# NOTES AND CORRESPONDENCE

## Transient Atmospheric Response to Interactive SST Anomalies

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ABSTRACT

The transient atmospheric response to interactive SST anomalies in the midlatitudes is investigated using a three-layer QG model coupled in perpetual winter conditions to a slab oceanic mixed layer in the North Atlantic. The SST anomalies are diagnosed from a coupled run and prescribed as initial conditions, but are free to evolve. The initial evolution of the atmospheric response is similar to that obtained with a prescribed SST anomaly, starting as a quasi-linear baroclinic and then quickly evolving into a growing equivalent barotropic one. Because of the heat flux damping, the SST anomaly amplitude slowly decreases, albeit with little change in pattern. Correspondingly, the atmospheric response only increases until it reaches a maximum amplitude after about 1–3.5 months, depending on the SST anomaly considered. The response is similar to that at equilibrium in the fixed SST case, but it is 1.5–2 times smaller, and then slowly decays away.

## 1. Introduction

Most studies with atmospheric general circulation models (AGCM) of the influence of sea surface temperature (SST) anomalies on the atmospheric circulation were focused on the equilibrium response. Recently, we investigated the transient atmospheric response to prescribed SST anomalies in the midlatitudes, using a three-layer quasigeostrophic (QG) model coupled in perpetual winter conditions to a slab oceanic mixed layer in the North Atlantic (Ferreira and Frankignoul 2005, hereafter FF). The first two modes of SST variability were diagnosed from a coupled run and prescribed as fixed anomalous boundary conditions for the model atmosphere. Strong air–sea heat fluxes that would damp the SST anomalies if they were allowed to vary immediately appeared and resulted in an anomalous diabatic heating of the lower atmosphere, which in turn forced a baroclinic response, as predicted by linear theory (Hoskins and Karoly 1981). The initial baroclinic response rapidly modified the transient eddy activity (2–6-days fluctuations) and thus the convergence of eddy momentum and heat fluxes. The latter transformed the baroclinic response into a growing barotropic one, consistent with the mechanism proposed by Peng and Whitaker (1999) and the sensitivity study of Li and Conil (2003). For both prescribed SST anomalies, the atmospheric response resembled the mode that had created them in the coupled run [namely, the North Atlantic Oscillation (NAO) for the dominant SST mode and the eastern Atlantic pattern (EAP) for the second mode]. It was thus associated with a positive feedback and a decrease of the implied heat flux damping.

An interesting feature of the responses is their long adjustment time to the fixed SST anomalies: about 4 months for the NAO-like response and about 2 months for the EAP-like response. In a recent study, Deser et al. (2007) using a full AGCM, but a somewhat different experimental setup, found a 2–2.5-month adjustment time scale for a NAO-like atmospheric response. Although our NAO adjustment time scale is somewhat

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inflated by the simplicity of the QG model, the two studies show that these time scales are comparable to both the persistence of extratropical SST anomalies and the seasonal scale. This suggests that a more realistic assessment of the atmospheric response to SST anomalies should include two elements: 1) a two-way interaction where SST anomalies are allowed to vary and 2) seasonally varying conditions.

In this note, we extend our previous work by performing the response study with an interactive mixed layer and thus taking into account the two-way interaction. We do not attempt to take into account seasonal variations, as our atmospheric QG model cannot efficiently simulate seasonal variations. An interactive mixed layer has been used in previous atmospheric response studies, but in a different context. Drévillon et al. (2003) and Peng et al. (2005) investigated the influence of a two-way coupling in the midlatitudes when an SST anomaly was prescribed in the tropics. Consistent with the reduced heat flux damping expected from the response of the mixed layer to atmospheric perturbations, the atmospheric response was enhanced where the mixed layer was allowed to vary (see Barsugli and Battisti 1998). Sutton and Mathieu (2002) investigated the response of the coupled atmosphere–ocean mixed layer system to a prescribed ocean heat flux convergence. Notably, they observed that the adjusted SST was not collocated with the forcing nor with the anomalous air–sea heat flux. These studies underscore the importance of an interactive mixed layer where SST anomalies are allowed to adjust, but they deal with cases where the mixed layer forcing was sustained.

