

A Skillful Prediction Model for Winter NAO Based on Atlantic Sea Surface Temperature and Eurasian Snow Cover

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ABSTRACT

A new statistical forecast scheme, referred to as scheme 1, is developed using observed autumn Atlantic sea surface temperature (SST) and Eurasian snow cover in the preceding autumn to predict the upcoming winter North Atlantic Oscillation (NAO) using the year-to-year increment prediction approach (i.e., DY approach). Two predictors for the year-to-year increment are identified that are available in the preceding autumn. Cross-validation tests for the period 1950–2011 and independent hindcasts for the period 1990–2011 are performed to validate the prediction ability of the proposed technique. The cross-validation test results for 1950–2011 reveal a high correlation coefficient of 0.52 (0.58) between the predicted and observed NAO indices (DY of the NAO). The model also successfully predicts the independent hindcasts for the period 1990–2011 with a correlation coefficient of 0.55 (0.74). In addition, scheme 0 (i.e., anomaly approach) is established using the SST and snow cover anomalies during the preceding autumn. Compared with scheme 0, this new prediction model has higher predictive skill in reproducing the interdecadal variability of NAO. Therefore, this study provides an effective climate prediction scheme for the interannual and interdecadal variability of NAO in boreal winter.

1. Introduction

The North Atlantic Oscillation (NAO) is the dominant pattern of atmospheric circulation variability in the Northern Hemisphere and it has a significant impact on the temperature and precipitation throughout the whole of Europe and East Asia (Hurrell 1995; Rodwell et al. 1999; Sun et al. 2008; Sun and Wang 2012; Tian and Fan 2012; Wang and Chen 2012; Kim et al. 2013). For this reason, there is growing interest in the predictability and ability of numerical climate models to simulate the NAO (Bojariu and Gimeno 2003a). An assessment of the skill of a set of wintertime NAO seasonal predictions within a multimodel ensemble framework has already

been carried out (Doblas-Reyes et al. 2003). Using the perfect model approach for the period 1959–2001, it has been shown that the mean winter NAO index is potentially predictable with a lead time of 1 month (Müller et al. 2005). Scaife et al. (2014) also demonstrated that the winter NAO is highly predictable months ahead using the Met Office Global Seasonal Forecast System 5.

The sea surface temperature (SST) in the North Atlantic in preceding months has been used as a predictor of the upcoming winter NAO. Many previous studies have pointed out that a positive NAO phase is associated with a positive SST anomaly southeast of Newfoundland and a negative SST anomaly to the northeast and southeast over the North Atlantic during several preceding months (Czaja and Frankignoul 1999, 2002). The seasonal predictability of the winter NAO has been examined from the perspective of lagged North Atlantic SST anomalies, which are significantly correlated to the next winter's NAO indices (Saunders and Qian 2002). In

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terms of atmospheric predictability, analysis has shown that significant links exist between wintertime NAO and SST anomalies in the preceding spring, summer, and autumn (Rodwell and Folland 2002). Meanwhile, Eurasian snow cover is also a predictor when forecasting winter NAO. Early season snow cover, which has a greater impact on winter atmospheric circulation than snow cover during the same period, is critical for understanding and predicting past, present, and future Northern Hemisphere climate (Cohen and Entekhabi 1999). Snow cover variability helps to explain the interannual variability, and can be regarded as a significant forcing mechanism of the Northern Hemispheric general circulation (Cohen and Entekhabi 2001). Because continental-scale snow cover variation can lead the atmosphere by several months through their mutual oscillations, snow cover is proposed as a potential contributor to the interannual variability of the leading boreal winter mode of the atmosphere (Saito and Cohen 2003; Saito et al. 2001; Gong et al. 2003; Peings et al. 2013). A statistical forecast model has been developed using observed October mean snow cover and sea level pressure anomalies to predict upcoming winter Arctic Oscillation (AO; Cohen and Fletcher 2007).

The year-to-year increment prediction approach utilizes the difference between the current year and the previous year as the predictand. In this approach, the predicted variable Y is obtained by adding the predicted DY of the variable to the observed variable of the preceding year; thus, the predicted Y (current year) = predicted DY + observed Y (preceding year).

