Tropical Atlantic SST Forcing of Coupled North Atlantic Seasonal Responses

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ABSTRACT

Recent observational studies reveal that a fall Pan-Atlantic sea surface temperature (SST) anomaly, composed of a horseshoe-like dipole in the North Atlantic and a southern center in the equatorial Atlantic, tends to precede the winter North Atlantic Oscillation (NAO) and its related SST tripole by several months. This study seeks to understand this relationship using large ensembles of atmospheric general circulation model (AGCM) experiments and experiments with the AGCM coupled to a mixed layer ocean (AGCM_ML). The models are forced either by the North Atlantic horseshoe (NAH) or by the tropical SST anomalies over the boreal winter months. The AGCM results show that the NAH anomaly induces a baroclinic response in geopotential heights throughout the winter, with little projection on the NAO. Since the NAH anomaly is ineffective in forcing the wintertime NAO, it cannot account for observations that the NAH SST leads the NAO. In contrast, in the AGCM_ML, the coupled North Atlantic response forced by the tropical anomaly exhibits a strong seasonal dependence. In early winter, the positive anomaly induces a trough east of Newfoundland with a wave train to the northeast, and in late winter the response projects strongly on a negative NAO. Correspondingly, the extratropical SST response features an NAH-like pattern in early winter and a tripole in late winter. These results suggest that tropical Atlantic SST anomalies can significantly influence the coupled extratropical variability. The observed relationship between the fall NAH SST and the winter NAO (or the SST tripole) may be a consequence of persistent forcing of the seasonally varying atmosphere by tropical SST anomalies.

Comparisons with the parallel AGCM results indicate that the largely sign-symmetric NAO responses developed in the AGCM_ML are in part due to active extratropical SST feedbacks. Diagnostic experiments using a linear model further illustrate that, in the absence of transient-eddy feedbacks, an idealized tropical heating induces anomalous flows that are qualitatively similar in early and late winter, with a trough southeast of Newfoundland and a ridge to the northeast. The enhanced seasonality in the SST-induced coupled response likely arises from the seasonal modulation of transient-eddy feedbacks on the heating-forced anomalous flow.

1. Introduction

Given the large thermal inertia of the ocean, it has long been suspected that oceanic forcing of the atmosphere may play a significant role in generating climate variability on seasonal-to-decadal time scales. Detecting extratropical oceanic influences on the atmosphere from observations is, however, difficult since extratropical coupled variability is dominated by vigorous atmospheric forcing of the ocean. The latter is clearly manifested in increased covariances between the atmosphere and the ocean when the atmosphere leads the ocean by a few weeks to a month, as shown in many observational studies (e.g., Wallace and Jiang 1987; Deser and Timlin 1997). Hardly any earlier studies found systematic increases in covariances with the ocean leading the atmosphere. This has led many to question whether the extratropical oceans exert any active feedbacks on the atmosphere, and whether such feedbacks, if they exist, are detectable in observations. Two recent observational studies by Czaja and Frankignoul (1999, 2002) approached the problem differently, and reported the detection of intriguing signals that suggest the existence of significant

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oceanic influences on the North Atlantic Oscillation (NAO).

Czaja and Frankignoul realized that in order to obtain unambiguous evidence of oceanic feedbacks the ocean lead time has to be long enough to exceed the persistence inherent in large-scale atmospheric variability. Using lead-lag maximum covariance analysis (MCA), they examined the leading covarying patterns of monthly 500-hPa geopotential heights (Z500) and SST anomalies over the Atlantic sector (70°N–20°S) for every adjacent three months, with the ocean lead time systematically extended to as long as six months (Czaja and Frankignoul 2002, hereafter CF02). As in earlier studies (see Marshall et al. 2001), the simultaneous winter covarying patterns feature a NAO-like Z500 dipole and a SST tripole in the North Atlantic. For most calendar months, the maximum covariance is reached with the atmosphere leading the ocean by one month. The covariance drops sharply and is not statistically significant with the ocean leading the atmosphere. Interestingly, for Z500 from October to February, the covariance between Z500 and SST remains significant with the SST leading up to two seasons, and it reaches a second maximum when the SST leads by four months (their Fig. 1). An NAO pattern in November–December–January (NDJ) is found preceded by a Pan-Atlantic SST anomaly in July–August–September (JAS), composed of a horseshoe-like dipole in the North Atlantic and a southern center in the equatorial Atlantic (their Fig. 2). This relationship is interpreted as reflecting the forcing of the winter NAO by the fall SST anomaly persisted into the winter. By repeating the MCA with only midlatitude or tropical SST anomalies, CF02 further suggest that the North Atlantic horseshoe (NAH) and the tropical anomalies are not strongly correlated, and that both may persist from the fall and force the winter NAO.

It is known that winter NAO is accompanied by a SST tripole in the North Atlantic, with the latter believed to be primarily a consequence of NAO-related surface heat fluxes (Cayan 1992; Seager et al. 2000; Marshall et al. 2001). Recent modeling studies, however, show that the SST tripole can also induce a NAO-like atmospheric response (Rodwell et al. 1999; Sutton et al. 2001; Peng et al. 2002, 2003). This potential positive feedback between the NAO and the SST tripole may organize coupled low-frequency variability (e.g., Latif and Barnett 1994; Czaja and Marshall 2001). Since the fall NAH anomaly and the winter SST tripole share certain common features, their relationship may be interpreted in at least two different ways. One is to consider the NAH as the "forcing" SST pattern that induces a NAO in winter and in turn is modified by the NAO to evolve into the tripole, as hypothesized by CF02. The effectiveness of the SST tripole in forcing the NAO would then derive from its projection on the NAH SST. This implies that the NAH anomaly should be more effective in forcing the NAO than is the SST tripole. Conversely, the SST tripole could be the more effective "forcing" for the NAO, and the NAH anomaly may precede the tripole without one necessarily evolving into the other. This would be the case if both the NAH and the tripole were induced by a third party, such as a tropical SST anomaly.

In this study we examine the validity of these hypotheses through systematic model experiments that determine the influences of both the NAH and the tropical parts of the fall Pan-Atlantic SST anomaly on the winter NAO. We mainly address two questions: (i) Is the NAH anomaly more effective in forcing the NAO than the SST tripole? (ii) Is the tropical Atlantic anomaly effective in forcing the NAO and its related SST tripole? And, if so, what is the role of extratropical oceanic feedbacks?

SST-induced extratropical responses are in general strongly modulated by intrinsic atmospheric variability (Peng and Robinson 2001; Hall et al. 2001; Deser et al. 2004). The forcing of the NAO by SST anomalies can be well simulated only if the NAO is realistically represented in the internal model variability. The atmospheric general circulation model (AGCM) used in our recent studies on the SST tripole has a good representation of the NAO, and the model produces a well-defined NAO response to the tripole in late winter, by effectively engaging transient-eddy feedbacks (Peng et al. 2002, hereafter PRL02; Peng et al. 2003, hereafter PRL03). Building on these studies, we now use the same AGCM to examine the effects of the NAH anomaly on the NAO, so that the model responses to the NAH and the tripole can be directly compared.

To investigate the influence of the tropical anomaly on the coupled extratropical variability, the AGCM is coupled to a slab mixed layer ocean north of the Tropics (AGCM_ML). Large ensembles of AGCM_ML and AGCM experiments forced by the tropical anomaly are conducted to determine the nature of the tropically forced response with or without extratropical oceanic feedbacks. Drevillon et al. (2003) recently conducted a related study with coupled and uncoupled model experiments, and their results indicate that tropical Atlantic SST may affect the NAO more strongly in a coupled system. They however used small (15-member) ensembles of experiments, and the midwinter NAO dipole in their models appears to be rotated off its observed meridional axis. Using models with a better NAO representation, and also much larger ensembles of experiments forced by a tropical anomaly fixed throughout the boreal winter, we investigate whether a robust NAO response may indeed be induced, and in particular its seasonality.

We present in this paper the key results from our model experiments, which demonstrate the impact of the NAH SST and the tropical SST anomalies on the NAO. The model results provide a new explanation for the observed relationship between the NAH SST and the NAO (or the SST tripole). The paper is organized as follows: section 2 describes the model experiments and the data analyses, section 3 presents the model results, and section 4 provides a summary and some discussion.

2. Methodology

The main results of this study are based on large ensembles of AGCM and AGCM_ML experiments. Diagnostic experiments with a linear baroclinic model are also performed to illustrate the difference between the SST-induced coupled response and the heating-forced linear response. Objective data analyses [i.e., MCA and empirical orthogonal function (EOF)] are conducted to obtain the SST anomalies used to force the models and to compare the observed and simulated variability in the NAO. The model experiments and the data analyses are described below.

a. Model experiments

1) AGCM EXPERIMENTS

As described in PRL02, the AGCM is a version of the operational seasonal forecast model we obtained from the National Centers for Environmental Prediction (NCEP) in 2000. The model is configured with a horizontal resolution of a T42 spectral truncation and 28 vertical levels. To determine the influence of the NAH or the tropical SST anomaly on the NAO over the boreal winter, ensembles of 8-month (September–April) model runs are performed with the SST anomaly either added to or subtracted from the climatological SST seasonal cycle, consistent with the experiments in PRL02. The ensembles are formed by initializing the runs with the NCEP reanalysis data of different dates from 1–5 September 1980–99. The 100-member control ensemble forced with the climatological SST is the same as that made by PRL02. For the NAH SST anomaly, two 40-member ensembles of experiments are made. Since the response from these experiments exhibits little resemblance to a NAO, no further runs are conducted, as the response is unlikely to change into a NAO with larger ensembles. For the tropical anomaly, two 100-member ensembles of runs are made for comparison with the AGCM_ML results. The model response is determined as the ensemble-mean difference between the positive and the negative SST-forced runs, or between the forced and the control runs. A student t test is used to determine the statistical significance of the response, considering each member of the ensemble as independent. All model results are based on twice-daily model outputs.