Here we consider a case where the initial SST is prescribed but no external forcing is present (other than that which created the SST anomaly in the first place). We show that with an interactive mixed layer, SST anomalies are damped by air–sea fluxes, leading to a weaker atmospheric response than in the uncoupled case. The paper is structured as follows. In section 2, we briefly describe the coupled model and the sensitivity experiments. The transient atmospheric response to the first empirical orthogonal function (EOF) SST mode is described in section 3. The EOF2 SST case is presented is section 4. Conclusions are given in section 5.

## 2. Model and procedure

The coupled model is that of FF. The atmospheric component is derived from Marshall and Molteni's (1993) three-layer QG spectral model. It has a realistic geometry (land–sea mask and topography). The QG potential vorticity is solved, in T31 resolution, at the pressure levels 800, 500, and 200 mb. Damping processes include temperature relaxation, Ekman friction, and a scale-selective diffusion of temperature and relative vorticity. The diabatic heating resulting from air–sea heat exchanges is explicitly represented but only over the North Atlantic between 20° and 60°N where the ocean is represented by a slab mixed layer, whose depth is a function of space only and is prescribed from the December–February climatology of Levitus and Boyer (1994). The SST is driven by turbulent air–sea heat flux and Ekman currents. Finally, the model includes time-independent, spatially variable forcing in both the atmosphere and ocean that corrects for the mean effects of neglected processes (e.g., radiative forcing, air–sea exchanges outside the North Atlantic, and SST advection by mean currents).

The coupled run of FF showed that, considering the simplicity of the model, the climatology and variability of both the atmosphere and the North Atlantic SST are satisfactory. The principal modes of low-frequency variability were determined by an EOF analysis using monthly means. The first two EOFs of the 500-mb geopotential height in the North Atlantic sector (15°–75°N, 90°W–10°E) projected onto the whole Northern Hemisphere are shown in Fig. 1 (left). The first mode is very similar to the NAO, the dominant mode in winter, while the second mode is a monopole centered south of Greenland that bears some resemblance to the EAP (Wallace and Gutzler 1981). The first two SST EOFs (right) capture about 75% of the variability. The first EOF is primarily a midlatitude dipole and the second one a monopole centered south of Newfoundland, Canada. The two modes show some similarity with the first two observed modes in the North Atlantic except south of about 25°N (the first observed mode is a tripole), presumably because of the QG approximation in the atmosphere. As in the observations, the NAO forces SST EOF1 while the EAP forces SST EOF2. The heat flux is the dominant forcing, although advection by Ekman currents significantly reinforces it north of 45°N. As in Czaja and Frankignoul (2002), once the SST anomaly is created, the surface heat flux tends to damp it; that is, there is a negative heat flux feedback (see FF for further details).

To investigate the atmospheric response to a SST anomaly in a coupled setup, we analyze a 600-member ensemble of 12-month simulations with the coupled model in which the SST EOF patterns added to the SST climatology of the coupled run are used as initial conditions. The polarity of the EOFs is that shown in Fig. 1, which gave the largest response in FF. To increase the signal-to-noise ratio, their amplitudes are multiplied by 3. The atmospheric initial conditions are independent snapshots (two months apart) taken from the

