Contribution of the Autumn Tibetan Plateau Snow Cover to Seasonal Prediction of North American Winter Temperature

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ABSTRACT

Predicting surface air temperature (T) is a major task of North American (NA) winter seasonal prediction. It has been recognized that variations of the NA winter T's can be associated with El Niño–Southern Oscillation (ENSO) and the North Atlantic Oscillation (NAO). This study presents observed evidence that variability in snow cover over the Tibetan Plateau (TP) and its adjacent areas in prior autumn (September–November) is significantly correlated with the first principal component (PC1) of the NA winter T's, which features a meridional seesaw pattern over the NA continent. The autumn TP snow cover anomaly can persist into the following winter through a positive feedback between snow cover and the atmosphere. A positive TP snow cover anomaly may induce a negative sea level pressure and geopotential height anomaly over the eastern North Pacific, a positive geopotential height anomaly over Canada, and a negative anomaly over the southeastern United States—a structure very similar to the positive phase of the Pacific–North America (PNA) pattern. This pattern usually favors the occurrence of a warm–north, cold–south winter over the NA continent. When a negative snow cover anomaly occurs, the situation tends to be opposite. Since the autumn TP snow cover shows a weak correlation with ENSO, it provides a new predictability source for NA winter T's.

Based on the above results, an empirical model is constructed to predict PC1 using a combination of autumn TP snow cover and other sea surface temperature anomalies related to ENSO and the NAO. Hindcasts and real forecasts are performed for the 1972–2003 and 2004–09 periods, respectively. Both show a promising prediction skill. As far as PC1 is concerned, the empirical model hindcast performs better than the ensemble mean of four dynamical models from the Canadian Meteorological Centre. Particularly, the real forecast of the empirical model exhibits a better performance in predicting the extreme phases of PC1—that is, the extremely warm winter over Canada in 2009/10—should the model include the autumn TP snow cover impacts. Since all these predictors can be readily monitored in real time, this empirical model provides a real-time forecast tool for NA winter climate.

1. Introduction

Large-scale seasonal anomalies of near-surface air temperature (T) in North American (NA) winters have a significant impact on societal and economical activities. In the past several decades, both extreme cold and warm events have become more severe and frequent. For instance, the 2009/10 winter was noteworthy for its anomalously below-normal T's across much of the United States, whereas in Canada, it was the warmest and shortest winter within the past several decades. A useful seasonal forecast of T's is thus of great value.

It is known that the potential seasonal predictability of the atmosphere mostly arises from coupled mechanisms that involve low boundary forcing (e.g., Charney and Shukla 1981), because the atmosphere, on its own, lacks the mechanisms to generate predictable variations beyond synoptic time scales (Lorenz 1963). El Niño–Southern Oscillation (ENSO) is a primary predictability source for interannual variations of NA winter T's (e.g., Ropelewski and Halpert 1986; Hurrell 1996; Shabbar and Khandekar 1996). The influence of ENSO extends to North America through atmospheric teleconnections related to tropical diabatic forcing (e.g., Horel and Wallace 1981). On the other hand, as one of the most important patterns of atmospheric variability, the North Atlantic Oscillation (NAO) accounts for 31% of the hemispheric T's interannual variance over the past 60 winters (Hurrell 1996). The NAO is a large-scale seesaw...
in atmospheric mass between the subtropical high and the polar low (e.g., Wallace and Gutzler 1981; Barnston and Livezey 1987; Hurrell 1995; Li and Wang 2003; Wu et al. 2009). Coherent with the NAO, the North Atlantic tripole sea surface temperature anomaly (SSTA) might be another potential predictability source for NA winter T’s (e.g., Rodwell et al. 1999).

As ENSO and the NAO account for only a portion of climate variability across the globe (Ting et al. 1996; Hurrell 1996; Wu and Li 2008), other lower boundary forcing mechanisms need to be investigated to further improve seasonal predictions (Wu et al. 2011). Of these, one of the most important is likely to be snow cover, as an anomaly in snow cover is associated with changes in solar radiation absorption and surface heat fluxes.