2) AGCM_ML EXPERIMENTS

To determine the influence of the tropical anomaly on the coupled North Atlantic variability, the AGCM is coupled to a slab mixed layer ocean of 50-m depth (AGCM_ML), from 10°N up to the climatological ice-boundary in both the Atlantic and the Pacific. This simple coupled system may be considered as a first-order representation of the extratropical air–sea interactions (e.g., Lau and Nath 1996; Alexander et al. 2002). For consistency with the AGCM experiments, the model is coupled for the September–April period. A flux correction is applied, based on the atmospheric surface heat flux averaged over 40 AGCM control reruns, to prevent drift in the ocean temperatures. Such drift may arise from various terms, primarily thermal advection, neglected in the mixed layer thermodynamic equation. A 100-member ensemble of AGCM_ML control runs forced with monthly climatological SST south of 10°N is performed. Over the coupled domain, the control ensemble-mean SST (or mixed layer temperature) averaged over September–April exhibits a small difference (about 0.2 K) from the climatological SST used in the AGCM. As a result, the ensemble mean atmospheric variables of the AGCM_ML control runs are nearly indistinguishable from those of the AGCM control. The extratropical thermal coupling, after application of the flux correction, thus has very little effect on the model mean climate. The effects of the coupling on internal model variability are also modest, as shown in section 3.

The extratropical coupled response to the tropical anomaly is examined by performing two ensembles of AGCM_ML runs with the tropical anomaly added to or subtracted from the SST climatology. Results from our preliminary runs indicated that the tropical anomaly tends to induce a NAO-like response in late winter. Hence, we enlarged each ensemble to 100 members to determine better the response, its seasonality, and its asymmetry about the sign of the tropical SST anomaly. The response in both the atmospheric variables and the North Atlantic SST is determined as the ensemble mean difference between the tropically forced and the control AGCM_ML runs.

3) LINEAR BAROCLINIC MODEL EXPERIMENTS

The linear baroclinic model (LBM) is the same as that used in PRL03. Briefly, the model is based on the primitive equations, and is configured with a T21 horizontal resolution and 10 equally spaced pressure levels between 950 and 50 hPa. The dissipation time scales for Rayleigh friction and Newtonian damping are set to one day at the lowest level and 7 days above 700 hPa. Basic states for the LBM are obtained by averaging over the AGCM-ML control runs. The LBM response to an idealized heating is determined by its nearly steady solution, reached after 25 days of integration.

b. Objective data analyses

1) MAXIMUM COVARIANCE ANALYSIS

To obtain the fall Pan-Atlantic SST anomaly associated with the winter NAO, as shown in CF02, we repeat their maximum covariance analysis (MCA) between the observed July–September (JAS) SST (70°N –20°S) and the following November–January (NDJ) Z500 (70°N –20°S, 100°W–20°E). Monthly Z500 from the NCEP–National Center for Atmospheric Research (NCAR) reanalysis dataset and the SST from the GISST dataset during 1958–98 (as in PRL02) are used following the same analysis procedure described in CF02, including the removal of the trends and low frequencies.

2) EOF ANALYSES

To illustrate how the NAO is represented in the AGCM and the AGCM_ML, an EOF analysis of the control run monthly Z500 in late winter (February–April) over the North Atlantic sector (20°–90°N and 90°W–90°E) is conducted for the two models. A corresponding analysis is also performed for the observed monthly Z500 of 1948–98, using the NCEP–NCAR reanalysis data.

3. Results

The leading NDJ Z500 and JAS SST covariance maps from our MCA calculations are similar to those shown in CF02 (their Fig. 2). A negative NAO in winter (NDJ) is preceded by a Pan-Atlantic SST anomaly in fall (JAS) with a NAH-like dipole north of 10N and a positive center in the equatorial eastern Atlantic, as shown in Fig. 1. The amplitude of the SST anomaly in Fig. 1 is enhanced, so that it has a maximum about 1.2 K, similar to the amplitude of the SST tripole used in PRL02. This Pan-Atlantic SST anomaly is divided along 10°N into the NAH anomaly and the tropical anomaly used to force the AGCM and the AGCM_ML.

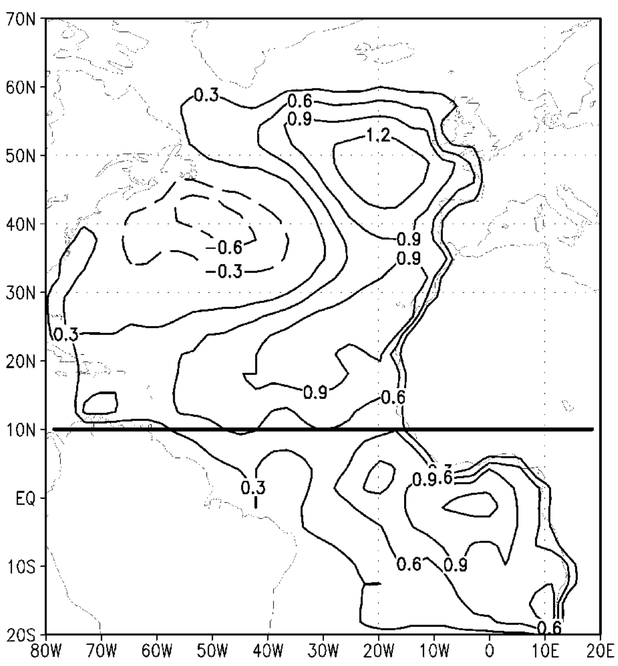

FIG. 1. The amplified fall Pan-Atlantic SST anomaly associated with the winter negative NAO from the MCA of observed JAS SST and NDJ Z500. The NAH and the tropical parts of the anomaly are divided by the solid line along 10°N. The contour interval is 0.3 K. In this and succeeding figures, dashed contours are used for negative values.

Before discussing the model responses to the SST anomalies, we examine how well the observed atmospheric variability, in particular the NAO, is simulated in the AGCM and the AGCM_ML. Figure 2 (left panels) shows the standard deviations of monthly mean Z500 anomalies averaged over late winter (February–April) for observations and for the AGCM and the AGCM_ML control runs. The structure of the observed Atlantic variability is simulated well by both models. The amplitude of the simulated variability is about 10%–20% weaker than the observed. This agreement is consistent with the idea that most atmospheric variability on monthly time scales is intrinsic to the atmosphere. The extratropical thermal coupling results in a slight increase of the variance in the AGCM_ML. The NAO variability is illustrated in Fig. 2 (right panels) by the leading EOF of the monthly Z500 in late winter for observations and for the AGCM and the AGCM_ML control runs. The NAO structure, as indicated by the EOF, is largely similar in observations and in the two models, but the observed NAO explains 2%–3% more of the variance. Again, the NAO variance is slightly increased in the AGCM_ML. Overall, Fig. 2 demonstrates that the intrinsic variability in the two models is very similar, and represents well the observed total variance and the NAO.

a. NAH SST-induced response

To address the question whether the NAH SST anomaly is more effective than the SST tripole in forcing the NAO, two 40-member ensembles of AGCM experiments, forced with positive and negative NAH anomalies north of 10°N (Fig. 1), are conducted. The geopotential-height response averaged over the boreal winter months (October–April), as determined by the ensemble mean difference between the positive and the negative SST-forced runs, exhibits a significant ridge at 250 hPa centered around 30°N over the eastern Atlantic and extending into the Mediterranean (Fig. 3a). The height response at 1000 hPa features a small ridge, above the negative SST anomaly, with troughs to the south and the east and stretching across northern Europe (Fig. 3b). Thus, unlike the NAO response to the SST tripole depicted in PRL03, the NAH SST-induced response is baroclinic and has little projection on a negative NAO. The baroclinic nature of the height response suggests that the response is likely more a direct response to atmospheric heating, in contrast to the more eddy-forced response to the SST tripole (PRL03). Apparently, the NAH SST is less effective than the tripole in stimulating strong eddy feedbacks. This may explain why the response also exhibits less seasonality