As we know, the seasonal climate variability includes both the interdecadal and interannual variability. But it is very difficult to capture the interdecadal variability of the variable. As the predicted variable is obtained by adding the predicted year-to-year increment to the observation for the preceding year that contains the real interdecadal component, the interdecadal prediction of the climate variable can be well captured in the increment approach.

In addition, the amplitude of the interannual variability for the year-to-year increment of the variable is much larger than that of the variable itself. The signals of the predictors and predictand are both amplified (Wang et al. 2010) and, hence, help to improve the prediction skill. In past years, the year-to-year increment approach has been used to explore the prediction skill for summer rainfall in China (Fan et al. 2008), wintertime heavy snow activity in northeast China, and the East Asian summer monsoon (Fan et al. 2012; Fan and Tian 2013), all resulting in improved skills. Since the power spectrum for the time series of the NAO had been shown to exhibit spectral peaks at

the quasi-biennial period (2–3 yr) (Stephenson et al. 2000; Marshall et al. 2001; Saito and Cohen 2003), we are motivated to apply the incremental approach to the seasonal prediction of winter [December–February (DJF)] NAO in this research. The year-to-year increments of the North Atlantic SST and Eurasian snow cover in the preceding autumn [September–November (SON)] are selected as the predictors, and we use both the cross-validation and independent retrospective hindcasts to evaluate the new prediction scheme.

2. Data and methods

In this study, monthly mean sea level pressure data at $2.5^\circ \times 2.5^\circ$ horizontal resolution are derived from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis dataset for the period 1950–2011. Extended reconstructed SST, version 3b (ERSST.v3b; $2.0^\circ \times 2.0^\circ$ resolution), from the National Oceanic and Atmospheric Administration (NOAA) are also used (Smith et al. 2008). The monthly snow cover data are from the Twentieth Century Reanalysis (20CR; Compo et al. 2011). The NAO index during DJF is defined using the time series of the leading empirical orthogonal function (EOF) of sea level pressure anomalies over the Atlantic sector (20° – 80° N, 90° W– 40° E) during the period 1950–2011 (Hurrell 1995).

First, we investigate the relationship between the DY of wintertime NAO and the DY of North Atlantic SST and Eurasian snow cover in the preceding autumn to identify two predictors to build the prediction model for winter NAO. Then, cross-validation tests and independent hindcasts are performed to verify the statistical scheme for forecasting boreal winter NAO.

3. The predictors

To choose the predictors for the new scheme, we consider the SST in the North Atlantic and the Eurasian snow cover in the preceding autumn, which both have a significant impact on winter NAO. Thus, we identify two predictors based on an analysis of the DY of SST in the North Atlantic and Eurasian snow cover with the DY of NAO.

Previous studies have indicated a relationship between the tripole pattern SST anomalies and the phase of NAO (Czaja and Frankignoul 2002; Saunders and Qian 2002; Feddersen 2003). The tripole SST signal may then project upon the overlying 500-hPa geopotential height variability via transient eddy fluxes, reinstating the previous winter's phase of the NAO (Rodwell and Folland 2002). Figure 1 shows the correlation coefficients