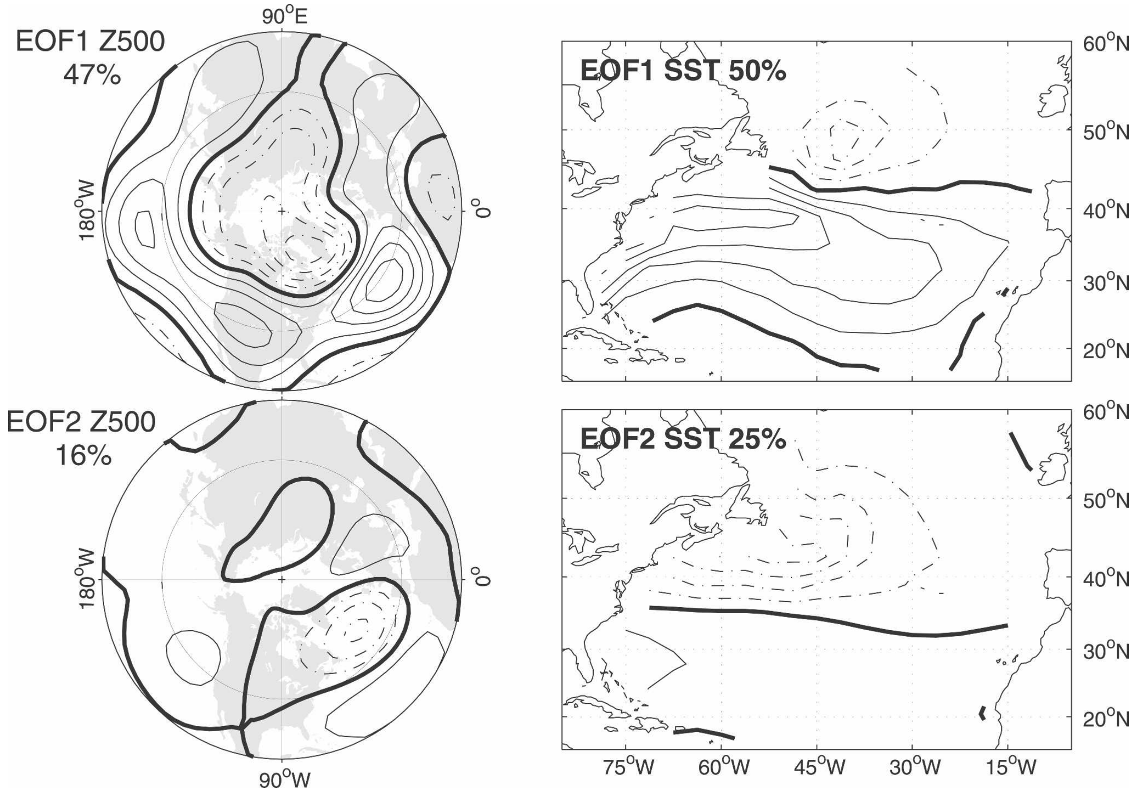


FIG. 1. First two EOFs of (left) geopotential height at 500 mb (m) and (right) SST (K) in the coupled run. The percentage of variance explained is indicated. The principal components have been normalized so that each EOF indicates the typical mode magnitude. Positive and negative contours are solid and dashed lines, respectively; the zero contour is highlighted. The contour interval is 20 m for the atmosphere and 0.2 K for the SST.

coupled run in FF. Weekly means are saved over the first four months and monthly means thereafter.

## 3. EOF1 SST anomaly

The ensemble mean geopotential height response to SST EOF1 [EOF1 $\times$ (+3) in FF] is shown in Fig. 2 at 500 (top) and 800 mb (bottom). The response at 200 mb is similar to that at 500 mb except for a twice-larger amplitude (not shown). As in the uncoupled case, the first-week response is mainly characterized by a baroclinic perturbation over the North Atlantic (cf. with Fig. 5 of FF). The response is small but significant at both levels. Then, the negative anomaly at 800 mb over the North Atlantic weakens and changes sign while, at both levels, the response pattern spreads latitudinally. By month 2, the growing response is equivalent barotropic almost everywhere and has teleconnections over the entire hemisphere. From then on, the response pattern hardly changes. It resembles the positive NAO phase, which created the SST EOF1 in the coupled run, suggesting a positive feedback onto the atmosphere. The amplitude of the response increases up to about month 3.5 (reaching 31 m at 500 mb) and then slowly decreases, reaching half its peak value by month 12. The atmospheric sensitivity, defined as the maximum of the atmospheric response divided by the maximum of the initial SST anomaly, is about 9 m K$^{-1}$ at 500 mb. However, the SST anomaly also evolves. To monitor this evolution, the SST anomaly is spatially correlated with and projected upon the initial SST anomaly (Fig. 3). The spatial correlation slowly decreases from 1 (by construction) to about 0.8, indicating that the SST pattern changes little, but the projection shows that its amplitude decays down to about 20% of its initial value at month 12. This is primarily due to surface heat fluxes, which everywhere damp the SST anomaly (not shown). The decay of the SST anomaly is similar to an exponential decline with a time scale close to the persistence (8 months) of SST EOF1 in the coupled model. However, the initial decay is faster because the atmosphere has not had enough time to adjust and the heat flux damping is very strong at first (see Fig. 5 below).