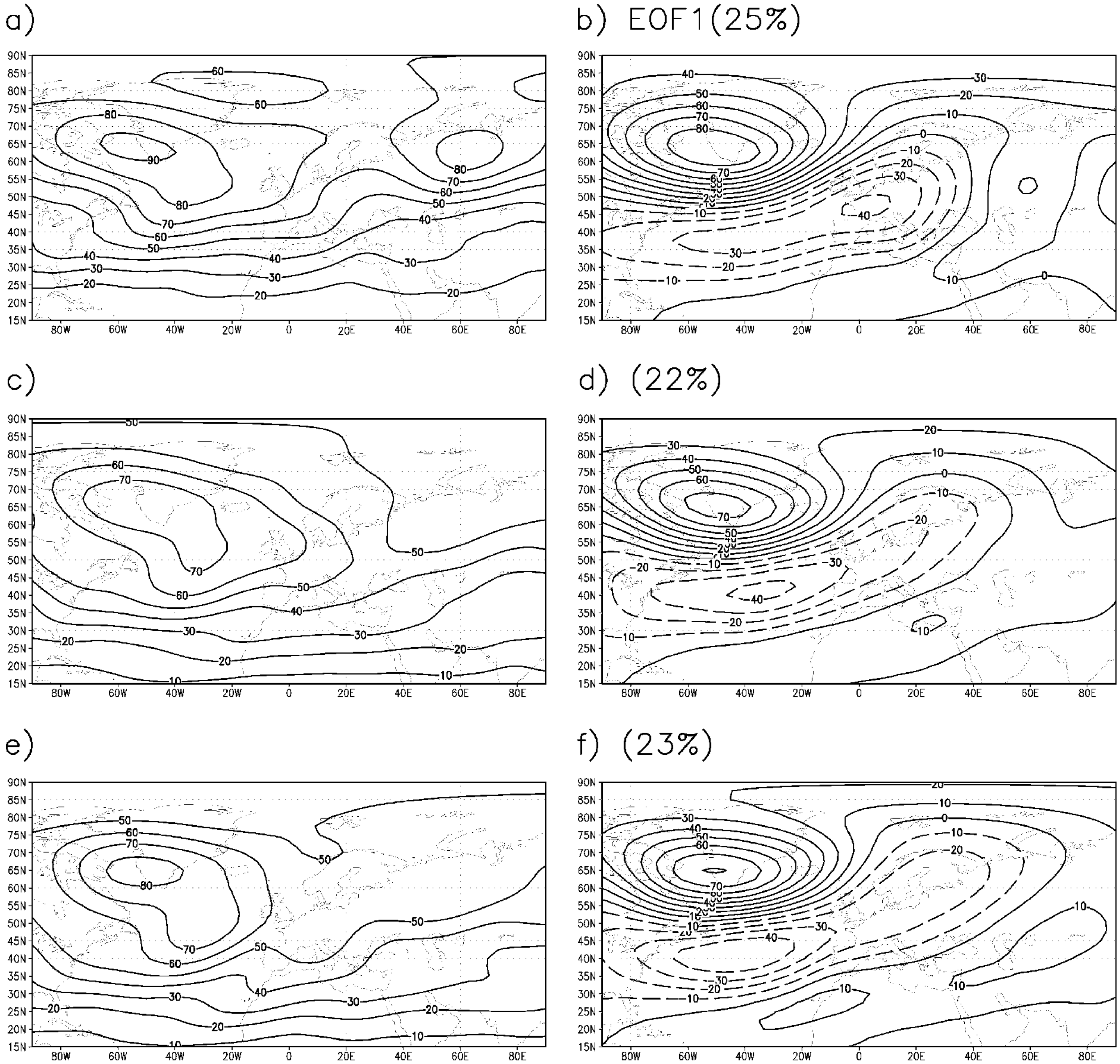

FIG. 2. The std dev and the leading EOF of monthly Z500 in late-winter (Feb–Apr) for (a),(b) observations; (c),(d) the AGCM; and (e),(f) the AGCM_ML. The contour interval is 10 m. The percentage of the variance explained by the EOF is given in the parentheses.

but is largely similar in early and late winter (not shown). Peng and Whitaker (1999) showed that the direct heating-forced response is far less sensitive to changes in the background state than is the eddy-forced response.

For a direct comparison with the response induced by the SST tripole, Fig. 4 shows the late-winter (February–April) Z500 response to the NAH anomaly, and the corresponding response to the tripole obtained by PRL02. Clearly, the two responses are drastically different, and the tripole-induced response is much stronger and projects highly on a negative NAO. Thus, contrary to what was suggested in CF02, our AGCM results indicate that the NAH SST is ineffective in forcing the NAO, and, consequently, it is unlikely to be modified by the NAO and evolve into the SST tripole. To understand the observed relationship between the fall SST and the winter NAO (or the SST tripole), alternative explanations are needed. We examine next the effects of the tropical SST on the NAO.

b. Tropical SST-induced response

1) AGCM_ML RESPONSE

To determine the coupled North Atlantic response to the tropical Atlantic SST anomaly south of 10°N (Fig.

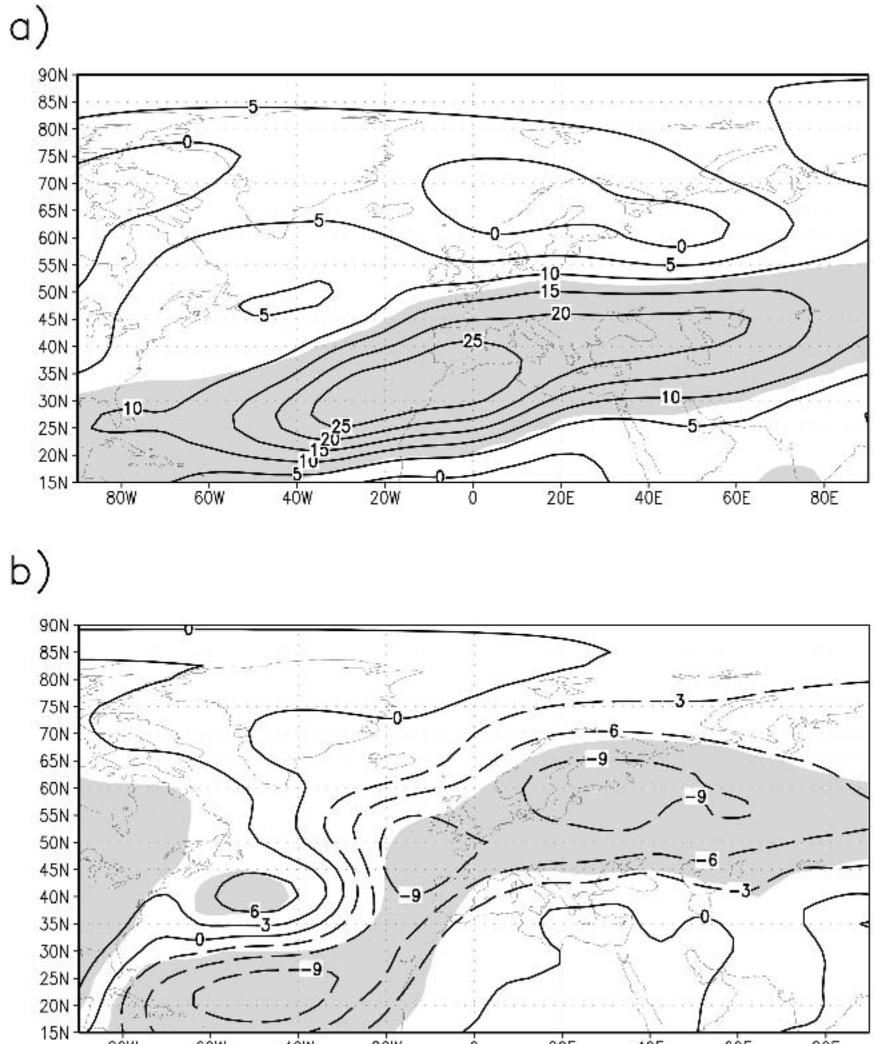

FIG. 3. AGCM geopotential height response at (a) 250 and (b) 1000 hPa for Oct–Apr as the ensemble mean difference between the runs with positive and negative NAH anomalies. The contour interval is 5 m in (a) and 3 m in (b). The shading in this and succeeding figures denotes areas with the response significant at 95% level as estimated by a *t* test.

1), and its seasonality and sign asymmetry, two 100-member ensembles of AGCM_ML experiments forced with positive and negative tropical anomalies are conducted as described in section 2. Since the tropical SST-induced geopotential height response is nearly equivalent-barotropic we show in Fig. 5 only the Z500 response to the positive and the negative anomaly for early (October–December) and late (February–April) winter as the ensemble mean difference between the SST-forced and the control runs. Note that for comparison the sign for the response induced by the negative anomaly is reversed in this and succeeding figures. The tropical SST-induced Z500 response exhibits a strong seasonality, in that the response is stronger and more NAO-like in late winter (Figs. 5b,d) than in early winter (Figs. 5a,c). In early winter, the positive anomaly induces a significant trough (~15 m) east of Newfoundland, with a wave train to the northeast, while the response to the negative anomaly is very weak. In late winter both the positive and the negative anomalies induce a significant NAO-like dipole (~30 m) over the North Atlantic. Qualitatively, this simulated Z500 response resembles the observed relationship between the fall positive (negative) tropical SST and the winter negative (positive) NAO depicted in CF02.

