NOTES AND CORRESPONDENCE

Annular Modes Forced from the Stratosphere and Interactions with the Oceans

Seiji YUKIMOTO

Climate Research Department, Meteorological Research Institute, Tsukuba, Japan

and

Kunihiko KODERA

*Graduate School of Environmental Studies, Nagoya University, Nagoya, Japan
Climate Research Department, Meteorological Research Institute, Tsukuba, Japan*

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Abstract

Long time-scale variations of the climate system can be induced by forcings loaded onto the stratosphere, such as increase of carbon dioxide, solar activity, and volcanic eruption, which is possibly modulated by the interaction with the oceans. Spatial structure of the atmospheric response resembles the Northern Annular Mode (NAM) in an experiment of a coupled atmosphere-ocean general circulation model with a westerly momentum forcing in the winter midlatitude stratosphere. The atmospheric response further induces distinctive patterns of the sea surface temperature (SST) change in the Atlantic and the Pacific, which are reminiscent of the observed decadal to interdecadal variability dominant in each basin. Investigation of oceanic feedback with the atmospheric model forced by the SST change suggests that the oceanic response has positive feedback effects on the NAM-like atmospheric structure, implying enforcement or maintenance of the response to the stratospheric forcing. The positive feedback effect in the Atlantic is significant in winter, while that in the Pacific is significant in spring.

1. Introduction

The Northern Annular Mode (NAM) or Arctic Oscillation (AO) is one of the most dominant variations in the atmospheric general circulation in the extratropics, and has decadal to interdecadal time-scales that cannot be explained by the inherent atmospheric dynamics alone. One of the conceivable causes of such long-term variation is modulation by interactions

with the ocean that has large thermal and dynamical inertia. Numerous studies (Czaja et al. 2003, and references there in) investigated association of the North Atlantic Oscillation (NAO), considered as an essential part of the NAM, with the oceanic variation in the North Atlantic. It is still an open question whether the atmosphere-ocean coupled variability can fully explain the observed decadal variation.

Another possible cause that we are focusing on is external forcing with long-term fluctuations, for instance, anthropogenic increase of carbon dioxide, solar activity, and volcanic eruptions. The observed decadal variability may not be only a response of the atmosphere to the ex-

Corresponding author: Seiji Yukimoto, Climate Research Department, Meteorological Research Institute, Tsukuba, 305-0052, Japan.
E-mail: yukimoto@mri-jma.go.jp
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ternal forcing, but may be contributed from some modulation by interaction with the ocean that has long timescales.

There is a controversy about origin of the NAM variability, namely tropospheric origin (e.g., Limpasuvan and Hartmann 1999; Kimoto et al. 2001) or stratospheric origin (e.g., Baldwin and Dunkerton 1999). The tropospheric origin includes forcing due to sea surface temperature change (Rodwell et al. 1999).

It is suggested that changes in the stratosphere due to external forcings lead to the tropospheric NAM-like response (Shindell et al. 2001). The tropospheric response would further influence ocean, though feedback from the ocean response is not fully investigated.

We examine how the coupled atmosphere-ocean system responds to an external forcing loaded in the stratosphere with a coupled ocean general circulation model (CGCM) and investigate the interaction by additional uncoupled experiments. Although there are various stratospheric forcing agents (changes of greenhouse gases, ozone heating by solar activity, volcanic aerosols, etc.) that may cause NAM-like response, we do not focus on identifying the forcing mechanism in the stratosphere, but aim at finding a pathway of the NAM-like change from stratosphere to the oceans with ocean feedbacks. The expected key processes in the pathway are coherent changes of planetary wave propagation and meridional circulation.

2. Experimental design

The CGCM used for this study is the MRI-CGCM2.3.2 (Yukimoto et al. 2006). The atmospheric component has T42 horizontal resolution and 30 vertical levels with 0.4 hPa level at the top. The oceanic part is a global ocean general circulation model with 2.5 (longitude) by 2 (latitude) degrees horizontal resolution (~ 0.5 degrees in latitude near the equator), and 23 vertical layers, and a thermodynamical sea ice model with accounting for free drift of sea ice.