As the atmospheric response is very similar to that

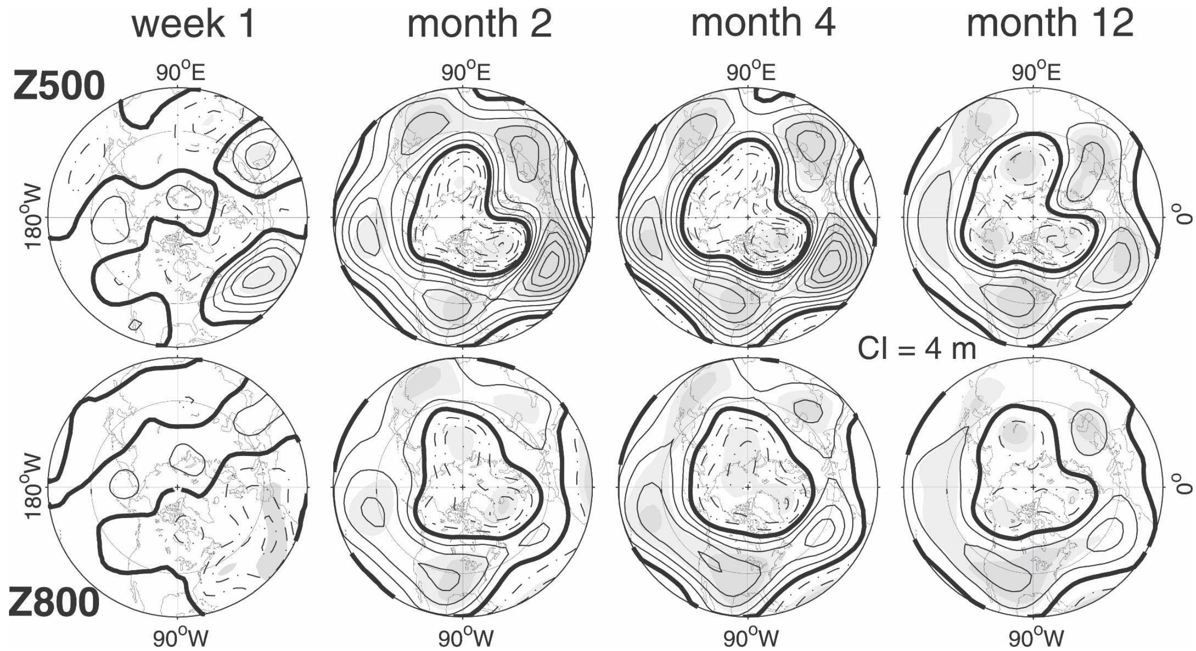


FIG. 2. Transient atmospheric response (geopotential height in meters) to SST EOF1 at (top) 500 and (bottom) 800 mb. Time average intervals are indicated. Positive and negative contours are solid and dashed lines, respectively; the zero contour is highlighted and the contour interval is 4 m. The 5% and 1% levels of significance (two-sided Student's $t$ test) are indicated by light and dark gray shading.

obtained in the uncoupled case, it should obey the same dynamics. The initial baroclinic structure is a direct linear response to the diabatic heating associated with the SST anomaly. As the wind blows over the SST anomaly, the surface heat flux generates an anomalous diabatic heating of the lower atmosphere (while starting to damp the SST anomaly). The initial baroclinic response primarily results from the thermal expansion of the air column due to the anomalous diabatic heating, resembling the time-independent linear response discussed by Hoskins and Karoly (1981) and the initial response in the AGCM experiment of Li and Conil (2003). Although daily outputs needed for the computation of storm track and transient eddy fluxes were not saved, we believe that the transformation of the baroclinic structure into an equivalent barotropic one essentially follows that observed in the uncoupled case through the transient eddy feedback on the large-scale circulation. However, since the SST anomaly is not artificially sustained but responds to surface heat fluxes and decays, the atmospheric response does not reach an equilibrium state. Rather, it peaks after 3–4 months and then decays away with the SST anomaly. This was to be expected since the coupled system slowly returns to a state where, in the ensemble mean, there is no atmospheric or SST anomaly.