The effect of Eurasian snow cover on global climate has long been noticed (e.g., Gong et al. 2002, 2003, 2004; Fletcher et al. 2009; Sobolowski et al. 2010). A possible relation between the variability of the Indian summer monsoon and Eurasian snow cover was reported (Bamzai and Shukla 1999). The snow cover anomalies over the Tibetan Plateau (TP) were found to have a close connection with Asian climate, especially the East Asian summer climate (e.g., Wu and Kirtman 2007). The winter Arctic Oscillation (AO) was reported to be modulated by the variability of Eurasian snow cover through troposphere–stratosphere interactions (Gong et al. 2007). On the other hand, Clark and Serreze (2000) suggested that the impact of snow cover anomalies over East Asia (105°–150°E) may be limited to short-to-medium-range weather forecasting applications, as they observed a lack of persistence in snow cover anomalies.

The Tibetan Plateau represents the highest elevated land on earth. The snow cover over the TP undergoes a strong annual cycle and interannual variability, which leads to anomalies in atmospheric diabatic heating in the middle troposphere. In this paper, we attempt to investigate whether and how the TP snow cover anomaly influences the NA winter climate, and to what extent it contributes to the seasonal prediction of NA winter T’s.

Section 2 introduces the datasets, model, and methodology used in this study. Section 3 identifies the leading modes of NA winter T’s using an empirical orthogonal function (EOF) analysis. The first principal component (PC1) of the first EOF (EOF1) is found to be significantly correlated with the interannual variability of the autumn TP snow cover anomaly. Section 4 presents the large-scale three-dimensional circulation features associated with the three leading modes. In section 5 the impact of autumn TP snow cover anomaly on the Northern Hemisphere circulation and NA temperature is discussed. In section 6, an empirical model is constructed to predict the PC1 of NA winter T’s based on the autumn TP snow cover and other predictors related to ENSO and the NAO. A seasonal hindcast is performed for the 1972–2003 period, and the result is compared with the multimodel ensemble (MME) forecast of four dynamical models from the Canadian Meteorological Centre. A real forecast experiment is conducted for the 2004–09 period. The last section summarizes our major findings and discusses some outstanding issues.

2. Data, model, and methodology

The main datasets employed in this study include 1) the 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) reanalysis (ERA-40; Uppala et al. 2005) data and the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) Global Reanalysis 1 (NCEP-1; Kalnay et al. 1996) data; 2) the Met Office Hadley Centre’s sea surface temperature (SST) datasets gridded at 1.0° × 1.0° resolution (Rayner et al. 2003); and 3) monthly Northern Hemisphere snow cover data (in unit of percent) gridded at 2.0° × 2.0° resolution, calculated with weekly snow cover data and programs from the National Oceanic and Atmospheric Administration’s (NOAA) Climate Prediction Center (www.cpc.ncep.noaa.gov/data/snow/).

To get a longer time length covering the period from December 1957 through February 2010, the ERA-40 and NCEP-1 data are combined. Since the NCEP-1 data may have systematic errors in the period before 1980 (Wu et al. 2005), we use the ERA-40 data for the period 1957–2001 and extend the data from December 2002 to February 2010 using NCEP-1 data (Wang et al. 2010). To maintain temporal homogeneity, the 2002–10 NCEP-1 data were adjusted by removing the climatological difference between the ERA-40 and NCEP-1 data. In this study, autumn refers to September–November (SON), and winter is defined as December–February (DJF).