The corresponding extratropical SST response to the tropical SST anomaly, by way of the "atmospheric bridge," also exhibits a strong seasonality (Fig. 6). Part of the seasonality in the SST response is expected, because the coupled AGCM_ML integrations begin in September, and the SST response should grow with

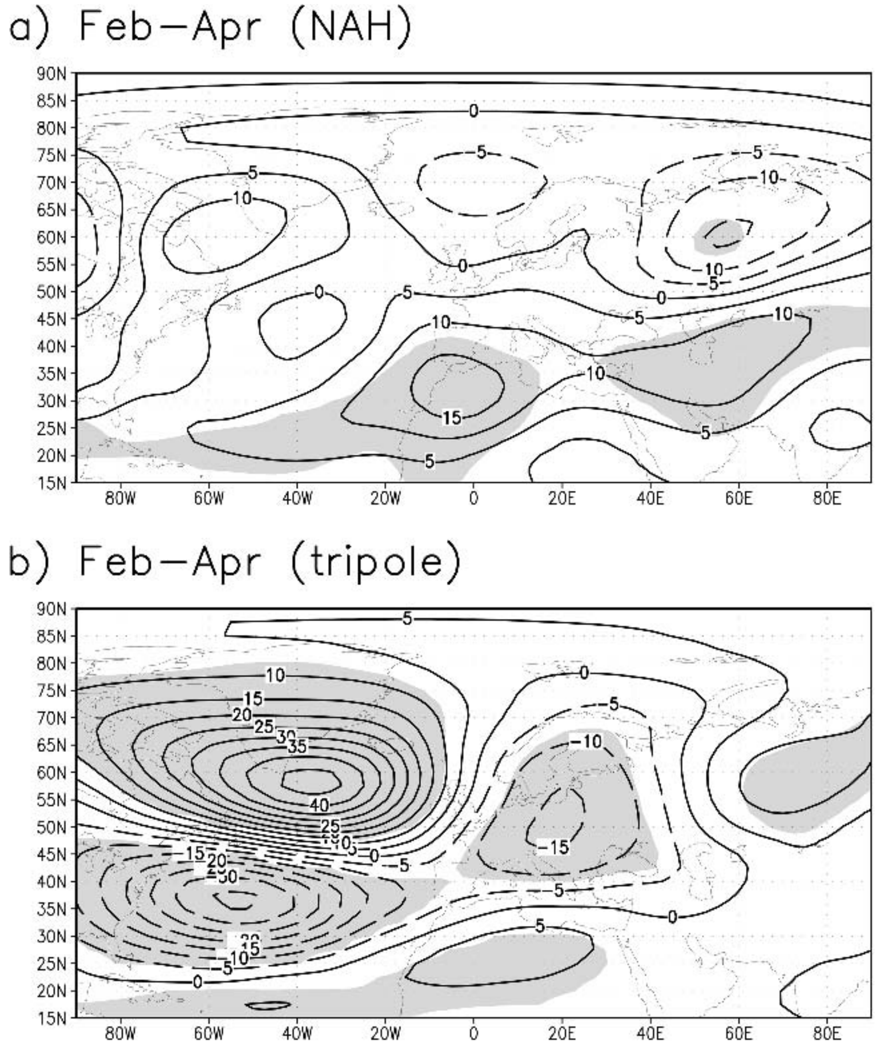

FIG. 4. As in Fig. 3 but for (a) the late-winter (Feb–Apr) Z500 response to the NAH anomaly (positive–negative), and (b) the corresponding Z500 response to the SST tripole replotted from PRL02. The contour interval is 5 m.

time even under the constant atmospheric forcing. More importantly, however, the seasonality in the SST response, especially in its spatial structure, is associated with the seasonality in the atmospheric response. Corresponding to the NAO-like response in late winter, the SST response features a tripole pattern with a maximum strength of about 0.5 K (Figs. 6b,d). This coupled NAO-SST tripole response is largely similar to the observed winter NAO and its associated SST tripole (Marshall et al. 2001). In early winter, the SST response is much weaker, but the response induced by positive tropical SST anomaly is significant and well defined, with a maximum of about 0.2 K (Fig. 6a). Most noteworthy, this early-winter extratropical SST response to the positive tropical anomaly resembles the fall NAH anomaly, with a negative center off Newfoundland surrounded by positive anomalies to the northeast and the southeast. Some differences between the simulated early-winter SST response (Fig. 6a) and the JAS NAH SST (Fig. 1) are not surprising, considering that the AGCM-ML is a simplified coupled system and that the observed NAH pattern also varies from JAS to OND, as indicated in CF02 (their Fig. 2). It should be mentioned that in both early and late winter, the tropical anomaly induces little significant SST response in the Pacific.

To demonstrate further that the NAH and the tripole SST anomalies induced by the positive tropical anomaly are directly related to the seasonality in the atmospheric response, we show, in Fig. 7, the responses in the total downward surface heat flux (latent + sensible + radiative) and in 925-hPa temperature (T925) for September–November and for January–March (i.e., 1 month leading the SST response). Even more clearly

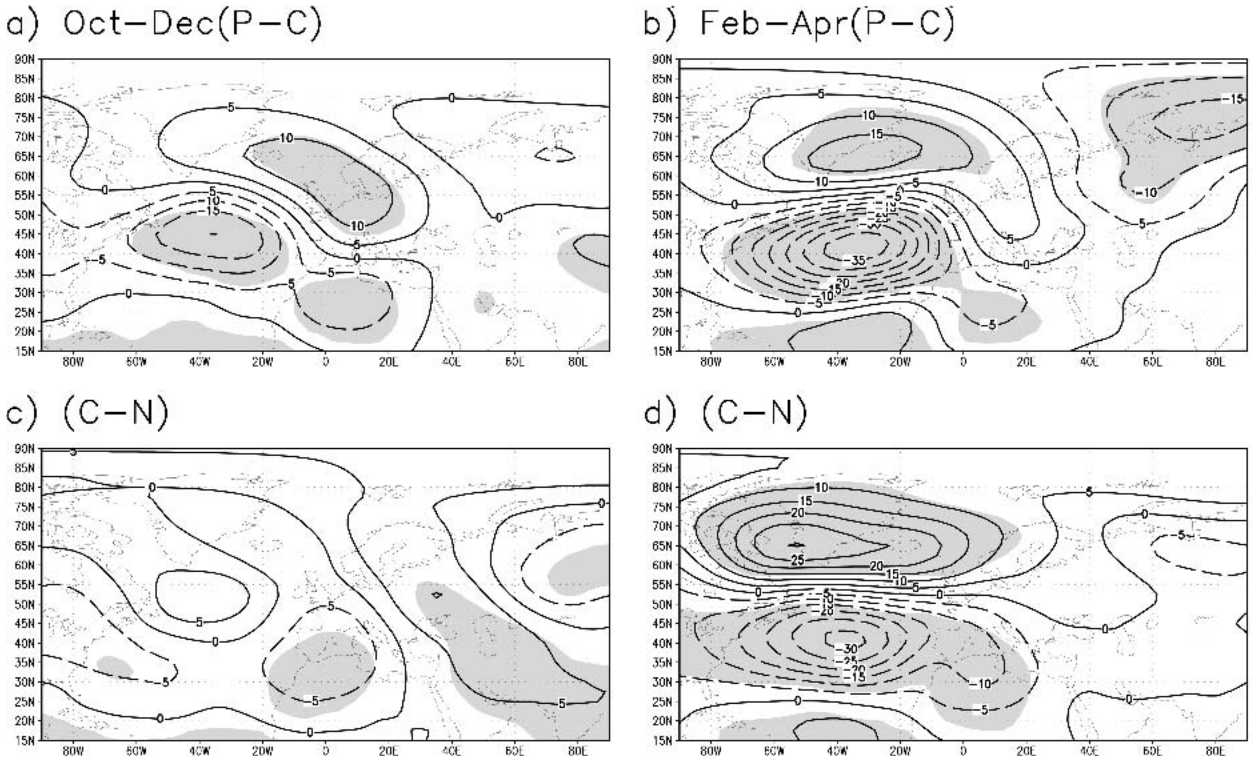

FIG. 5. AGCM_ML Z500 response to positive and negative tropical anomalies for (a),(c) early winter (Oct–Dec) and for (b),(d) late winter (Feb–Apr) as the ensemble mean difference between the SST-forced and the control runs. For comparison, the sign for the negative anomaly-induced response in (c), (d) and in succeeding figures is reversed. The contour interval is 5 m.

than the SST response, the preceding heat flux and the T925 responses both exhibit a seasonality of the NAH-like pattern leading the tripole. The strong similarity between the SST and the low-level temperature responses suggests that the NAH and the tripole SST anomalies result mainly from the seasonality in the atmospheric temperature response.

The above AGCM_ML results reveal that the tropical Atlantic anomaly can significantly influence the coupled North Atlantic variability, but with a strong seasonal dependence. An identical positive tropical anomaly induces an anomalous trough east of Newfoundland accompanied by a NAH-like SST in early winter, and a negative NAO and SST tripole in late winter. This suggests that the observed relationship between the fall NAH SST and the winter NAO (or SST tripole) may arise from persistent tropical SST forcing of the seasonally varying atmosphere.

2) COMPARISON WITH THE AGCM RESPONSE

To determine the influence of extratropical oceanic feedbacks on the coupled AGCM_ML response to the tropical anomaly, parallel 100-member ensembles of AGCM experiments are performed. Since the extratropical SST response in the AGCM_ML, and therefore, its potential back-influence on the atmosphere, is much stronger in late than in early winter, we compare the responses in the two models for this period. Absent oceanic feedbacks in the AGCM, the late-winter Z500 response to the tropical anomaly (Fig. 8) exhibits a pronounced asymmetry about the sign of the SST anomaly. The positive SST induces a negative NAO-like dipole (~25 m), whereas the negative SST induces only the trough (~20 m) southeast of Newfoundland but not the ridge to the north. Comparing the Z500 responses in the two models (Figs. 5 and 8) reveals that the influence of oceanic feedbacks on the response is more evident in the negative case than in the positive case. The response to a positive anomaly in both models features a similar NAO-like dipole, except that the southern trough is stronger in the AGCM_ML. The extratropical thermal coupling in the negative case results not only in an enhanced southern trough but also a comparable northern ridge that is nearly absent in the AGCM. Overall, with the thermal feedbacks from the tripole-like SST anomaly present, the equilibrium Z500 response in the AGCM_ML is stronger and more NAO-like. This is consistent with our earlier AGCM results that showed a significant influence of the SST tripole on the NAO (PRL02; PRL03). The strength of the SST feedback (~20 m K^{-1}), as deduced from the difference in the southern trough between the two models, is also largely consistent