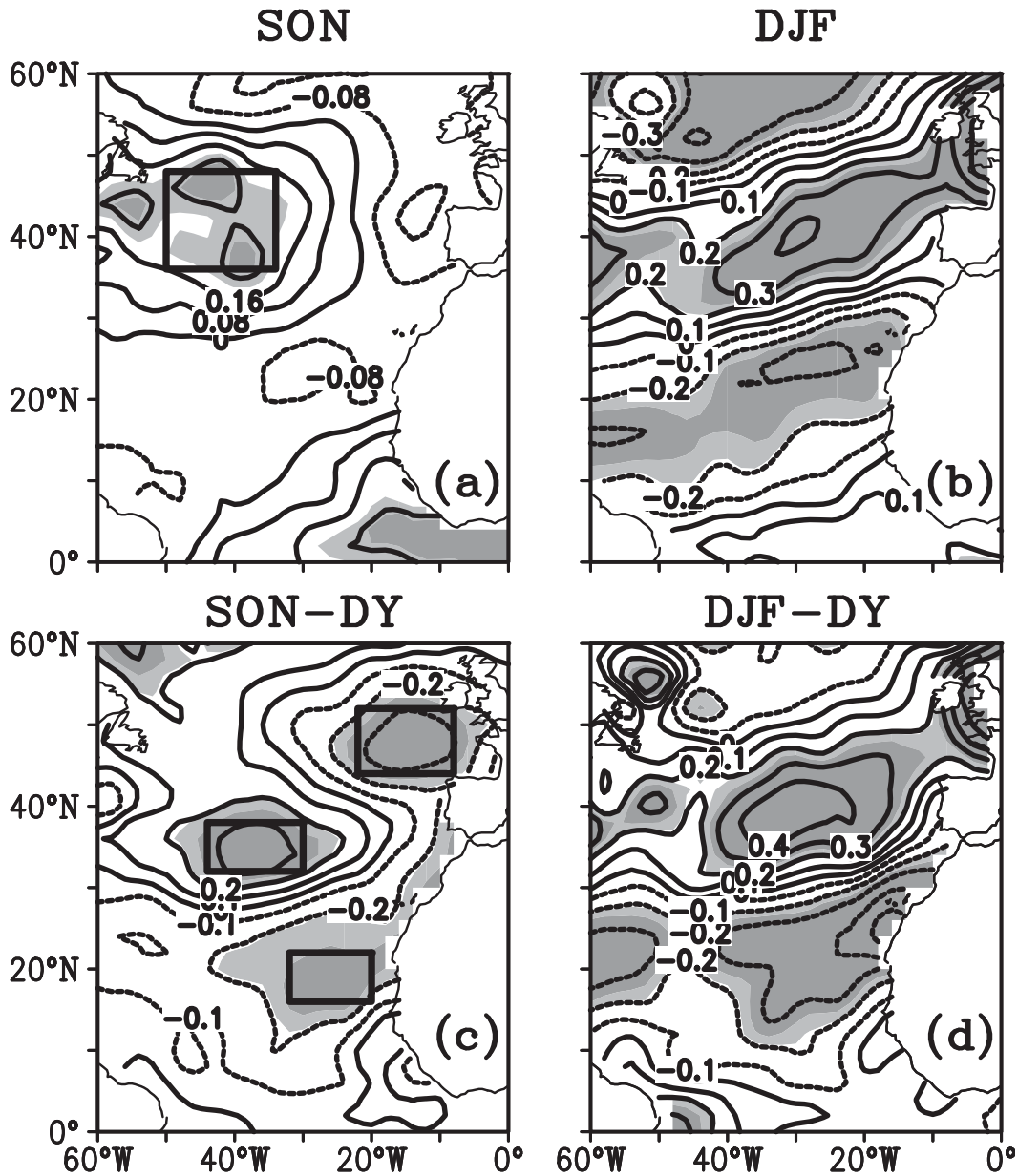


FIG. 1. Correlation coefficients between the DJF NAO index and SST in (a) SON and (b) DJF; the DY form in (c) SON and (d) DJF, for the period 1950–2011. The shaded areas indicate significance at the 90% (light shaded) and 95% (dark shaded) levels estimated by a local Student's *t* test.

between the DJF NAO index and the SST in SON and DJF for the period 1950–2011. It is found that the correlation between the SON SST and the DJF NAO is only significant when year-to-year increments are considered; the relationship is insignificant when the original form is used (see Figs. 1a and 1c). The forcing of SST on the NAO can continue from autumn to the following winter (Figs. 1c and 1d).

The predictor x_1 is defined by the difference of the area-averaged DY of SST in the preceding autumn

between the area of positive correlation in the mid-latitudes and the areas of negative correlation in the high and low latitudes. The average DY of SST fields in the North Atlantic include the curvilinear rectangles (32°–38°N, 30°–44°W; 44°–52°N, 8°–22°W; 16°–22°N, 20°–32°W) in Fig. 1c. The correlation coefficient between the predictor x_1 and observed DY of NAO for the period 1950–2011 is 0.38, above the 99% significance level.