The output of the MME hindcast conducted with four dynamical models from the Canadian Meteorological Centre under the second phase of the Historical Forecasting Project (HFP2) is used in this study. The four models are the second- and third-generation atmospheric GCMs (GCM2 and GCM3, respectively) of the Canadian Centre for Climate Modelling and Analysis (CCCma) (Boer et al. 1984; McFarlane et al. 1992), a reduced-resolution version of the global Spectral à Éléments Finis (SEF) model of the Recherche en prévision numérique (RPN) (Ritchie 1991), and the Global Environmental Multiscale-Climate Version (GEM-CLIM) model of the RPN (Côté et al. 1998). The MME hindcast covers 24 winters from 1969/70 through 2002/03. An ensemble of 10 parallel integrations of 4-month duration...
was conducted using each model starting from the beginning of each month. The initial atmospheric conditions were at 12-h intervals preceding the start of the forecasts, taken from the NCEP–NCAR reanalysis. Global SST was predicted using the persistence of the anomaly of the preceding month; that is, the SST anomaly from the previous month was added to the climate of the forecast period. Sea ice (ICE) extents were initialized with the analysis and relaxed to climatology over the first 15 days of integration. The experiment setup of the hindcast was described in Lin et al. (2008) and Jia et al. (2010). In this study the MME hindcast for winter seasons starting from 1 December is used to compare with the statistical model, as introduced in section 6.

To identify the dominant modes, we perform an EOF analysis with winter (DJF) T’s over the NA domain (10°–70°N, 150°–40°W). The EOF analysis is carried out by constructing an area-weighted covariance matrix.

3. Distinct leading modes of NA winter T’s variability

To better understand the relationship between the Tibetan Plateau snow cover and NA winter T’s, first we derive the leading modes of NA winter T’s. Figure 1 presents the spatial patterns and the corresponding PCs of the three leading modes that jointly account for 63% of the total variances of the interannual variability of NA wintertime seasonal mean T’s. The first mode explains around 30% of the total variance and is statistically distinguished from the rest of the eigenvectors in terms of the criterion defined by North et al. (1982).

The most prominent characteristic of the EOF1 mode is a meridional dipole structure with anomaly centers of opposite sign over the northern and southern NA continent (Fig. 1a), at approximately 60° and 30°N, respectively. A positive (negative) PC1 winter is characterized...
by a warm–north, cold–south (cold–north, warm–south) T distribution over the NA continent. PC1 is dominated by interannual variability, as is evidenced in Fig. 1d. Its power spectrum peaks at a periodicity around 3.3 years (Fig. 1g). Embedded in Fig. 1d is the autumn snow cover anomaly averaged over the TP and the adjacent region (25°–50°N, 90°–105°E), which is defined as a TP snow cover index (TPSCI) (red curve). A remarkable feature of PC1 is that it experiences a strongly coherent interannual variability with TPSCI. The correlation coefficient between them reaches 0.57, exceeding the 99% confidence level based on the Student’s t test. It is noteworthy that the extremely anomalous 2009/10 winter had its T’s strongly projected to a positive EOF1, with PC1 reaching the highest value within the past 53 yr (shown by a red circle in Fig. 1d). Meanwhile, the autumn TP snow cover anomaly also reached its highest value since 1972. This, along with the apparent coherence between TPSCI and PC1, indicates that the autumn TP snow cover may be linked to the origin of the EOF1 mode. We will come back to this point in section 5.

The second mode shows a monosign pattern over most of NA, except for the northeastern flank of the continent, with the maximum loading located in the central NA continent and Davis Strait (Fig. 1b). A positive (negative) phase of EOF2 corresponds to a cold (warm) NA continent. The spectrum of PC2 exhibits significant periodicities of 3–4 and 7 yr (Fig. 1h). It is interesting that the EOF2 mode (not shown fully) resembles an NAO-like anomaly pattern over the North Atlantic, namely, a tri-pole T’s anomaly pattern (e.g., Wu et al. 2009). The third mode features a zonal dipole pattern with an opposite sign between the eastern and western NA continent (Fig. 1c). A strong (or weak) third mode corresponds to a cold–west, warm–east (warm–west, cold–east) T distribution. PC3 basically bears three significant spectrum peaks: 2, 3.3, and 25 yr (Fig. 1i).

