Impact of the wintertime North Atlantic Oscillation (NAO) on the summertime atmospheric circulation

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[1] Using the NCEP/NCAR reanalysis dataset and other observations, we show that the summer high-latitude climate in the Northern Hemisphere is influenced by the NAO of the previous winter. We find this influence in the summertime surface air temperature, the geopotential height, the sea surface temperature (SST), sea-ice/continental snow cover extent fields as well as in the zonal mean geopotential height and zonal wind fields. This summertime NAO signal is annular but its meridional scale is much smaller than the winter annular mode. Distinct summer anomalies are located at the nodal latitudes of the winter anomalies. We suggest that the sea-ice, SST and snow cover anomalies provide the memory allowing the winter NAO to affect the summer climate. *INDEX TERMS:* 1620 Global Change: Climate dynamics (3309); 1863 Hydrology: Snow and ice (1827); 3309 Meteorology and Atmospheric Dynamics: Climatology (1620); 3322 Meteorology and Atmospheric Dynamics: Land/atmosphere interactions. **Citation:** Ogi, M., Y. Tachibana, and K. Yamazaki, Impact of the wintertime North Atlantic Oscillation (NAO) on the summertime atmospheric circulation, *Geophys. Res. Lett.*, *30*(13), 1704, doi:10.1029/2003GL017280, 2003.

1. Introduction

[2] The North Atlantic Oscillation (NAO), the simultaneous strengthening and weakening of the Icelandic low and the Azores high [e.g., *van Loon and Rogers*, 1978; *Wallace and Gutzler*, 1981], is the most powerful phenomenon in winter and dominates the climate variability over the wintertime Northern Hemisphere, especially over northeast America, the Atlantic and Eurasia. Many studies have been done on the influence of the NAO upon the regional and hemispheric climate in winter [e.g., *Hurrell*, 1995, 1997; *Kodera et al.*, 1999; *Rodwell et al.*, 1999; *Xie et al.*, 1999]. Recently, emphasizing a more hemispheric circulation pattern, *Thompson and Wallace* [1998, 2000] introduced the Arctic Oscillation (AO) or the Northern Annular Mode (NAM). The AO also has an impact on the winter climate of the Northern Hemisphere in interannual and longer timescales [e.g., *Thompson and Wallace*, 2001]. However, few studies have been performed on the impact of the wintertime NAO/AO on the summertime climate. *Rigor et al.* [2002] showed that the summertime sea ice concentrations over the Arctic Ocean are linked to the phase of the AO in the previous winter. However, their study is limited to the Arctic region. The impact of the wintertime NAO/AO on the summertime atmospheric circulation in the hemispherical scale has not yet been clarified.

[3] The purpose of this study is to illustrate the impact of the winter NAO/AO on the summertime atmospheric circulation in the extratropics. Although the NAO index is used as an index for winter circulation in this study, the results of the AO are similar to those of the NAO.

2. Data and Method

[4] Atmospheric data used in this study is the NCEP/NCAR reanalysis dataset for the period from 1958 to 2000 [*Kalnay et al.*, 1996]. In this study winter is defined as December, January and February (DJF). The NAO index is defined as the SLP difference between Stykkisholmur, Iceland and Ponta Delgada, Azores by *Hurrell* [1995] and we use the period from 1958 to 2000. The monthly mean sea surface temperature (SST) and sea-ice concentration dataset used in this study is the global sea ice coverage and SST data (GISST2.3b) with a $1^\circ \times 1^\circ$ resolution from 1958 to 2000. The snow cover frequency is based on the digital NOAA-NESDIS Weekly Northern Hemisphere Snow Charts from 1971 to 1995, compiled by *Robinson et al.* [1993].

[5] In order to clarify the relation between the winter NAO and the spring and summer atmospheric circulation in an interannual time scale, we calculate the correlation coefficients and the linear regression coefficients of the spring and summer fields with the preceding winter NAO index. The statistical significance is determined by t-test assuming sample data are independent.

3. Winter NAO and Spring-Summertime Atmospheric Circulation

[6] To grasp the scope of the impact of the winter NAO on the subsequent months, we present lag correlations of the monthly averaged zonal mean 500-hPa geopotential height with the winter NAO index in Figure 1. In January and February, a seesaw between 40–50°N and 60–75°N is dominant. After February, the location of positive maximum shifts poleward and the correlation is the weakest in March. In April, the positive correlation located in the nodal latitudes of the winter anomaly increases again and the Arctic negative correlation reappears in June. From this temporal variation of correlation coefficients, we define summer as May, June and July (MJJ) and the transient season, i.e., spring as March and April (MA).