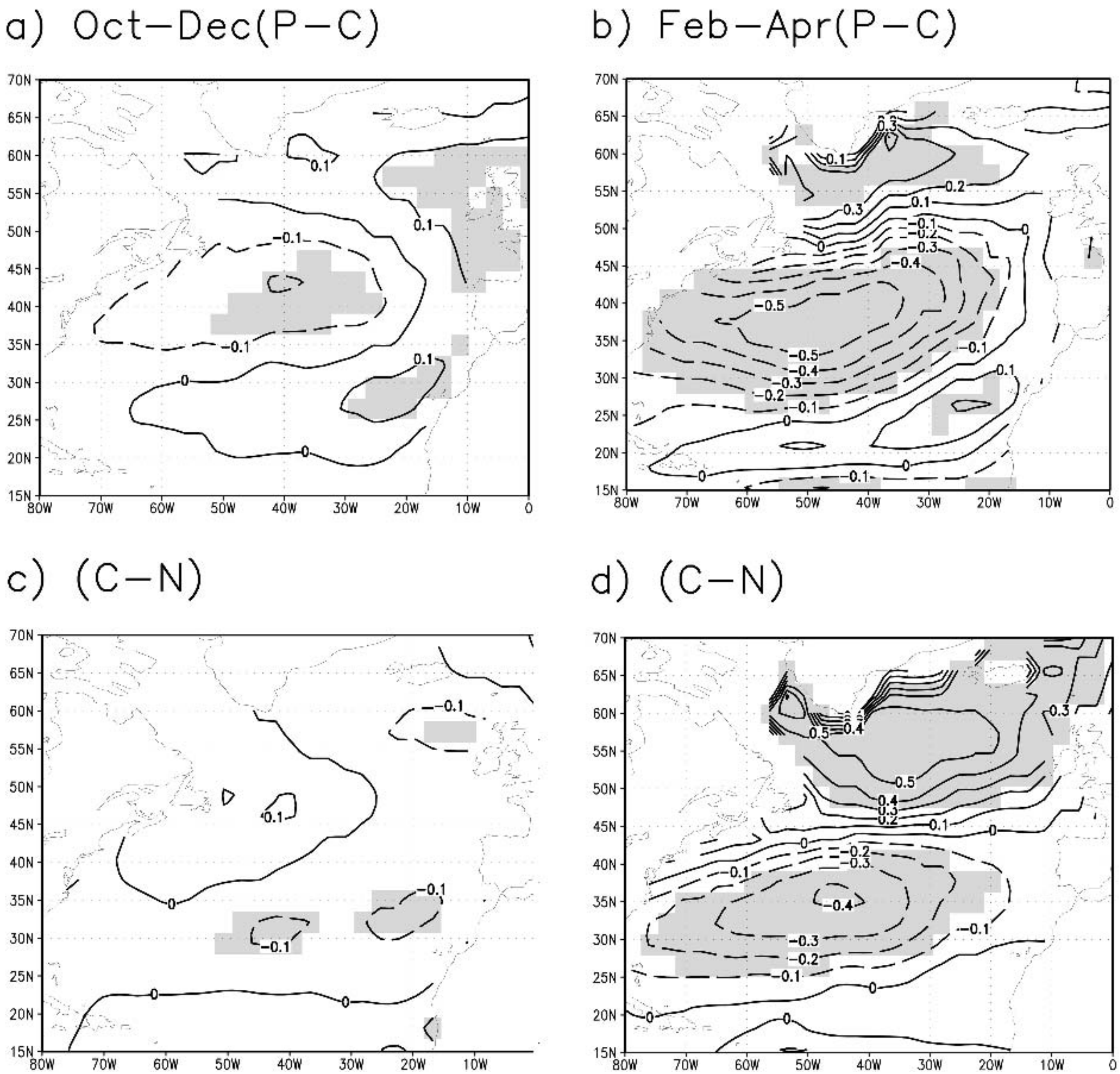

FIG. 6. As in Fig. 5 but for the AGCM_ML SST response. The contour interval is 0.1 K.

with that of the modeled response to the tripole (see Fig. 4b).

A basic effect of oceanic thermal coupling is to reduce the anomalous temperature difference between the atmosphere and the ocean and, consequently, the anomalous surface heat fluxes and the thermal damping of the atmospheric anomalies. This is termed the "reduced thermal damping effect" (Barsugli 1995; Bladé 1997; Barsugli and Battisti 1998). Since no radiative fluxes were saved for the earlier 100 AGCM control runs, we compare only the responses in the surface sensible and latent heat fluxes between the two models. Figure 9 shows that the response of the extratropical heat flux to the tropical SST anomaly in the AGCM_ML is indeed reduced by about 50% in comparison with that in the AGCM. Assuming that the response is about 0.5 K in the 1000–300-hPa depth-averaged temperature in both models and that the response in the surface flux is 7.5 W m^{-2} in the AGCM_ML and 15 W m^{-2} in the AGCM, the damping time scale is estimated to be about 5.5 days in the AGCM_ML and half of that in the AGCM. Since the actual temperature response is slightly stronger in the AGCM_ML, this suggests that coupling to a mixed layer reduces the thermal damping by about a factor of 3. We note, however, that, apart from the amplitude difference, the heat-flux responses in the two models also exhibit substantial differences in their patterns, as-

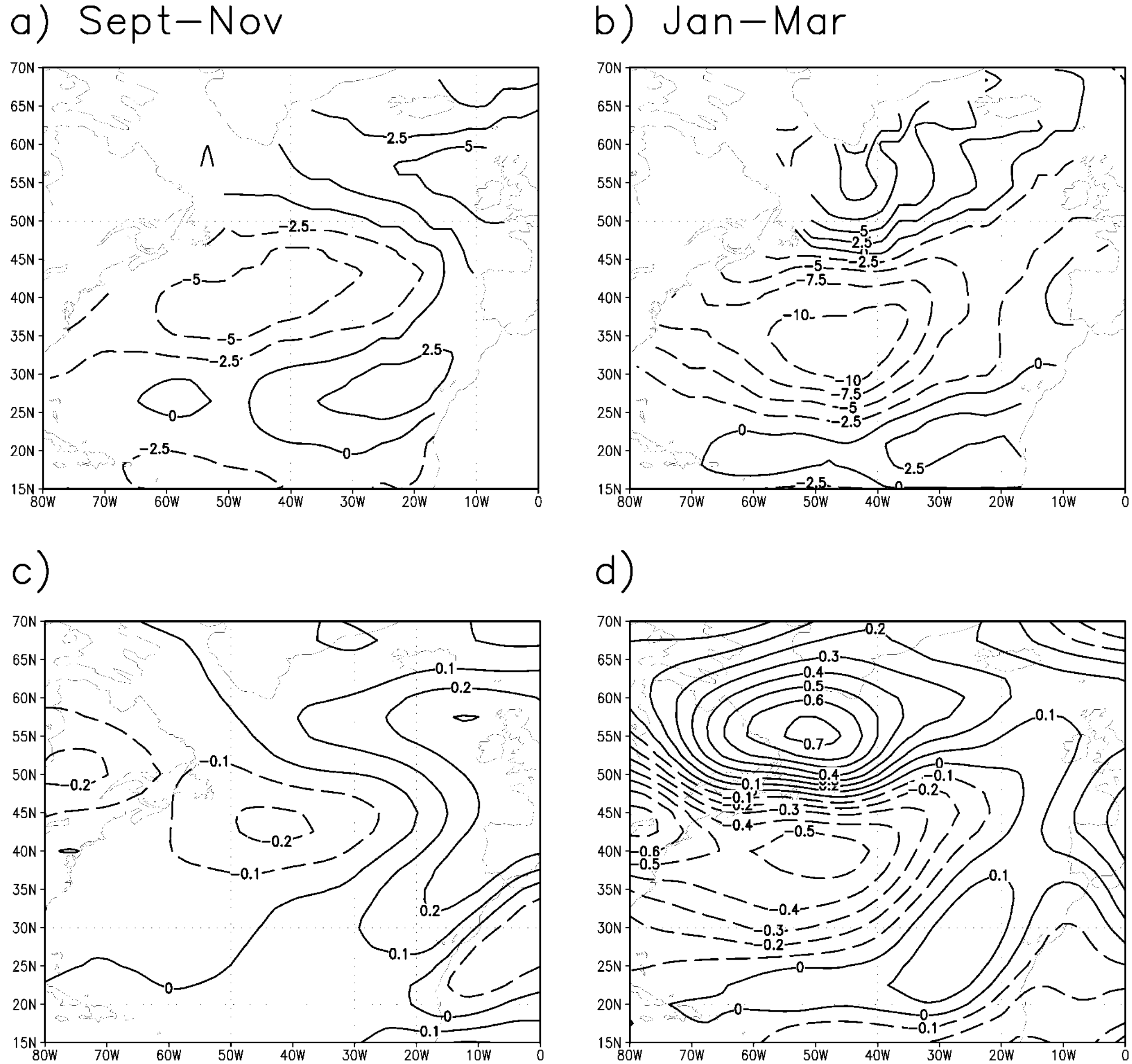

FIG. 7. AGCM_ML response to positive tropical anomaly in the total downward surface heat flux and in 925-hPa temperature for (a),(c) Sep–Nov and for (b),(d) Jan–Mar. The contour interval is 2.5 W m^{-2} in (a) and (b), and is 0.1 K in (c) and (d).

sociated with the different responses of the atmospheric circulation. Such differences are not easily explained by the reduced thermal damping effect, but are consistent with an active forcing of the atmosphere by the extratropical SST (Watanabe and Kimoto 2000; PRL03). Watanabe and Kimoto suggest that a positive SST southeast of Newfoundland (as can be inferred from Fig. 9d with a sign reversal) may be effective in forcing a positive NAO. This may, in part, be responsible for the different height (Figs. 5d and 8b) and flux (Figs. 9c,d) responses to the negative anomaly in the two models.