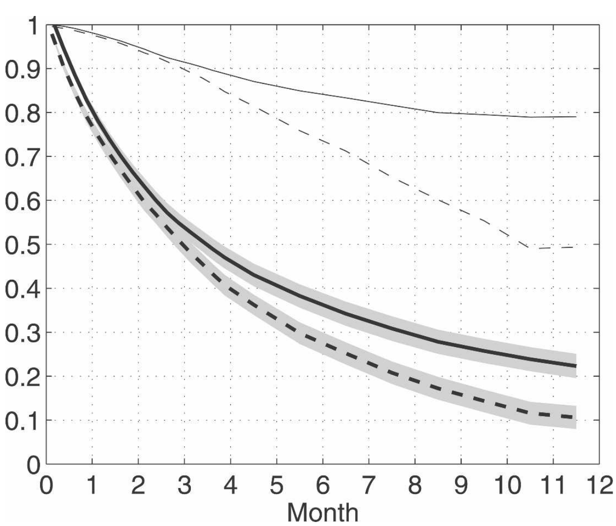


FIG. 3. Spatial correlation (thin) of the SST anomaly with and projection (thick) upon the initial SST anomaly as a function of time for the SST EOF1 (solid) and EOF2 (dashed) experiments. Shading indicates the 95% confidence interval.

As mentioned above, the SST anomaly creates an atmospheric pattern that becomes similar to the one that generates it in the coupled system. To quantify this positive feedback, the atmospheric response is compared in Fig. 4 to the atmospheric pattern that creates SST EOF1 in the coupled run, namely, a positive phase of the NAO very similar to that in Fig. 1 (see Fig. 4 of FF), but multiplied by 3 to take into account the larger

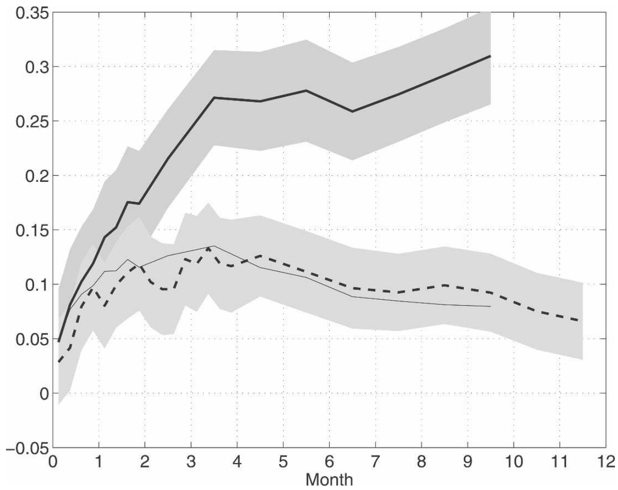


FIG. 4. Projection of the atmospheric response to EOF1 at 500 mb onto the forcing pattern of EOF1 as a function of time. The dashed line corresponds to the coupled case and the solid line to the prescribed SST anomaly case (Fig. 11 in FF). Shading indicates the 95% confidence interval. The thin line is the product of the uncoupled projection (solid line) by the SST anomaly amplitude (thick solid line from Fig. 3).

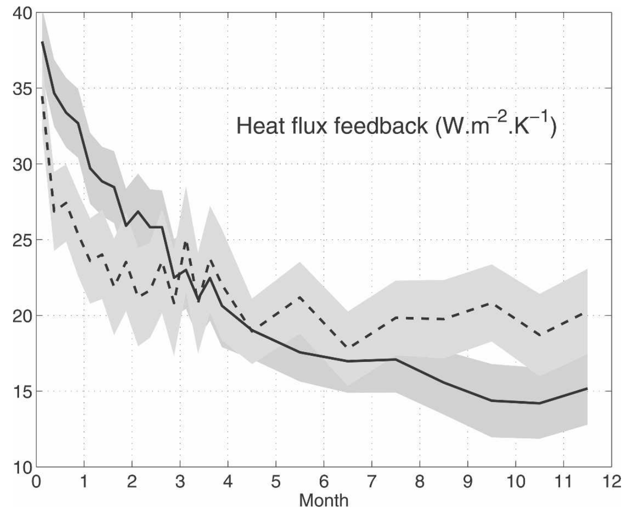


FIG. 5. Heat flux feedback (W m$^{-2}$ K$^{-1}$) as a function of time for the EOF1 (solid) and EOF2 (dashed) experiments. It is estimated by projecting the anomalous surface heat flux onto the SST anomaly and is positive when the former damps the latter. Shading indicates the 95% confidence interval.