[7] Horizontal patterns in spring and summer related to the winter NAO are shown in Figure 2. The spring SLP field (Figure 2a) shows a seesaw between a positive area

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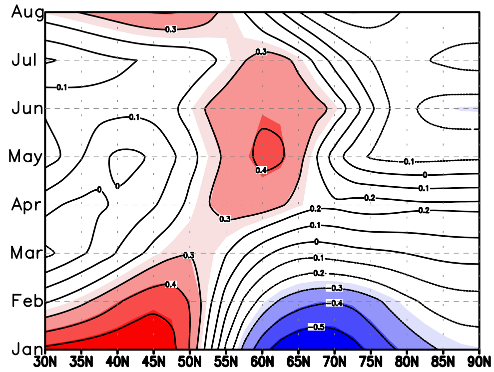

Figure 1. Lag correlation of the monthly averaged zonal mean 500-hPa geopotential height with the winter NAO. Light, middle light, and heavy shadings indicate that correlation coefficients exceed the 90%, 95%, 99% confidence levels, respectively.

over the North Atlantic Ocean and a negative area centered over the Greenland Sea. The 500-hPa height field (Figure 2b) also shows a similar seesaw pattern over the North Atlantic Ocean and the Greenland Sea and it can be seen throughout the troposphere. The spring NAO signal shows an equivalent barotropic structure except for central Siberia. The warm 850-hPa temperature signal (Figure 2c) is observable over the North Atlantic and extends eastward throughout northern Eurasia and the largest signal is found over central Siberia. The surface air temperature (2-m temperature) shows a similar pattern. The strong warm signal over central Siberia generates a significant positive height anomaly at 500-hPa there. The spring NAO signature is similar to the winter NAO signature [*Hurrell*, 1995].

[8] The summer SLP and 500-hPa height signatures (Figures 2d and 2e) are characterized by a seesaw pattern between the Arctic and the subarctic region; a negative correlation area in the Arctic and a positive correlation in the subarctic, especially over the British Isles, central Siberia and the northern Sea of Okhotsk. The anomalies in the northern Sea of Okhotsk are related to the Okhotsk high, which is a stationary anticyclone that sometimes occurs in summer over the Sea of Okhotsk. Because the Okhotsk high influences monsoon rainfall in Japan, Korea and China [e.g., *Ninomiya and Mizuno*, 1987], many studies on the Okhotsk high have been performed, motivated by local interests. The 500-hPa height signals (Figure 2e) show that Arctic negative regression/correlation values are surrounded by positive regression/correlation values. In particular, positive values over the British Isles, central Siberia, eastern Siberia and subarctic North America are significant. These features can be seen throughout the troposphere. The summertime circulation pattern is annular but its meridional scale is smaller than that of the winter AO.

[9] Here, we discuss the field significance of the results. The spatial area above 90% (95%) statistical significant level for SLP in MJJ is 12% (7%) of the analysis region (north of 35°N), which is larger than 10% (5%), the expected percentage that occurs by chance. The corresponding percentage for 500-hPa height is 15% (9%). These percentages of significant areas increase when the analysis area is restricted to higher latitudes. The areas exceeding the 90% statistical level shown in other figures (Figures 1, 2, 3, and 4) are also above 10%. Therefore, we can not conclude that the correlation patterns occur by chance, though the percentage of 12% for SLP, 90% significant level is close to 10%.

[10] The regression/correlation map between the winter NAO and summertime 850-hPa temperature (Figure 2f) displays large positive values over the circumpolar zone, especially over northwestern Europe, central Siberia, and northeastern Siberia. That is, when the winter NAO is in a