The model was evaluated for performance in reproducing mean climate of the fundamental fields, such as solar and terrestrial radiation budgets, surface temperature, precipitation, and atmospheric circulation (Yukimoto et al. 2006). The model also exhibits NAM-like variability in month-to-month and interdecadal timescales comparable to the observed NAM

(Yukimoto and Kodera 2005). In the experiments for the 21st Century climate change scenario, the simulated NAM revealed positive trend in a warmer climate, the same as in recent experiments by many other climate models (Miller et al. 2006).

A control simulation (referred to CC) and a forced experiment (referred to CM) are performed with the CGCM. The forced response ΔCM is defined as $\Delta CM = CM - CC$. In the CM experiment, we imposed a momentum forcing in the stratosphere rather than a thermal forcing for the following reason. In the real climate system, since mechanical effects and thermal effects closely interact each other, it is difficult to impose exclusively either thermal or mechanical forcing component so that the response resembles the observed variation in nature. For instance, when we impose ozone heating change due to solar variation, the ozone distribution will be changed by the circulation change due to the original heating change. Therefore, we do not stick to whether the original forcing is thermal or mechanical. We intend to impose a zonal momentum forcing that concurrently causes coherent changes of planetary wave propagation and meridional circulation.

The momentum forcing F_m loaded in the experiment CM, is expressed as

$$F_m = A_s f(p) (\sin 2\phi)^2 \cos(2\pi n_d/365),$$

where p and ϕ denote pressure (hPa) and latitude (radians), and n_d is number of days from 15 January. The vertical profile $f(p)$ is expressed as

$$f(p) = \begin{cases} 1.0 & p < 10 \\ (\ln p - \ln 100)/\ln 0.1 & 10 < p < 100 \\ 0.0 & 100 < p \end{cases}$$

which gives 1.0 above 10 hPa and zero below the 100 hPa with log-linear decrease for pressure between the 10 hPa and the 100 hPa. The spatial structure is the same as that used by Thuburn and Craig (2000), but the amplitude of the forcing varies with the season and applied only in winter hemisphere. This idealized forcing gives maxima $A_s = 5.0 \text{ ms}^{-1}\text{day}^{-1}$ at the 45°N latitude of the stratosphere above the 10 hPa in winter (maximum in mid-January in the northern hemisphere).

The magnitude of this forcing at the maximum phase is as large as in Thuburn and Craig

Fig. 1. Responses in the stratospheric westerly momentum forcing experiment (ΔCM) for (a) zonal-mean zonal wind (unit: ms^{-1}), (b) EP-flux (vectors) and EP-flux divergence (shading, unit: $1 \times 10^{-4} \text{ ms}^{-2}$), and (c) Eulerian mean meridional circulation (unit: 10^9 kgs^{-1}), averaged for January through March. Shadings in (a) and (c) denote regions with 99% significance.

(2000). The forcing may be too strong when considering a long-term variation (e.g., solar cycle: Kodera 2002), however, it is comparable to the order of breaking by planetary waves in a short-term (e.g., day-to-day) variation of the real winter stratosphere, such as in the stratospheric sudden warming case. We considered a ‘possible extreme’ so that the response becomes sufficiently distinct compared to the internal variability.

In addition to the experiments with the CGCM, the atmospheric model (AGCM; a component of the CGCM) is forced by the SSTs reproduced in the experiments CM and CC, referred to as AF and AC experiments respectively. The difference between the experiments AF and AC ($\Delta\text{AF} = \text{AF} - \text{AC}$) enables us to examine feedback from the oceans. An AGCM experiment with the same stratospheric momentum forcing as in CM is also made, where the climatological SST and sea ice in the CC experiment are given as the bottom boundary. This experiment (referred to AM) will represent response ($\Delta\text{AM} = \text{AM} - \text{AC}$) of the atmosphere without ocean feedback, and will also be used for confirming linearity of the coupled response (i.e., $\Delta\text{CM} = \Delta\text{AM} + \Delta\text{AF}$).