Numerous subcontinental-scale studies of European snow cover have been carried out and have yielded negative

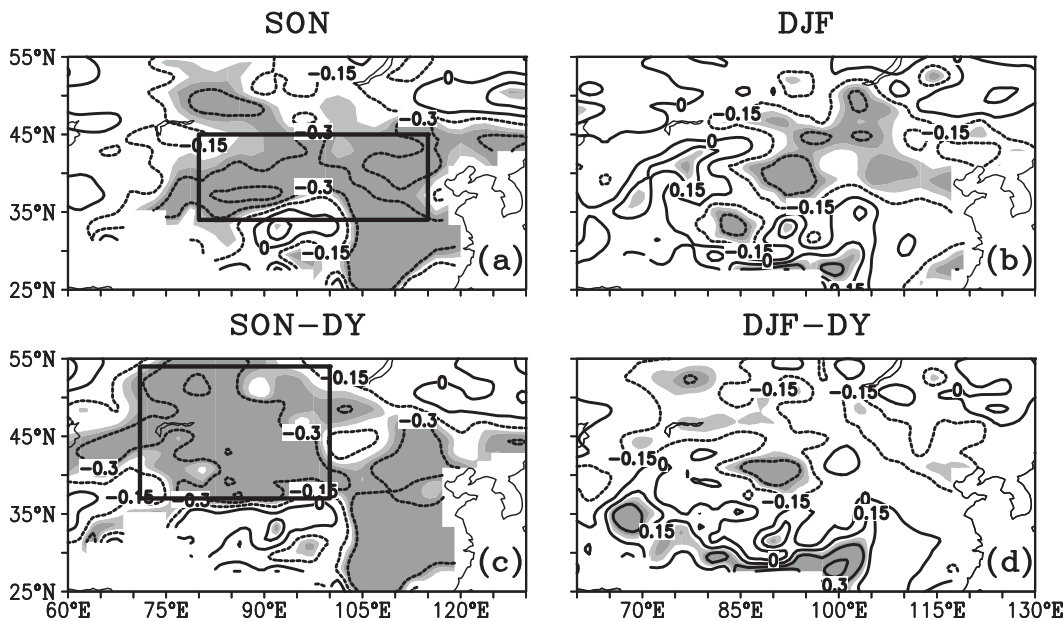


FIG. 2. Correlation coefficients between the DJF NAO index and snow cover in (a) SON and (b) DJF; the DY form in (c) SON and (d) DJF, for the period 1950–2011. The shaded areas indicate significance at the 90% (light shaded) and 95% (dark shaded) levels estimated by a local Student's *t* test.

correlation relationships between snow cover variability and the phase of the NAO pattern (Falarz 2004). The high albedo of snow, which reduces the total incoming solar radiation and hence the surface temperature, plays a role, especially in the lower latitudes where incoming radiation is extremely strong. The responses of the Northern Hemispheric atmosphere to these continental processes seem to be partly nonlocal, leading to the excitation of NAO

circulation patterns (Bojariu and Gimeno 2003b). The intensification of the Siberian high, along with the thermal impacts of enhanced snow cover and topographic forcing, could lead to stratospheric warming and to the January tropospheric negative AO response (Cohen et al. 2007). Figure 2 shows that the autumn snow cover anomaly can persist into the following winter through a positive feedback between snow cover and the atmosphere.