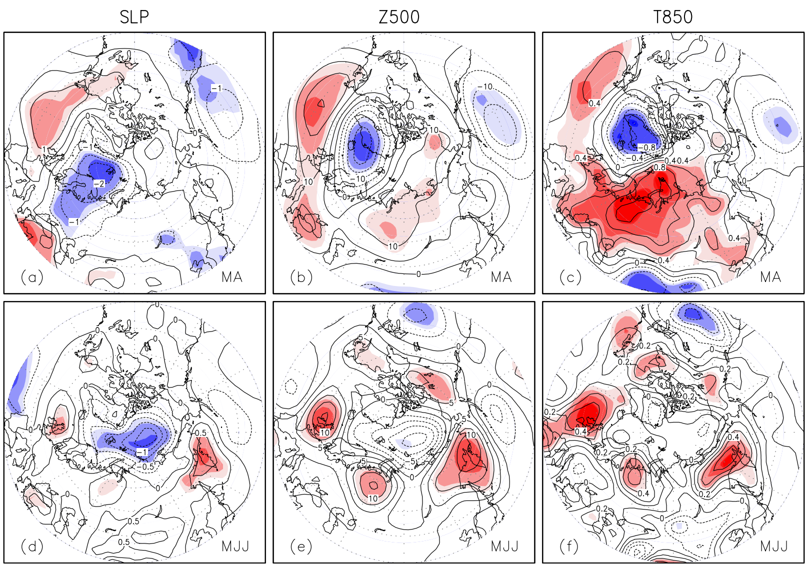

Figure 2. Maps for regression coefficients of spring (MA) and summer (MJJ) atmospheric conditions with the NAO in the previous winter (DJF). (a) SLP in spring, (b) 500-hPa height in spring, (c) 850-hPa temperature in spring, (d) SLP in summer, (e) 500-hPa height in summer, (f) 850-hPa temperature in summer. Contours are regression values and the intervals in summer are halves of those in spring. The shadings denote the statistical confidence levels calculated from correlation coefficients and they are the same as in Figure 1.

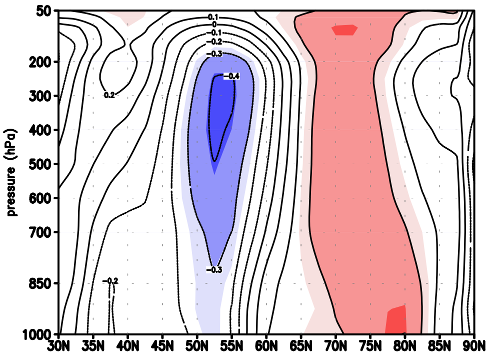

Figure 3. Latitude-height cross section of correlation coefficients between the zonal mean zonal wind in summer and the winter NAO index. The shadings are the same as in Figure 1.

positive phase, the temperatures in the circumpolar regions tend to be warmer. The areas of the maximum anomalies in temperature are consistent with the areas of the maximum anomalies in geopotential height.

[11] Figure 3 shows correlation coefficients between the zonal mean zonal wind in summer and the winter NAO index. A dipolar structure, with positive correlations located at 70–75°N and negative ones at 50–55°N, is clear. That is, when the winter NAO is in a positive phase, summertime westerly winds at 70–75°N tend to be stronger and those at 50–55°N tend to be weaker. Around 40°N, where the climatological subtropical jet is located, correlations are nearly zero. This indicates a double jet structure, which is closely related to atmospheric blocking events, appearing in summer after the positive NAO winter.

[12] Since the summer signal is annular, the winter-to-summer linkage might be caused by a persistence of the winter annular mode. *Thompson and Wallace* [2000] defined the AO index for all the months. The NAO index defined by *Hurrel* [1995] is also obtainable for all the months. It is valuable to examine the decay time scales of the winter NAO and AO indices as a function of the calendar month. Table 1 shows such auto-correlations from January through July with the winter NAO/AO indices. After February, the correlation coefficients drop rapidly and lose their statistical significances after a month or two. Therefore, the linkage between the winter NAO and the summer atmosphere found in this study is not a simple persistence of the specific atmospheric pattern; it accompanies a structural change.

4. Winter NAO and Spring-Summer Snow Cover, Sea Ice and SST

[13] The results in the previous section show that the winter NAO influences spring-summertime atmospheric circulation. Since the atmosphere itself does not have long memory beyond one month and the winter NAO does not have a significant auto-correlation after March, something else that has longer memory should link winter to summer. Surface ocean, sea-ice and snow cover are possible candidates for this long memory. Figure 4a shows the correlation between the winter NAO and the spring SST, sea-ice concentration and snow cover. The warm color in the figure denotes a warmer SST, less sea-ice and less snow cover for the positive NAO index in winter. Colder SSTs and more sea ice are found in the North Atlantic and the Labrador Sea. This cold area is surrounded by warm SSTs to the south and east and by less sea-ice over the Greenland Sea and the Barents Sea. These SST and sea-ice anomalies associated with the NAO are well-known by previous studies [*e. g., Rodwell et al.*, 1999, and references therein] and they persist through summer (Figure 4b). This indicates that the influence of the wintertime NAO is memorized in the ocean and the sea ice, both of which have larger thermal inertia than the atmosphere. For the Arctic sea ice, *Rigor et al.* [2002] pointed out that a dynamical process through wind forcing and ocean current is also important and our result on the summer Arctic sea ice is qualitatively similar to Figure 13b of theirs.

[14] Another candidate for the long memory is snow cover [e.g., *Watanabe and Nitta*, 1999, *Rajeevan*, 2002]. Negative anomalies in snow cover over western Eurasia and central Asia in spring (Figure 4a) and those over eastern Siberia in summer (Figure 4b) are statistically significant.