The CGCM is integrated for 100 years and the last 50 years are analyzed, and the AGCM is integrated for 40 years and the last 30 years are analyzed, for each experiment. As for time-scale of the forcing, we primarily consider a

timescale longer than multidecadal, specifically the anthropogenic changes of greenhouse gases and ozone, or the secular change of solar activity. However, the result may also be applicable to shorter timescale variation such as the 11-year solar cycle, depending on how quickly the ocean response is built. Our result suggests that a large part of ocean response is built up within a decade by the change of wind-driven circulation as shown later.

3. Response to the stratospheric forcing

3.1 NAM-like atmospheric response

In this section we examine the atmospheric response in the CM experiment (ΔCM). Figure 1 depicts the response in (a) zonal-mean zonal wind, (b) Eliassen-Palm flux (E-P flux) and (c) meridional circulation for January–March mean.

Direct effect of the imposed momentum forcing appears as a strengthening of the westerly jet. However, due to the instantaneous interaction of the planetary waves, it induces a seesaw of the anomalous zonal winds, stronger westerly in the polar region, but weaker westerly in midlatitudes. Evidently a dipole type anomaly in the stratosphere is a coherent direct change in the zonal wind anomaly and meridional circulation. A dipole anomaly is found in the troposphere as well as in the stratosphere; a westerly anomaly in the high latitude and an easterly anomaly in the low latitude with the

node at around the 40°N. The response is statistically significant even in the troposphere.

The E-P flux response exhibits enhanced upward wave propagation in the midlatitude troposphere, and equator-ward refraction in the upper troposphere and the lower stratosphere. The E-P flux anomaly pattern makes a convergence around the 30°N–40°N and a divergence in the north of it. This acceleration and deceleration corresponds to the zonal wind and the meridional circulation responses. Thus, the tropospheric circulation changes seem mainly produced by the change in the tropospheric wave propagation, which is affected by stratospheric circulation change. This response structure resembles that in the observed NAM.

Consistent with the zonal mean structure (Fig. 1), response in sea level pressure (SLP) (Fig. 2) reveals an annular pattern with a low-pressure anomaly in the polar region and a surrounding high-pressure anomaly in the midlatitude. There are two maxima of the positive anomaly in the North Atlantic and in the eastern North Pacific. Most of the region has statistical significance for the response. The response

pattern is similar to the month-to-month NAM in the control simulations (not shown), though the center of negative anomaly exists in the Beaufort Sea rather than around Iceland.

3.2 Oceanic response

The significant response of the troposphere further extends the stratospheric forcing influence to the oceans. Figure 3 indicates the SST response for February–April average; the response pattern basically does not change throughout a year. Response of the North Atlantic SST exhibits positive anomalies in the midlatitude off the east coast United States, the Norwegian Sea, and the Barents Sea and a negative anomaly around the Labrador Sea. The SST anomaly pattern reminiscent of the known North Atlantic tri-pole pattern that is related to NAO (Deser and Blackmon 1993).

This SST response with the NAO-like SLP pattern (Fig. 2) is consistent with interpretations as follows. The intensified westerly wind over the Labrador Sea cools more sea surface, while the anti-cyclonic circulation anomaly in the midlatitude induces the stronger North Atlantic Current (Fig. 3b), which leads to a larger poleward heat transport.

Response of the North Pacific SST accompanies a positive anomaly in the middle to western part of the midlatitude surrounded by negative anomalies extending from the subarctic gyre region to the eastern part of the midlatitude. There is also a significant response in the central-eastern equatorial region, which consequently reveals resemblance with the Pacific Decadal Oscillation (PDO, Mantua et al. 1997). The location of positive SST anomaly in the western Pacific is northward of a corresponding anomaly in the PDO signal. This might be due to a northward bias of the Kuroshio extension region in the used model. This anomalously northward shifted warm SST region might cause effective sensible heat transport to the cold atmosphere in the high latitudes, which may result in overestimation of oceanic forcing to the atmosphere.