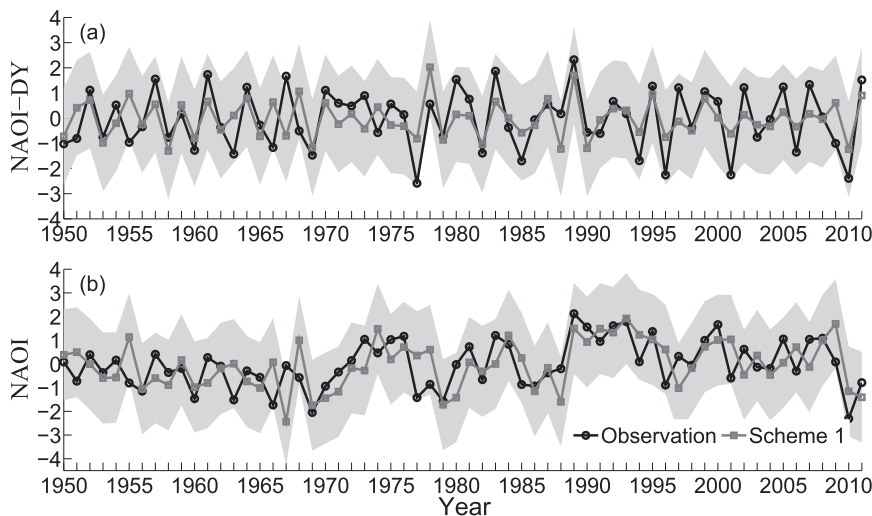


FIG. 3. Predicted (gray line) and observed (black line) (a) DY of NAO and (b) NAO in the cross-validation test for the period 1950–2011. The gray shading is the 95% prediction interval.

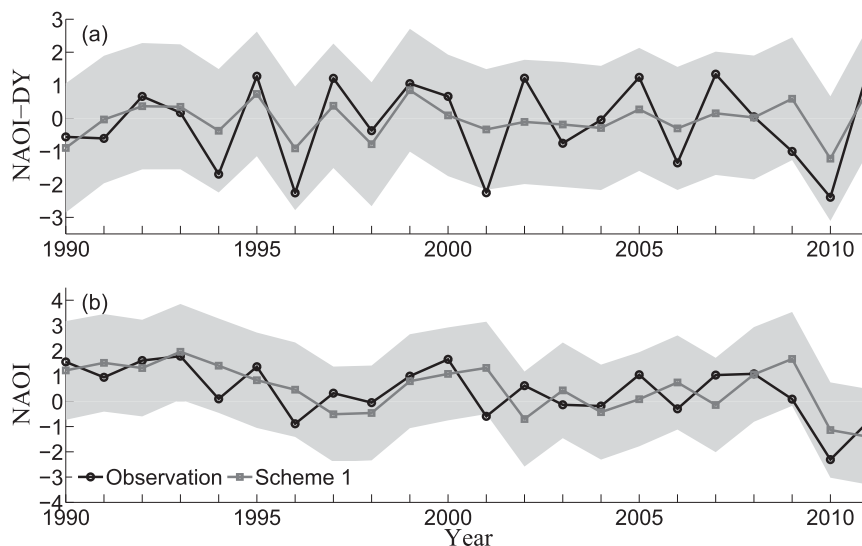


FIG. 4. Predicted (gray line) and observed (black line) (a) DY of NAO and (b) NAO in the independent hindcasts for the period 1990–2011. The gray shading is the 95% prediction interval.

The predictor x_2 is defined as the area-averaged DY of snow cover over the region (37° – 54° N, 71° – 100° E) in the preceding SON (see Fig. 2c). The correlation coefficient between the predictor x_2 and the observed DY of NAO for the period 1950–2011 is -0.46 , exceeding the 99% significance level. It is found that x_1 is insignificantly correlated with x_2 , and the correlation coefficient is only 0.06. That is to say, the tripole pattern SST over the North Atlantic and the Eurasian snow cover in the preceding autumn should be viewed as an independent external forcing affecting wintertime NAO.

4. The prediction model and its validation

In this section, a new prediction model is established using the multilinear regression analysis method. Both cross-validation tests for 1950–2011 and independent hindcasts for 1990–2011 are performed to assess the skill

of our new forecast model. The cross-validation tests for the period 1950–2011 can be applied by removing each year in turn, and building a new regression model using the x_1 and x_2 from the remaining 61 yr to produce a hindcast for the DY of the NAO of the withdrawn year (Michaelsen 1987). Using this process, we can evaluate the prediction skill by the correlation between the hindcasts and observations for the period 1950–2011. To assess the NAO forecast model, we also apply an independent hindcast for the period 1990–2011, similar to the real-time forecast. During the training period of 1990–2011, the real-time forecast models are built using the x_1 and x_2 from 1950 to the year before the forecast year. For example, we use the 1950–2000 dataset to forecast the 2001 values. Based on the cross-validation test, the predicted DY of NAO agrees well with the observed DY of NAO (Fig. 3a). The correlation coefficient between the predicted and observed DY of