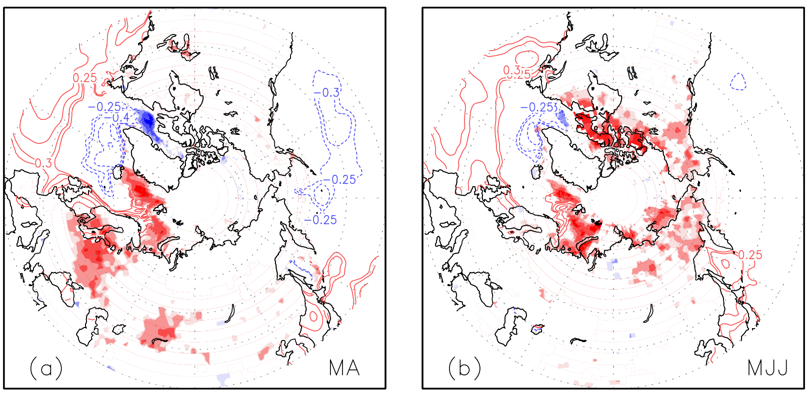

Figure 4. (a) Correlation coefficients between winter (DJF) NAO and the spring (MA) SST, sea ice and snow cover. (b) Same as in (a) but for summer (MJJ) SST, sea ice and snow cover. Contours over oceans are correlations of the SST. Solid (red) and broken (blue) lines show the positive and negative correlations, respectively. The shadings over oceans are correlations with sea-ice. The shadings over land are correlations with snow cover. The denotations of the shading areas are the reverse of Figure 1 (red denotes negative, blue denotes positive).

Table 1. The Auto-Correlation Coefficients of the Monthly NAO/AO Indices From January Through July With the Winter NAO/AO Indices

	Jan	Feb	Mar	Apr	May	Jun	Jul
NAO	0.75	0.61	0.16	−0.15	−0.17	0.15	−0.20
AO	0.87	0.77	0.30	0.00	0.09	0.00	0.21

Because eastern Siberia is mountainous, the climatological snowmelt season there is delayed relative to other regions and the snow cover signal over eastern Siberia appears in summer. The snow cover anomalies over subarctic North America in summer are also negative. Early snowmelt over western Eurasia is considered to be caused by warm temperature and less snowfall during winter [e.g., *Serreze et al.*, 1997]. In general, the timing of snowmelt in the circumpolar regions in both continents is earlier after the positive phase of the winter NAO than after the negative phase.

[15] Snow cover affects local atmospheric heating through snow-albedo feedback during the melting season. Even after snowmelt, snow cover affects the heating through soil moisture. These snow cover and ground hydrological processes are in agreement with *Yasunari et al.* [1991], in which the effect of the excessive snow over the Eurasian continent (30–60°N) in the beginning of March upon the Asian climate in spring and summer was investigated by a GCM. *Kodera and Chiba* [1989] calculated correlation coefficients between snow cover in April over central Asia and 500-hPa height in June, based on the data for 1968–86. They found that the snow anomalies in spring affect the climate in Eurasia and North America. Their results generally agree with our results. In particular, Figure 15 in *Yasunari et al.* [1991] is quite similar to ours (Figure 2e).

[16] Using the same method as was done for the NAO, we investigated the relationship between the wintertime AO and the summertime atmospheric circulation. The summertime patterns correlated with the wintertime AO are similar to those with the NAO.

5. Discussion and Conclusions

[17] We investigated the influence of the wintertime NAO on the subsequent spring and summer atmospheric circulation in the extratropics with regression and correlation analyses based on the NCEP/NCAR reanalysis dataset. It was found that when the wintertime NAO is in a positive phase, the summertime surface air temperatures over circumpolar regions in northern Eurasia and subarctic North America are warmer and the geopotential height is higher and vice versa. Although this summertime NAO signal appears similar to the winter annular mode, its structure is quite different from the winter one; the meridional scale of the summertime NAO signal is much smaller than the winter annular mode.

[18] The SST and sea-ice anomalies over the North Atlantic Ocean and the Arctic Ocean persist from winter through summer. In addition, the snow-cover anomalies over the western Eurasia and central Asia in spring and those over eastern Siberia in summer are significant. We suggest that the signal of the wintertime NAO is memorized in the snow, sea ice and the ocean surface in the circumpolar regions, and these anomalies then influence the summertime atmospheric circulation in the extratropics.

[19] For seasonal forecast, the ENSO has been considered as a primary factor so far. The present study suggests the possibility of seasonal forecast for summer climate from the winter NAO through spring cryosphere such as snow. To this end, the climate model must have a realistic ground hydrological sub-model.

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