Figure 4 illustrates the temporal variation of the ocean response. The temperature change (Figs. 4a, 4b) for different depths (surface, 200 m, 500 m, and 1000 m) around the regions where large responses were seen in the SST field (Atlantic: 60°W–30°W, 40°N–48°N; Pa-

Fig. 2. Sea level pressure response (unit: hPa) in the stratospheric westerly momentum forcing experiment (Δ CM) averaged for January through March. Shadings denote regions with 99% significance.

Fig. 3. (a) Sea surface temperature response (unit: K) averaged for February through April and (b) annual-mean ocean barotropic streamfunction (unit: Sv) in the stratospheric westerly momentum forcing experiment (Δ CM). Shadings denote regions with 99% significance.

cific: 150°E–180°E, 40°N–48°N). It is shown that the Atlantic and Pacific anomalies significant in the SST field extend down to subsurface to thermocline. In the Atlantic, most of the temperature response at 200 m depth is completed within the first several years, and that at 500 m depth is also largely established in the first decade. In the Pacific, on the other hand, the response time seems somewhat longer than in the Atlantic, reflecting the difference of adjustment time depending on the basin width.

Response in the ocean barotropic volume

transport (Fig. 3b) reveals positive and negative anomalies in the northern and southern sides of the subtropical gyres both in the Pacific and the Atlantic. These changes are consistent with wind-driven ocean circulation anomalies implied from the atmospheric circulation changes (Fig. 1, Fig. 2). Anti-cyclonic wind anomalies with axis around 40°N–45°N in the both oceans enhance northward ocean transport around the Kuroshio and the North Atlantic Current. Magnitude of the northward ocean volume transport anomalies at 40°N

Fig. 4. Temporal variations in the CM experiment for the ocean temperature changes in the (a) Atlantic (60°W – 30°W , 40°N – 48°N) and (b) Pacific (150°E – 180°E , 40°N – 48°N) regions at the surface (cross) and depths 200 m (triangle), 500 m (dots), and 1000 m (solid line), and (c) the maximum volume transport changes of the subtropical ocean cells in the North Atlantic (closed square) and the North Pacific (open circle).

(Fig. 4c) are attained to the full response in a shorter period than that of the subsurface temperature anomalies.

4. Feedbacks from ocean

Differences between the AGCM experiments AF and AC ($\Delta\text{AF} = \text{AF} - \text{AC}$) represent feedbacks from the ocean to the NAM-like atmospheric response. Figure 5 illustrates SLP signals in ΔAF for winter (December–February) and spring (March–May) averages. The SLP change reveals characteristic signals in the North Pacific and the North Atlantic depending on seasons, although any annular pattern that strides the pole is not established.

A NAO-like dipole SLP pattern is found in

the North Atlantic in winter. The pattern is coherent within the Atlantic sector to the annular responses in the AM (not shown) and CM (Fig. 2) experiments, though its amplitude is about half of those. It is suggested that there is a feedback from the ocean for the stratospheric forcing in the Atlantic sector.

In the Pacific sector, however, the winter response has no coherency with the stratospheric forced response. In spring, on the other hand, the North Pacific reveals a significant signal that is congruent with the NAM-like pattern, which possibly contributes to the persistent response until late spring with the oceanic feedback.

Seasonality of the feedback is examined (Fig.

Fig. 5. Sea level pressure response (unit: hPa) in the oceanic feedback experiment (ΔAF) averaged for (a) December through February and (b) March through May. Shadings denote regions with 99% significance.