3) MECHANISMS

While an in-depth mechanistic diagnosis of the tropically forced response is beyond the scope of this paper, we nevertheless briefly explore possible key processes that determine the coupled response and its seasonality. Previous studies suggest that a SST-induced extratropical response in GCMs is primarily sustained by two forcings—namely, the diabatic heating and the vorticity forcing from synoptic transients (Ting and Peng 1995; Peng and Whitaker 1999; PRL03). Initially, the heating comes directly from the SST-induced

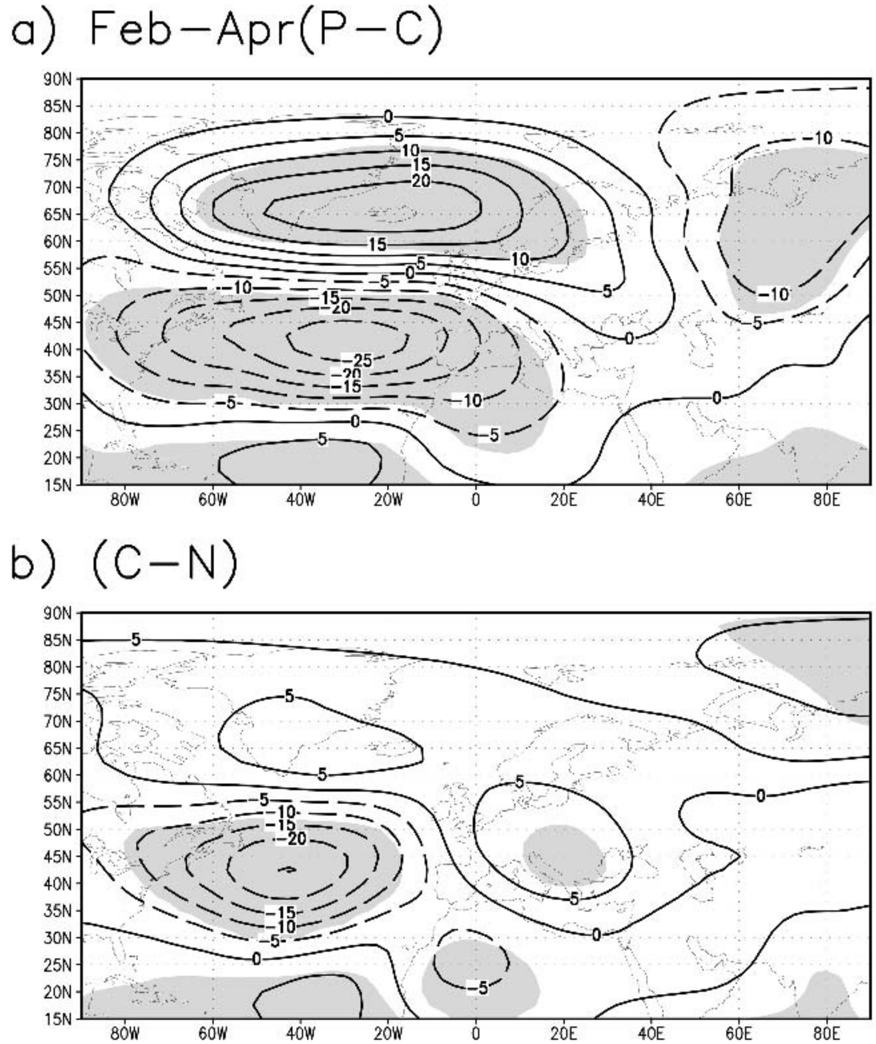

FIG. 8. As in Fig. 5 but for the AGCM late-winter Z500 response to (a) positive and (b) negative anomalies.

anomalous surface heat fluxes, and the eddy forcing results from the interaction of the heating-forced anomalous flow with the extratropical storm tracks. The two forcings (or the forced anomalous flows) subsequently interact with and modify each other and eventually determine the equilibrium response. In the tropical SST-forced case, since the heat source and the eddy forcing are geographically separated, eddy feedbacks cannot directly modify the heating, and alter the nature of the heating-forced anomalous flow. Hence, eddy forcing plays a somewhat secondary role in modulating and sustaining the tropically forced response, and the response exhibits less qualitative sensitivity to the choice of model, to the seasons, and to the background flow (Hall and Derome 2000; Hoerling and Kumar 2002). In contrast, in the extratropical SST-forced case, the heat source and the eddy forcing reside in the same region, and the heating can be strongly modified by eddy feedbacks, so much so in some cases to even reverse the sign of surface heat-flux anomalies (Latif and Barnett 1994; Peng et al. 1995). Hence, eddy feedbacks play a more determining role in defining the extratropical SST-forced response, and the nature of the response depends sensitively on background intrinsic variability associated with different seasons and models.

Since the coupled Z500 response shown in Fig. 5 is forced by the tropical SST anomaly, despite the influence of the extratropical SST, we expect that the SST-induced tropical diabatic heating plays a significant role in determining the nature of the extratropical response. As a proxy for the depth-averaged diabatic heating, the precipitation response shown in Fig. 10 indicates that the anomalous heating is largely similar for positive and negative cases, but differs more between early and late winter. The early-winter heating structure is simpler

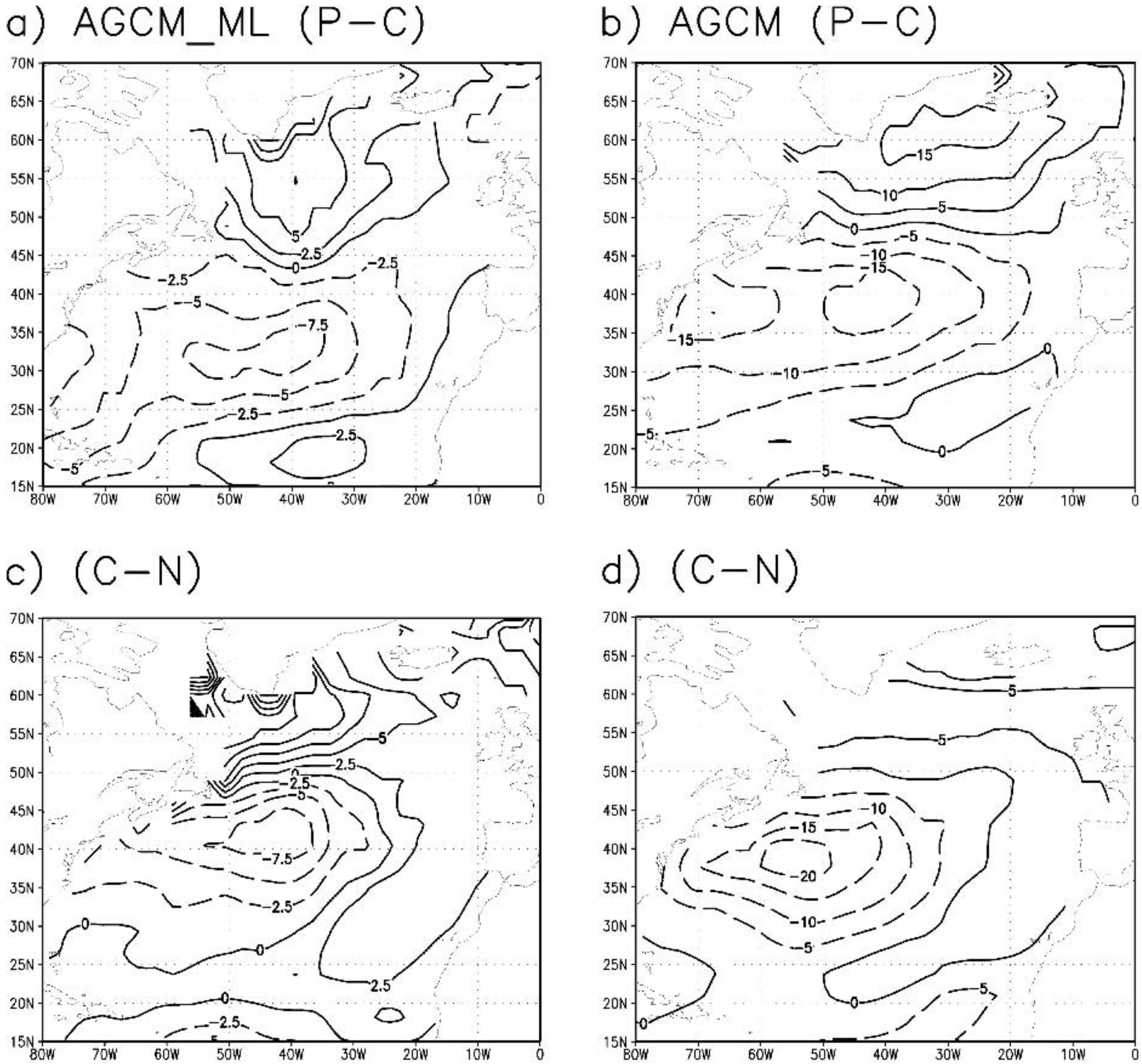

FIG. 9. The late-winter response in the downward surface sensible and latent heat fluxes to positive and negative tropical anomalies in (a),(c) the AGCM_ML and (b),(d) the AGCM. The contour interval is 2.5 W m^{-2} in (a) and (c), and 5 W m^{-2} in (b) and (d).

with a roughly elliptical center near 25°W on the equator, while the late-winter heating pattern is more complex with the heating maximum shifted to 5°S , 15°W. To determine the anomalous flow forced directly by the tropical heating, in the absence of extratropical eddy feedbacks and thermal coupling, linear baroclinic model (LBM) experiments are conducted with an idealized elliptical heating centered at different locations for the early- and late-winter AGCM_ML basic states. It should be mentioned that the tropical precipitation response in the AGCM is nearly identical to that in the AGCM_ML. Since the basic states from the two models are also similar, these LBM results are equally applicable to the AGCM responses.