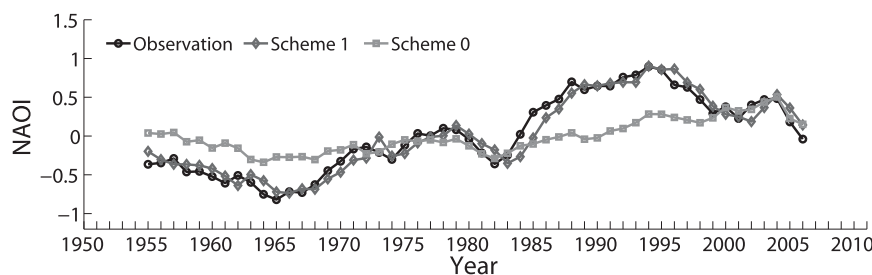


FIG. 5. The 11-yr running averages of NAO for the observations (black line), scheme 1 in the cross-validation test (dark gray line), and scheme 0 in the cross-validation test (light gray line).

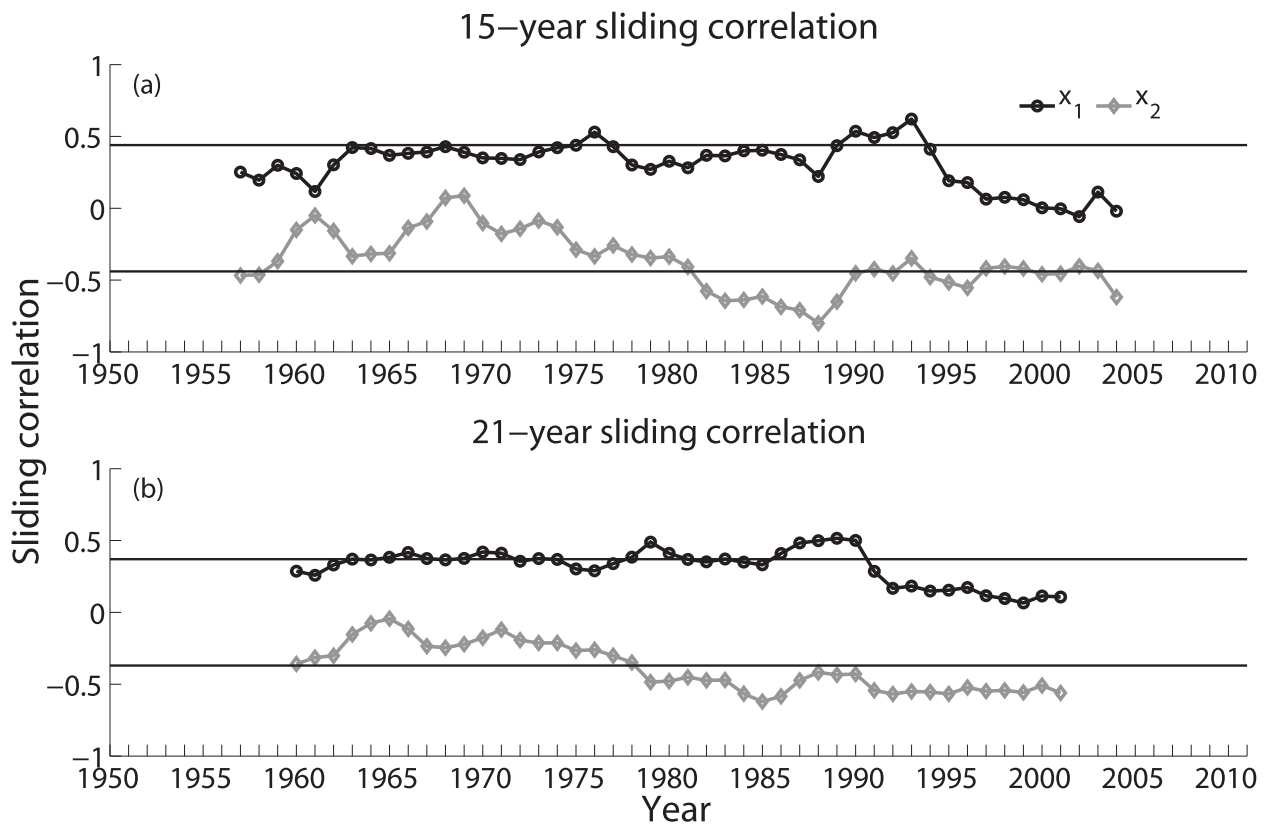


FIG. 6. The (a) 15-yr and (b) 21-yr sliding correlation coefficient between the predictor x_1 and the NAO in winter (black curve lines) and between the predictor x_2 and the NAO in winter (gray curve line). Straight lines show the significance level for correlation coefficients at 0.01.

NAO from 1950 to 2011 is 0.58, accounting for 34% of the total variance. Figure 3b shows the predicted NAO in the cross-validation test, which can be obtained by adding the predicted DY of NAO to the observed NAO during the previous year. The gray shading is the 95% prediction interval around the cross-validation test for 1950–2011. The winters of 1955 and 1967 are the only two years in which the observed NAO index does not fall within the predictive range. The correlation coefficient between the predicted and observed NAOs during the period 1950–2011 is 0.52, accounting for 27% of the total variance, with a root-mean-square error (RMSE) of 0.96.

As is shown in Fig. 4, scheme 1 performs well in the independent hindcasts for the period 1990–2011. The correlation coefficient between the predicted and observed DYs of NAO from 1990 to 2011 is 0.74, accounting for 55% of the total variance. The replicated real-time predicted NAO is closely related to the observed NAO during the period 1990–2011, with a correlation coefficient of 0.55, accounting for 30% of the total variance, and an RMSE of 0.93. Only during the winter