The distinct spatial–temporal structures of the three leading modes imply they may have different physical origins and predictability sources. In the next section, we will analyze the large-scale circulation anomalies and predictability sources associated with the three leading modes.

4. Dynamic structures and predictability sources of the leading modes

a. Planetary-scale circulation anomalies

Figure 2 shows the anomalous surface circulation regressed to three PCs along with the climatology. For EOF1, there are two negative sea level pressure (SLP) anomaly centers (dashed contours in Fig. 2b). A large area of negative SLP anomalies is located over the northeastern Pacific, with a major trough extending...
southeastward along the west coast of NA. Compared with the climatology (Fig. 2a), this pattern reflects a deeper-than-normal and eastward-shifted Aleutian low pressure system. The northerly surface wind anomalies prevail in the central North Pacific (NP), accompanied by increased advection of cold air from the north, which decreases SST over the central North Pacific. The southerly surface winds along the west coast of the NA continent transport from the south warmer and moister air into this region, leading to increased $T$'s in Alaska and the west coast of NA. The other negative SLP center, which is relatively minor, is located in the midwestern Atlantic, which implies weakening of the subtropical high in this region. Cold anomalies are found in the southeastern United States, extending eastward toward the midwestern Atlantic. The northerly surface winds over the eastern and southeastern United States transport cooler and drier air toward the Gulf of Mexico and the western Atlantic, which decreases SST over that region.

A pronounced feature of EOF2 is the NAO-like structure in SLP and 925-hPa winds over the North Atlantic (Fig. 2c). The correlation coefficient between PC2 and the DJF NAO index is $0.76$, well exceeding the 99% confidence level. A strong positive SLP center associated with anticyclonic surface wind dominates the Greenland and Iceland region, whereas a negative SLP center associated with cyclonic surface wind appears over the midlatitude North Atlantic. This pattern is similar to the negative phase of the NAO, with a weak subtropical high and a weak Icelandic low. A warm anomaly center is located over Greenland. The NA continent is basically controlled by positive SLP and cold $T$ anomalies. The SLP and $T$ anomaly centers tilt southeastward from western Canada to the southeastern United States. The pattern accompanying EOF3 is a wave train with two positive SLP centers over the northeastern Pacific and the eastern NA continent and a negative SLP center over the western NA continent.

Figure 3 compares the anomalous large-scale circulations at midtroposphere that are associated with the three leading modes along with the climatology. For the first mode, there is a negative height anomaly center over the North Pacific, a positive center in central Canada, and a negative center over the southeastern United States and southwestern Atlantic region—a distribution similar to the PNA pattern. The strong low-pressure anomaly center is located over the northeastern Pacific above the cold-air mass, which is consistent with the hypsometric equation (Fig. 3b). Because the climatological ridge over the northeastern Pacific and the west coast of the NA continent tilts southeastward from Alaska to the Rocky Mountains (Fig. 3a), the anomalous high center over central Canada implies an eastward shift of the high ridge toward the central continent, which decreases cold-air activities over
the NA continental region north of 40°N and increases $T$'s in the mid–high latitudes. Another low-pressure anomaly center is seen over the midwestern North Atlantic. Since the climatological NA trough tilts southwestward along the east coast of the NA continent, the anomalous low over the midwestern Atlantic indicates a southward extension of the NA trough, which favors more cold-air activities over the eastern United States and the midwestern North Atlantic. The 500-hPa circulation anomalies have centers that in general match those in the lower troposphere (Fig. 2b), indicating a quasi-barotropic vertical structure of the circulation patterns. The 500-hPa circulation patterns associated with the second and the third modes also resemble those at the surface. For the second mode, pronounced NAO-like anomalies prevail over the North Atlantic sector with a meridional seesaw in geopotential height ($H$) between the high and middle latitudes. For the third mode, a notable wave train propagates from the northeastern Pacific to the eastern NA continent.

