from model temperature profile, $\gamma$ is easily estimated from the covariance between SST and $F_S$ anomalies) to the explicit computation of each term of the mixed layer heat budget. Of course, the latter will still be needed to identify the physical mechanisms (horizontal or vertical advection) responsible for the enhanced power of $F_S$ at low-frequency.

One implication of this study is that one can not specify externally the time behavior of $F_S$ in forced ocean-only numerical simulations. The atmospheric control of $F_S$ variability was shown to be limited by the finite heat capacity of the ocean mixed layer and the non zero sensitivity of $F_S$ to SST anomalies. The presence of significant power in $F_S$ spectra at timescales of decades or longer must reflect the impact of ocean dynamics on the mixed layer heat budget. This impact can easily be included by adding an SST-
dependent term to any prescribed low-frequency for $F_S$ (as in (3)), keeping in mind that it implicitly makes an assumption about the atmospheric adjustment to SST anomaly.

Whether the small changes in the heat flux observed at interdecadal timescales over the Gulf Stream extension region (peak to peak amplitude of about 10 $Wm^{-2}$) and attributed to ocean dynamics in this study could significantly impact the atmospheric circulation deserves further study.

**Acknowledgements:** Mike Alexander kindly provided monthly outputs of $F_S$. Discussions with C. Wunsch are gratefully acknowledged. This work was supported by a grant from NOAA’s Office of Global Programs as part of Atlantic CLIVAR.
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