Figure 11 shows the depth-averaged amplitude of the idealized heating pattern together with its vertical profile that mimic the heating distribution in the AGCM_ML. The heating amplitude (5 K day^{-1}) is enhanced from the AGCM_ML data in order to compensate for the LBM's deficiency in responding too weakly to forcings, in comparison with GCMs, due to the limitations of the linear dynamics (PRL03). This affects the amplitude, but not the structure, of the response. Under the early-winter basic state, with the heating centered at 25°W on the equator, the LBM 550 hPa height response features a large-scale wave train over the North Atlantic with a trough southeast of Newfoundland and a ridge to the northeast (Fig. 12a). This linear

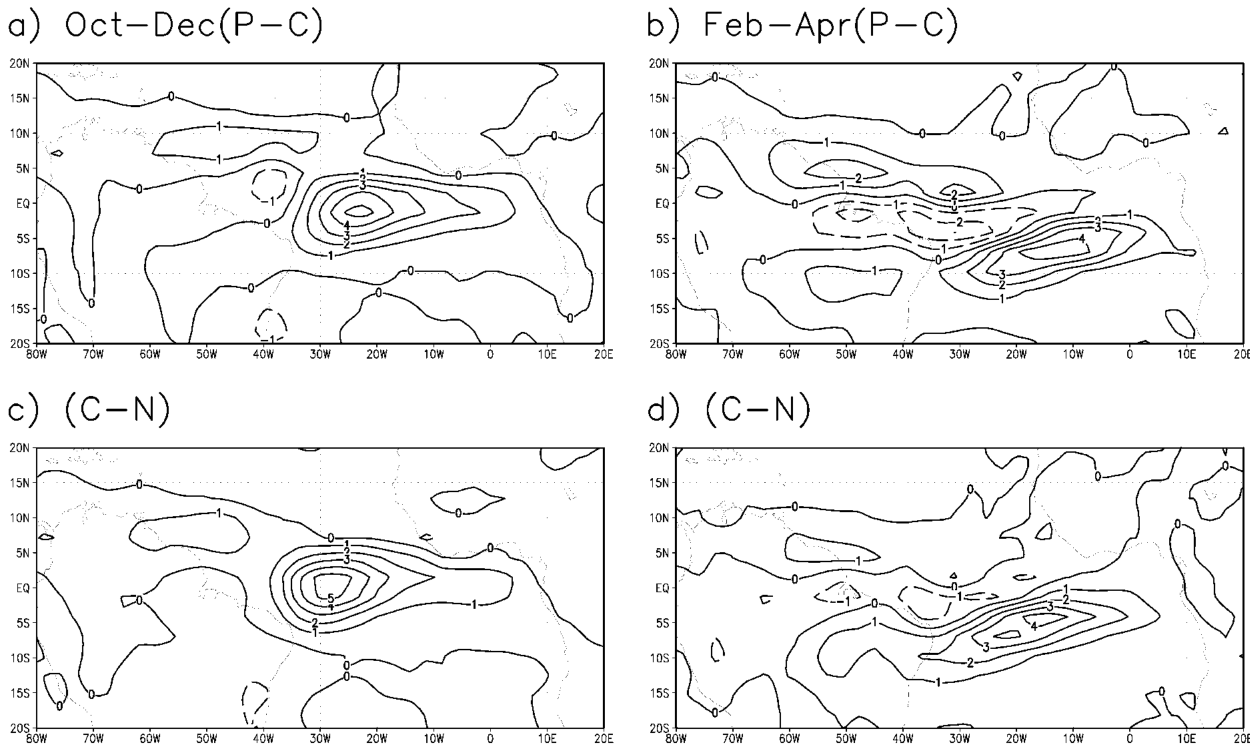

FIG. 10. As in Fig. 5 but for the AGCM_ML precipitation response. The contour interval is 1 mm day^{-1}.

response to tropical heating bears a clear resemblance to the early-winter Z500 response to the positive SST in the AGCM_ML (Fig. 5a), but it differs strongly from a typical NAO pattern characterized by a north–south dipole with little downstream propagation. With the late-winter basic state and the heating shifted to 5°S and 15°W, the LBM produces a stronger but qualitatively similar wave train–like response (Fig. 12b). The enhanced amplitude is mainly due to the difference in the basic state not in the location of the heating. With the heating locations reversed for early and late winter, the LBM responses are nearly indistinguishable from those shown in Fig. 12. Thus, regardless of the details in the heating, the tropical heating induces a wave train–like response that is similar in nature in early and late winter. By itself, the heating cannot directly force a NAO-like dipole in either season.

The LBM results suggest that the late-winter NAO-like responses to the tropical SST produced by the AGCM_ML likely result from strong eddy feedbacks on the heating-forced anomalous flow. Since the synoptic eddies and the NAO are more active later in the winter, the heating-forced anomalous flow can perturb the storm tracks more vigorously. Such perturbations of the storm tracks lead to anomalous transient eddy fluxes of vorticity, which in turn generate the NAO. Indeed, we find that the eddy forcing associated with the tropical SST-induced response in the AGCM_ML is much stronger in late winter than in early winter (not shown). While the late-winter NAO response may be strongly modulated by eddy feedbacks, the influence of the tropical heating on the equilibrium extratropical coupled response is still readily visible. Note that the linear heating-forced anomalous flow (Fig. 12) projects more on a negative than on a positive NAO. Hence, the polarity of the coupled Z500 response in both early and late winter is determined by the tropical heating.

4. Summary and discussion

To understand the observed relationship between the fall Pan-Atlantic SST anomaly and the winter NAO identified by CF02, the influence of both the NAH and the tropical parts of the anomaly on the NAO over the boreal winter months is examined in a series of systematic model experiments. First, ensembles of AGCM experiments are conducted to determine if the NAH anomaly persisting from the fall may be more effective in forcing the winter NAO than its associated SST tripole. The model results reveal that the NAH anomaly induces a baroclinic response throughout the winter with little projection on the NAO. Since the fall NAH anomaly is ineffective in forcing the NAO, it cannot account for the observation that the fall NAH SST leads the winter NAO. Ensembles of AGCM_ML experiments are then performed to determine the coupled North Atlantic response to the tropical anomaly. The tropically forced Z500 response exhibits a strong sea-

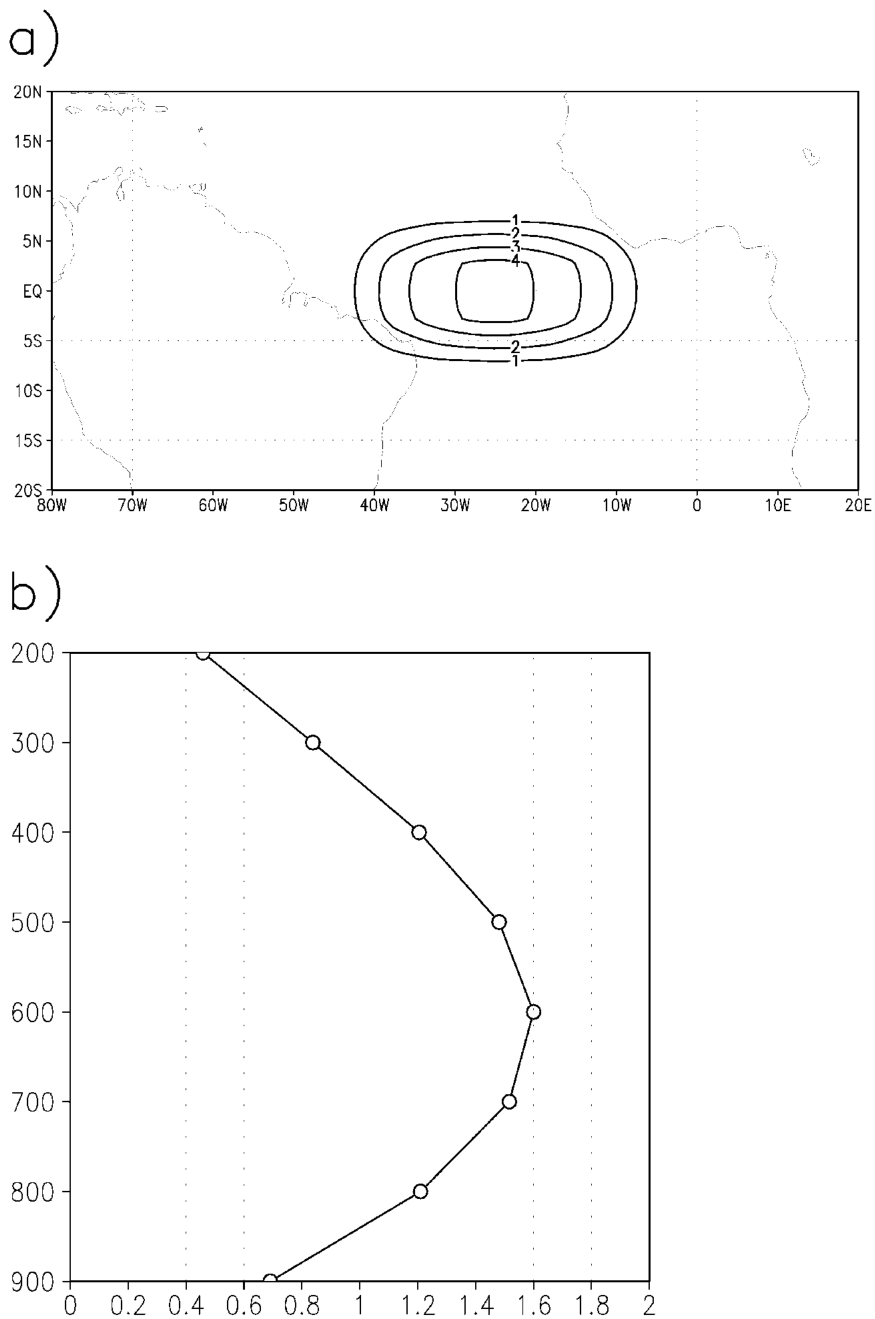

Fig. 11. (a) Idealized heating pattern with depth-averaged heating rates, and (b) its vertical heating profile. The contour interval in (a) is 1 K day^{-1}.