6) by calculating covariances and correlations for each month between the SLP patterns in ΔAF , ΔAM , and ΔCM . Relative importance (or magnitude) of the feedbacks and linearity of the responses are evaluated with the SLP pattern covariances. The Atlantic sector ($60^{\circ}W-0^{\circ}W$, $30^{\circ}N-90^{\circ}N$) and the Pacific sector ($120^{\circ}E-120^{\circ}W$, $30^{\circ}N-90^{\circ}N$) are separately analyzed. It is not easy to make a significance test for these values, since determining the degree of freedom for the spatial variation is difficult. We checked robustness of the result by making the same analysis for half period (25 and 15 years) of each experiment, and found there is no large difference that would alter the conclusion.

In the Atlantic sector, the coupled response (variance of ΔCM) is large through winter to late spring (December through May). The coupled response is mostly reproduced in atmospheric response (the contribution estimated by covariance between ΔAM and ΔCM). However, the coupled response is consistently larger (except for December) than that by the contribution from oceanic feedback (estimated by covariance between ΔAF and ΔCM). This suggests that the coupled response in the Atlantic sector can be explained by a linear combination of the atmospheric response to the forcing and the

oceanic feedback. It is noted that the ratio of amplitude of the ocean feedback to the atmospheric response is larger in December than in midwinter through spring (January to May). For spatial correlations with the coupled response in the Atlantic sector, the oceanic feedback indicates more than 0.6 for November through March and substantially no correlation in summer.

In the Pacific sector, on the other hand, the oceanic feedback exhibits low correlations in winter and high correlations in spring through late summer, except for July. Relative importance of the oceanic feedback is large in spring. Contribution from the oceanic feedback is not detectable in winter, while the atmospheric response is almost as large as the coupled response. There appears some relative importance of the positive feedback in May and June, though the total variance is small in this season.

5. Discussion and conclusions

It is suggested that an intensification of the zonal wind in the winter midlatitude stratosphere induces a NAM-like response extending from the stratosphere to the troposphere. This finding implies that the decadal to interdecadal annular variation of the troposphere is attrib-

Fig. 6. Seasonal variation of the SLP response in the Atlantic sector (60°W – 0°W , 30°N – 90°N) for the experiments AF and AM measured by (a) pattern covariances with the experiment CM [labeled $\text{Cov}(\Delta\text{AF}, \Delta\text{CM})$ and $\text{Cov}(\Delta\text{AM}, \Delta\text{CM})$], and (b) pattern correlation between ΔAF and ΔCM . (c, d) Same as (a, b) except for the Pacific sector (120°E – 120°W , 30°N – 90°N). Variances of SLP response in the experiment CM [labeled $\text{Cov}(\Delta\text{CM}, \Delta\text{CM})$] is also plotted in (a) and (c).

utable (in some part) to the stratospheric zonal wind change caused by external forcings, for instance, ozone heating change due to solar activity, heating by volcanic aerosol, or cooling due to increase of carbon dioxide.

We consider that the connection of stratospheric change to tropospheric change is associated with coherent changes in zonal wind, wave propagation, and meridional circulation. The

major E-P flux divergences are located below 100 hPa (Fig. 1b) where the original momentum forcing is not imposed. This can be explained as follows. Change in mean zonal wind (and vertical shear) due to the momentum forcing leads to change of propagation path of tropospheric waves, which results in the E-P flux divergence and zonal wind acceleration at the lower levels. This is a similar mechanism as

suggested for NAM in previous studies (e.g., Kodera and Kuroda 2000).

The response is substantially larger than the observed month-to-month NAM, and much larger than the observed one for a longer time-scale. As for magnitude, here we showed a “possible extreme,” though we also made additional experiments (not shown) where the maximum forcing of $2.0 \text{ ms}^{-1}\text{day}^{-1}$ (weaker forcing) and $-5.0 \text{ ms}^{-1}\text{day}^{-1}$ (easterly forcing). The results suggest the circulation response is qualitatively similar (though with opposite sign and partly different pattern for the latter case). Accordingly, we believe the qualitative conclusion of the present study does not differ broadly from the results with more realistic forcing.