SST amplitude. Their spatial correlation becomes rapidly high, exceeding 0.8 after slightly more than a month and 0.9 after 3 months, which confirms the NAO-like character of the response (not shown). Correspondingly, the feedback, defined as the projection of the atmospheric response onto the SST anomaly forcing pattern, increases from zero at the initial time to its maximum of 0.13 at about month 3.5 and then slowly decreases, being still about 0.07 (significantly different from zero at the 5% level) at month 12 (dashed line in Fig. 4). This is compared to the feedback in the uncoupled case (Fig. 11 of FF), which increases to 0.28 after 3.5 months and then stabilizes (solid line). The feedback thus is overestimated by more than a factor 2 in the uncoupled case. However, the coupled feedback is well estimated from the uncoupled case by weighting the latter by the decaying SST anomaly amplitude (thin line in Fig. 4). Hence, if the feedback is estimated from the atmospheric forcing that would have generated the SST anomaly at any given time, rather than at the initial time, the uncoupled feedback is recovered. This is equivalent to making a two-time scale approximation. Consistently, the uncoupled atmospheric sensitivity at 500 mb (20 m K$^{-1}$) weighted by the SST decay at month 3.5 (0.5) is 10 m K$^{-1}$, close to that estimated in the coupled experiment. To a good approximation, the smaller positive feedback in the coupled case thus results from the SST anomaly damping, while involving similar atmospheric dynamics.

As shown in Fig. 5 (solid line), the heat flux feedback (positive for a negative feedback) obtained by projecting the heat flux anomaly onto the SST anomaly rapidly decreases from the initial estimate of 38 W m$^{-2}$ K$^{-1}$, which is close to that in the uncoupled experiment (see Fig. 11 of FF). Since the atmospheric response resembles the positive NAO phase that forces SST EOF1-like anomalies, the negative heat flux feedback diminishes with time, as discussed in FF. However, the decrease is faster and leads to smaller values than in the uncoupled case (15 versus 27 W m$^{-2}$ K$^{-1}$). This is to be expected since SST anomalies adjust to air–sea heat fluxes, reducing the air–sea contrast and thus the heat flux feedback (Barsugli and Battisti 1998).

## 4. EOF2 SST anomaly

The transient atmospheric response to SST EOF2 (multiplied by 3) shows similar features but on a shorter time scale. As before, the initial baroclinic response, with a trough at 500 mb and a ridge at 800 mb above the cold SST pole, rapidly evolves into an equivalent barotropic low at week 2 (Fig. 6). By week 3, the low is slightly shifted northward as a smaller equivalent barotropic high appears above the warm southern SST pole. The pattern hardly changes afterward and is similar to that shown for month 2, resembling the EAP pattern as in the uncoupled case. The response amplitude rapidly increases and reaches 24 m at 500 mb after 1 or 2 months. The atmospheric sensitivity with respect to the initial SST anomaly is then about 9 m K$^{-1}$, similar to