of 2001 does the observed NAO index not fall within the predictive range.

To verify the skill of the prediction model with respect to the interdecadal variability of the NAO, we calculate an 11-yr running average of NAO from the observations and scheme 1 in the cross-validation test for the period 1950–2011 (see Fig. 5). The correlation coefficient between them reaches 0.97 and the RMSE is only 0.13 for the period 1955–2006. Scheme 1 has considerable ability in reproducing the interdecadal variability of NAO. In addition, we also establish scheme 0 using the SST and

TABLE 1. Verification for the scheme used in the cross validation for 1981–2000. Scheme-SST denotes the scheme based on the DY of SST. Scheme-SST-SNOWC denotes the scheme based on the DY of SST and DY of snow cover. TCC refers to the time correlation coefficient between the predicted value and the observed value.

	TCC (significance level)	RMSE (decrease in RMSE)
1981–2000		
Scheme-SST	0.63 (99.84%)	0.83
Scheme-SST-SNOWC	0.78 (99.99%)	0.64 (23%)

snow cover anomalies instead of their year-to-year increments in the preceding autumn. The two predictors that we use are the area-averaged SST anomalies over the North Atlantic (36°–48°N, 34°–50°W; see Fig. 1a) and the snow cover anomalies over Eurasia (35°–45°N, 80°–110°E; see Fig. 2a). The 11-yr running average of NAO for scheme 0 in the cross-validation test for the period 1950–2011 is also shown in Fig. 5. The correlation coefficient between the 11-yr running average of NAO for the observations and for scheme 0 in the cross-validation test is 0.74, and the RMSE is 0.36 for the period 1955–2006. When compared with the new scheme, the NAO interdecadal variation prediction skill of the old scheme declines significantly.