The NAM-like response in the troposphere influences the underlying oceans through changes of surface wind and heat flux. The SST response exhibits virtually global extent (Fig. 3a). The SST response in the North Atlantic is reminiscent of the tri-pole pattern that is observed associated with NAO (Deser and Blackmon 1993), though the negative anomalies in the Labrador Sea and in the west off Africa are much weaker than the positive anomalies prevailing in most of the North Atlantic. The ocean indicates relatively strong response in the western North Pacific (Fig. 3), even though the SLP response is not significant in that region (Fig. 2). The subtropical ocean gyres (both in the North Atlantic and the North Pacific) are intensified in their northern edge (Fig. 3b) and are shifted northward, in association with the northward intensified subtropical anti-cyclone (Fig. 2). It is implied that ocean heat transport change is more responsible than surface heat flux change for the SST response. For the relative importance between these effects, however, it should be noted that the forcing in the experiment is loaded in the same direction for a long term (i.e., 100 years), which is probably different from the forcing in natural variability.

The SST response (Fig. 3a) resembles the pattern in PDO, which may imply a similar mechanism with PDO in the connection from the extratropical to tropical ocean. One of the possible processes associated with the connection is the “ocean bridge” with advection (Gu and Philander 1997) or wave propagation (Kleeman et al. 1999) of the ocean temperature

anomaly along the shallow overturning cell in the subtropical ocean. However, another “atmospheric bridge” is also possible, since a tropical meridional circulation is evident (Fig. 1c) as a secondary response with the NAM-like response to the stratospheric forcing.

Wiles et al. (2004) suggested century-scale solar variation in association with a modulation of the PDO based on Alaska glacial records. No significant indication is found for relationship between the 11-year solar cycle and the PDO, however Christoforou and Hameed, (1997) pointed out that the observed action center of the Aleutian Low migrates in correlation with the 11-year solar cycle. These may imply an association between the SLP in the North Pacific and the solar activity. The cause of the PDO modulation is not fully understood. Schneider and Cornuelle (2005) argue that the signal consists of several responses from different forcings, such as variations of the Aleutian Low, the Kuroshio extension, and ENSO. These subjects should be pursued in a future study.

By analyzing AGCM experiments with the ocean boundary from the CGCM experiments, the oceanic response has feedback effects onto the initial atmospheric circulation response. The oceanic feedback effects reveal different features between the Atlantic and Pacific sectors. In the Atlantic sector, there is a positive feedback in winter that amplifies the NAO-like SLP pattern. In the AF experiment, the outstanding pattern with a positive SLP anomaly in MAM in the north Pacific (Fig. 5b) reveals high correlation (~ 0.8 in May) with the pattern in the CM experiment of the same season (Fig. 6d). This suggests that the atmospheric response to the SST anomaly formed by winter forcing becomes large in late spring rather than midwinter, which means a possible time lag between the forcing and the ocean feedback.

The oceanic feedback is small in midwinter, compared to the direct response of the atmosphere, whereas it is relatively large in early winter in the North Atlantic and in spring in the North Pacific. This seasonality suggests a relationship between the oceanic condition and storm-track activity. It is conjectured that oceanic forcing to the atmosphere is reinforced by storm track activity. The seasonal dependency of the storm track activity is shown by Nakamura (1992). He showed that the storm track

activity is suppressed in midwinter over the Pacific, while it is not over the Atlantic. For the response in the Pacific, influence from the tropical Pacific through tele-connections should also be considered. This should be studied in future experiments with SST forcing in separate regions.

It is shown that a stratospheric forcing, such as solar activity, volcanic eruptions, or increase of carbon dioxide, possibly causes a NAM-like long-term variation. In addition, it is implied that the forcing further influences the decadal to interdecadal variability of oceans with feedback from ocean to atmosphere. To validate these speculations, more realistic and quantitative estimation of forcing (e.g., seasonality of ozone variation and 11-year cycle of solar activity) and detailed analyses will be required in future studies.

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