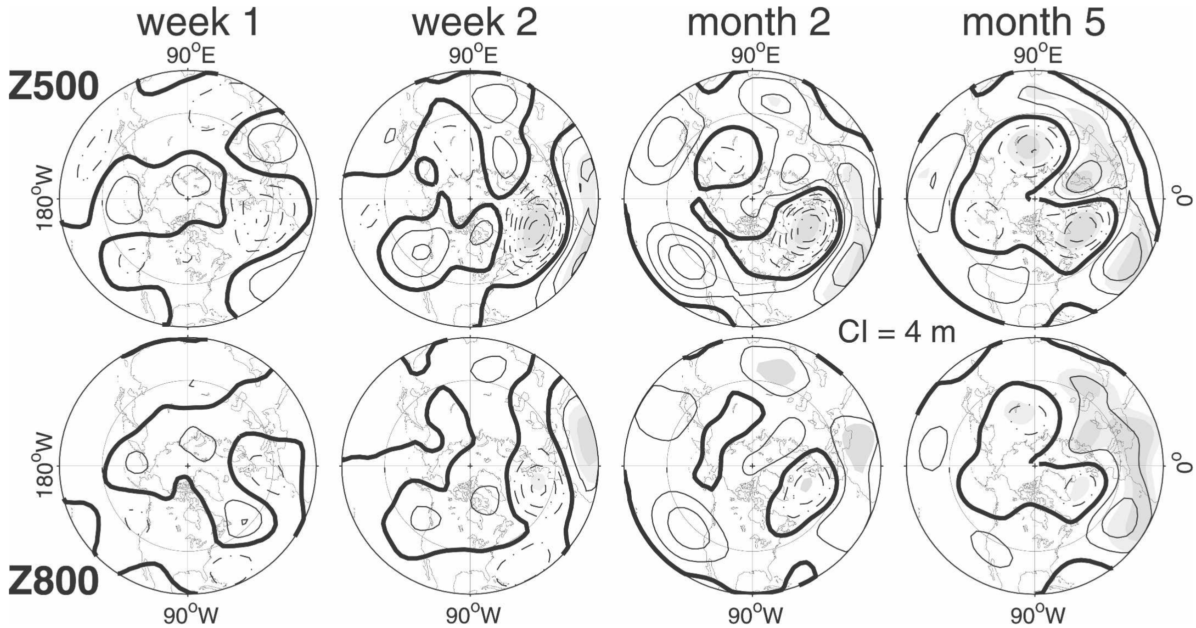


FIG. 6. As in Fig. 2 but for the SST EOF2 experiment.

that obtained with SST EOF1. The signal subsequently decreases, although it is still observed at month 5, and loses significance over the northern Atlantic sector by month 6. This is because the SST anomaly evolves more rapidly than in the SST EOF1 case with a decay time scale of about 5 months (Fig. 3), similar to the persistence of SST EOF2 in the reference coupled run.

The SST anomaly decreases because of the heat flux damping. This is measured, as before, by projecting the heat flux anomaly onto the SST anomaly (Fig. 5, dashed line). The heat feedback is always negative, decreasing from about 35 W $m^{-2}$ $K^{-1}$ (similar to the uncoupled case) to 20 W $m^{-2}$ $K^{-1}$, which is significantly smaller than in the uncoupled case (about 28 W $m^{-2}$ $K^{-1}$) again because of the SST adjustment to the lower tropospheric temperature.

Since the atmospheric response to SST EOF2 is very similar to the EAP pattern that forces it in the coupled setup (except for more a symmetric dipole over the North Atlantic and larger hemispheric teleconnections—see Fig. 4 of FF), it acts again as a positive feedback. The latter is quantified as in section 3 and increases to 0.14 in a month (Fig. 7, dashed line). This is sustained throughout month 2, and then decreases to be indistinguishable from zero by month 7. The maximum feedback is 2/3 of that estimated from the equilibrium response of the uncoupled experiment (solid line), and the feedback estimated by multiplying the uncoupled one by the SST evolution again reproduces the gross behavior of the coupled feedback (thin line).

## 5. Conclusions

We have extended the analysis of FF and studied the transient atmospheric response to SST anomalies in a coupled setup, using a three-layer QG model coupled in perpetual winter conditions to a slab oceanic mixed layer in the North Atlantic. The two dominant modes of SST variability, diagnosed from a coupled run, were prescribed as initial conditions in a 600-member en-

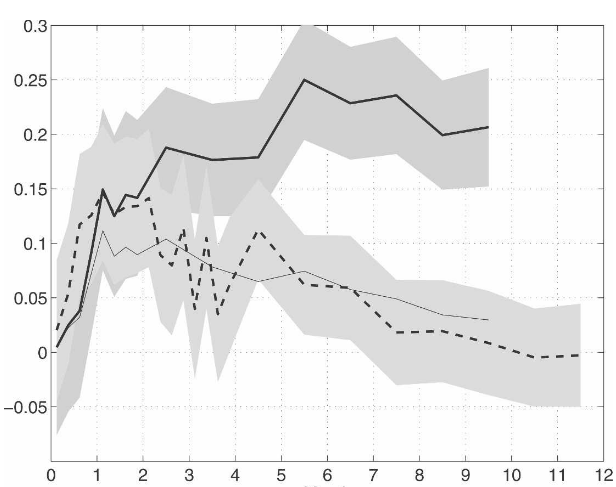


FIG. 7. As in Fig. 4 but for the SST EOF2 experiment. The uncoupled atmospheric feedback (solid line) is shown in Fig. 13 of FF. The thin line is the product of the uncoupled feedback by the SST anomaly amplitude (thick dashed line in Fig. 3).

semble with independent initial atmospheric state and the system allowed to evolve.