sonality. In early winter the positive anomaly induces a trough east of Newfoundland with a wave train to the northeast, and in late winter the response projects strongly on a negative NAO. The corresponding response in the extratropical SST features a NAH-like pattern in early winter and a tripole in later winter. A similar NAO-like response is induced by the negative tropical SST anomaly in late winter, but the response to the negative anomaly is weak in early winter. Overall, the AGCM_ML results suggest that tropical Atlantic SST anomalies can significantly influence the coupled North Atlantic ocean–atmosphere system. The observed relationship between the fall NAH SST and the winter NAO (or SST tripole) may result from the seasonal march of responses to persistent tropical forcing. We thus view the fall NAH SST state to be mostly coincidental, rather than causal, to the subsequent winter NAO state.

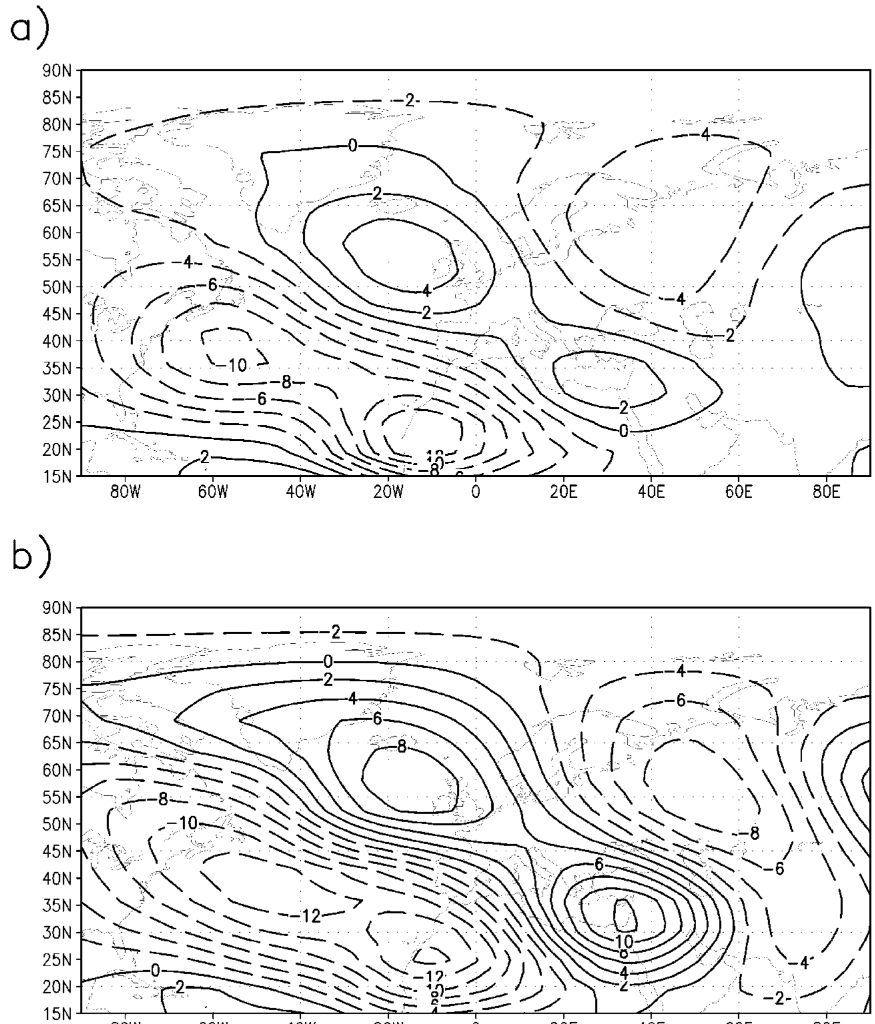

FIG. 12. LBM 550-hPa height response to the idealized heating at (a) 0°, 25°W for early-winter basic state and (b) 5°S, 15°W for late-winter basic state. The contour interval is 2 m.

The AGCM_ML results are further compared with the results of parallel AGCM experiments, in order to determine the influence of extratropical thermal coupling on the tropically forced NAO response in late winter. In the absence of extratropical oceanic feedbacks, a negative NAO dipole is still induced by the positive anomaly, but the response to the negative anomaly has only a monopole southern center. Hence, the largely symmetric NAO responses, with respect to the sign of the tropical anomaly, arise in the AGCM_ML in part as a result of active extratropical SST feedbacks.

Mechanisms that determine the seasonality of the tropically forced coupled response are briefly explored using linear model experiments. Without eddy feedbacks, an idealized tropical heating induces a qualitatively similar large-scale wave train over the North Atlantic with a trough southeast of Newfoundland and a ridge to the northeast in early and in late winter. The enhanced seasonality in the SST-induced response likely arises, at least in part, from the seasonal modulation of transient-eddy feedbacks on the heating-forced anomalous flow. Further in-depth diagnostic studies will be conducted to quantify the maintenance of the SST-induced response by various forcings, and to elucidate better the associated dynamic processes. We are particularly interested in understanding the causes for the strong sign asymmetry in the tropically forced AGCM response, the causes for the sign asymmetry in the extratropical oceanic influence on the response, and their possible connections.

The coupled AGCM_ML experiments in this study are simplified in several respects in order to isolate the problem. The model is coupled only with the extratropical oceans, and is forced by a tropical anomaly that is held fixed throughout the boreal winter. This experimental design isolates the tropically forced extratropical coupled response and its dependence on the atmospheric seasonality, but it does not allow the tropical anomaly to be modified by the response. In reality, the tropically forced response may exert a back influence on the tropical ocean and cause the tropical SST anomaly to change with time. The results of CF02 (their Figs. 2 and 10) appear to indicate that a similar tropical SST anomaly, accompanied by the NAH in the extratropics, persists from summer into early winter, but disappears once the NAO-SST tripole becomes dominant. This apparent lack of persistence in the tropical anomaly beyond midwinter may be determined by local air–sea interactions. It is also possible that the tropically forced NAO-tripole response may tend to destroy the tropical SST anomaly, and thus prevent it from surviving through the winter months. The potential back-influence of the NAO response on the tropical anomaly will be explored in future studies.

The strength of the Z500 response induced by the tropical anomaly in the AGCM_ML is about 30 m K^{-1} in the southern trough in February–April (Fig. 5). In the MCA modes of CF02, the observed NDJ Z500 is related to the preceding tropical SST by about 50 m K^{-1} in the southern trough with an even stronger signal in the northern ridge (their Figs. 2 and 10). Since the MCA modes cannot be interpreted as a response to a particular forcing exclusively, one should avoid making direct quantitative comparisons between the MCA modes and the model response, without taking into consideration limitations with the MCA analyses and with the model. We note, for example, that, unlike the MCA modes, the tropically forced extratropical SST response in the AGCM_ML is only about half as strong as the tropical anomaly. This discrepancy may be due in part to the omission of certain important processes in the slab-ocean model, such as the absence of Ekman transport. Based on the surface wind response induced by the tropical anomaly, we estimate that Ekman transport would reinforce the extratropical SST response and potentially lead to a stronger NAO response. We plan in future studies to expand the mixed layer ocean to include the Ekman transport, and perhaps also a seasonally varying mixed layer depth, in order to determine their effects on the extratropical coupled response and on the seasonality.

Apart from the simplicity of the slab ocean, it is known that the atmospheric seasonality simulated in models often deviates somewhat from the seasonality of the real atmosphere and consequently affects the seasonality in the simulated SST-forced response. The transition from a trough to a NAO response to the positive tropical anomaly in our model (Fig. 5) appears to occur in midwinter (around January), whereas that in observations is probably earlier—around December as indicated in CF02. Nevertheless, the broad qualitative agreement of our AGCM_ML results with the observations indicates the likely influence of persistent tropical SST forcing on the extratropical coupled variability and its seasonality in nature.

Last, it should be emphasized that, apart from tropical Atlantic SST, the observed relationship between the fall NAH SST and the winter NAO (or SST tripole) may also result from other persistent forcings, such as tropical or extratropical Pacific SST anomalies. In fact, CF02 showed that, excluding the tropical Atlantic SST in the analyses, the leading MCA mode still indicates a significant relationship between the JAS NAH and the NDJ NAO. Moreover, the global SST correlation with this mode exhibits a stronger connection with the extratropical Pacific SST than with the tropical SST (Fig. 6; CF02). Further observational and modeling studies are required to elucidate such potential connections between Pacific and Atlantic variability.

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