We also compare scheme 1 with the prediction scheme of winter NAO (i.e., scheme-SST) that is established using two lagged modes of the North Atlantic SST variability from June to October, as proposed by Saunders and Qian (2002). For the cross-validation tests for the period 1950–2000, similar predictive skill is found in the two schemes. The correlation coefficient between the observed NAO and the predicted NAO is 0.55 for scheme 1, and 0.63 for the scheme-SST. For the independent hindcasts for the period 1986–2000, scheme 1 has higher predictive skill than the scheme-SST, with the correlation coefficient increasing from 0.57 to 0.73. To explain this result, the relationships between wintertime NAO and the two predictors in the scheme are analyzed. Figure 6 shows the 15- and 21-yr sliding correlations between the predictors and the winter NAO. The relationship between winter NAO and North Atlantic SST in the preceding months has remained stable over the past several decades, whereas a significant negative correlation between snow cover over Eurasia in the preceding autumn and wintertime NAO has increased since the late 1970s. Similarly to this conclusion, previous studies have found that the significant relationship between the fall Siberian snow and the wintertime AO did not emerge until the 1970s (Peings et al. 2013). However, we admit that there are some problems in the quality of the 20CR snow cover before the satellite era, which may affect the NAO–snow cover relationship. Quasi-Biennial Oscillation in the equatorial stratosphere may have played a role in the modulation of the AO–snow relationship (Saito et al. 2001; Gong et al. 2003; Peings et al. 2013). Because snow cover also has significant impact on winter NAO after the late 1970s, scheme 1 shows a higher level of predictive skill than the existing scheme, which only considers the influence of SST on the NAO, for winter NAO after the late 1970s. To further illustrate the importance of snow cover in the new prediction scheme, we compare the scheme based on the DY of SST with the scheme based

on the DY of snow cover and DY of SST during the period 1981–2000. The relationship between North Atlantic SST, Eurasian snow cover, and winter NAO is statistically significant over this period (see Fig. 6). For the cross-validation tests for the period 1981–2000, the correlation coefficient between the observed NAO and the predicted NAO is 0.63 (exceeding the 99.74% significance level) and the RMSE is 0.83 for the scheme based on the DY of SST (Table 1). When the snow cover is contained in the prediction model, the correlation coefficient is increased to 0.78 (exceeding the 99.99% significance level) and the RMSE is decreased by 23% (Table 1). Thus, the fall snow cover plays a key role in improving the predictive skill of the winter NAO after the late 1970s.

5. Discussion and conclusions

In this paper, a new prediction scheme for winter NAO based on North Atlantic SST and Eurasian snow cover in the preceding autumn has been established using the year-to-year increment approach. Two predictors of the tripole pattern SST and snow cover, which are independent external forcings affecting wintertime NAO, are defined in the prediction model.

Cross-validation tests and independent hindcasts were performed to verify the statistical scheme for forecasting boreal winter NAO. The cross-validation test results for the period 1950–2011 reveal a high correlation coefficient of 0.58 between the predicted and observed DYs of NAO. The correlation coefficient between the predicted and observed DYs of NAO from 1990 to 2011 is 0.74, accounting for 55% of the total variance. We obtain the predicted NAO by adding the predicted DY of NAO to the observed NAO during the previous year. The correlation coefficient between the predicted and observed NAOs during the period 1950–2011 is 0.52, accounting for 27% of the total variance, and with an RMSE of 0.96. The predicted NAO is also closely related to the observed NAO during the period 1990–2011, with a correlation coefficient of 0.55, accounting for 30% of the total variance, and an RMSE of 0.93.

Compared with scheme 0 using the SST and snow cover anomalies in the preceding autumn, our new prediction model shows a higher degree of skill in reproducing the interannual and interdecadal variability of NAO. Therefore, this research provides an effective climate prediction model for the interannual and interdecadal variability of NAO in boreal winter. The year-to-year increment approach can better deal with both the interdecadal component and the interannual component. This is the major reason why the model has higher skill when the increment approach is used than

for the same model that predicts the NAO. We also compare scheme 1 with the scheme-SST, which was established using two lagged modes of North Atlantic SST variability from June to October. Because of the unstable winter NAO–snow cover relationship, scheme 1 for winter NAO has higher predictive skill after the late 1970s than does the scheme-SST. In the future, we will try to further investigate the physical processes of NAO variability and identify new predictors in hopes of improving the forecasting of interannual variability of wintertime NAO. In addition, further research is needed to explore the reasons behind the instability in the relationship between snow cover and winter NAO.

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