The atmosphere initially behaves as in the uncoupled case. The atmospheric response is initially baroclinic and then becomes equivalent barotropic, presumably because of the transient eddy feedback. As the SST decays because of a negative heat flux feedback, the atmospheric response does not reach an equilibrium, but peaks after 3–4 (1–2) months in the SST EOF1 (EOF2) case. The pattern of the response is similar to that which creates the SST anomaly in the coupled run, so there is a positive feedback on the atmosphere, as in FF. However, the feedback, defined as the projection of the response onto the atmospheric signal that generated the initial SST anomaly, is, at its peak, about 2 and 1.5 times smaller than in the uncoupled experiments for SST EOF1 and EOF2, respectively, therefore amounting in each case to 13%–14% of the atmospheric forcing pattern. The initial negative heat flux feedback is decreased by the atmospheric response as well as by the SST adjustment (Barsugli and Battisti 1998), which results in a smaller heat flux feedback than in the uncoupled case even though the atmospheric response is smaller.

Interestingly, the atmospheric feedback is well reproduced by the uncoupled feedback weighted by the SST amplitude evolution. In other words, in our setup, the atmospheric response to the time-varying SST anomaly is similar to that found with the fixed SST provided the SST anomaly decay is taken into account. This suggests that a two time scale approximation remains approximately valid despite the slowness of the atmospheric response. This does not mean that the magnitude of the atmospheric response can be estimated from uncoupled AGCM experiments as the SST evolution cannot be predicted a priori. In our experiments, the SST anomalies behave simply: their patterns change little and they decay on a time scales similar to the EOFs' persistence in the coupled system. This may not hold in other models or settings as, for example, the atmospheric response pattern could differ from the one that forces the SST anomaly, which would in turn lead to rapid changes in the SST pattern and thus the atmospheric response. Ocean dynamics may also modify the SST pattern and/or time scale (see, e.g., the reemergence mechanism investigated by Cassou et al. 2007). Finally, SST anomalies need not be created by the atmosphere in the first place but can originate from oceanic dynamics, in which case its evolution might be complicated (e.g., Sutton and Mathieu 2002 for the case of anomalous ocean heat advection). Hence, in the more general case, the atmospheric response to SST anomalies and the associated feedbacks should be best estimated from coupled sensitivity experiments.

In a comparison between a coupled and an uncoupled run with the coupled model, FF showed that coupling enhances the monthly NAO variance by 14% and the EAP one by 10%. The NAO persistence increased from 2 to 2.8 months while the EAP one remains unchanged. FF attributed the larger NAO changes to the stronger positive feedback onto the NAO found in the uncoupled sensitivity experiments. The coupled experiments discussed here suggest that the feedback is comparable in the two cases and that it is the duration of its impact that matters. The positive feedback acts for a longer period on the NAO than on the EAP because of the longer persistence of SST EOF1 compared to that of SST EOF2, which itself results from the larger NAO persistence as well as climatological properties such as the deep mixed layer at northern latitudes. Thus, in our model, the stronger impact of the coupling on the NAO may simply be due to the properties of the SST anomaly that it forces.

The similarity of the results of FF and Deser et al. (2007), who used a much more realistic AGCM, gives us confidence that our QG model behaves realistically. Still, it has numerous shortcomings and the present results need to be confirmed in a more complex model. Furthermore, in the present study, the QG model was used in perpetual winter conditions. As the mixed layer is very deep and the SST anomalies are only damped by the heat flux feedback, they decay slowly and so does the atmospheric response after it reaches its peak. However, taking into account oceanic damping should somewhat reduce the SST time scales (Frankignoul 1985). Also, the atmospheric response to SST anomalies is very sensitive to the background atmospheric circulation, and hence to time of a year, as shown in sensitivity studies by Peng et al. (1997) and in observations by Czaja and Frankignoul (2002). A better assessment of the atmospheric response to SST anomalies thus requires us to also take into account the seasonal cycle. For this purpose, a more realistic model is needed.

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