Overall, the circulation anomalies associated with the three leading modes may be viewed as departures from the climatological norm. The first mode tends to shift the Aleutian low eastward (Fig. 2). In the midtroposphere, the first mode represents an eastward shift of the high ridge over the western NA continent with a positive anomaly center over central Canada, and a southward extension of the NA trough with negative anomalies over the southeastern United States and the western Atlantic (Fig. 3). EOF1 is closely related to the PNA pattern. The second mode primarily exhibits NAO-like circulation anomalies, whereas the third mode bears a west–east-oriented wave train pattern. Thus, the three leading temperature modes are closely related to large-scale circulation variability.

### b. Predictability sources for the three leading modes

In this paper, we select the autumn ENSO, NAO, and the TP snow cover as three major potential predictability sources for wintertime NA temperature. To examine how the three leading modes of NA winter $T$’s are related to these predictability sources, we calculate the correlation coefficients between the three leading PCs and the prior autumn Niño-3.4 index, NAO index, and the TP snow cover index (Table 1). PC1 has a significant correlation not only with the Niño-3.4 SST but also with the TP snow cover anomalies in SON. Interestingly, the correlation with the TP snow cover is even higher than with the Niño-3.4 index. The second mode also shows a link with the autumn TP snow cover, whereas the third mode does not exhibit any notable connection with the above major potential predictability sources. Thus, the first mode is likely to be the most predictable mode, and this study will focus on seasonal prediction of the first mode.

An important feature that has not been noticed before is that the TP snow cover might provide a new predictability source for seasonal mean NA winter $T$’s besides ENSO. The SON TP snow cover has only a weak correlation with the SON Niño-3.4 SST, with a correlation coefficient of 0.22 for the 1972–2008 period. Clark and Serreze (2000) found that the East Asian snow cover may significantly affect the circulation over the downstream North Pacific on the synoptic time scale. However, according to their lag correlation statistics, the East Asian snow cover anomalies cannot persist beyond about a week. Therefore, they proposed that the East Asian snow cover was limited for short-to-medium-range weather forecasting applications other than for problems on longer time scales. In the present work, we found that the autumn snow cover anomaly over the TP has a good persistence, and that it is closely connected to the first two leading modes of the NA surface air temperature—the first mode in particular. We will discuss how the autumn TP snow cover influences the NA winter $T$’s in the next section.

Although the NAO is correlated with the simultaneous NA winter $T$’s (see section 1; Hurrell 1996), the autumn NAO does not show any considerable impact on the NA winter $T$’s because of a lack of persistence (see Table 1). Thus, the autumn NAO, by itself, may not serve as a good predictor for the NA winter $T$’s. The NAO in fall, however, may extend its influences through some kind of lower boundary forcing, such as the tripole SST anomalies in the North Atlantic (e.g., Wu et al. 2009). Indeed, the autumn tripole SST anomalies in the North Atlantic do show a strong correlation with the first leading mode in the following winter (see section 6). Thus, the autumn tripole SST relative to the NAO is also used as a predictor in this study.

### 5. Impacts of autumn TP snow cover

The first mode of the NA winter $T$’s has a significant positive correlation with the preceding autumn snow cover anomalies over the TP and the adjacent areas (Fig. 4); 12.4% of the entire figure area exceeds the 95% significance level. A Monte Carlo test is performed to test
the field significance of Fig. 4 (Livezey and Chen 1983).

To do so, the PC1 time series is replaced with a series of 38 numbers randomly selected from a normal distribution that has the same autocorrelation as PC1. This experiment is repeated 1000 times. The case of a greater than 12.4% area with correlations statistically significant at the 95% level happens only 5 times. Therefore, it is highly unlikely (at the 99% level) that the results in Fig. 4 were a chance occurrence. An excessive TP snow cover in autumn is likely an antecedent anomalous boundary condition for the occurrence of a high PC1 winter over the NA continent. Likewise, a reduced TP snow cover in autumn tends to lead a low PC1 winter in NA.

To describe the interannual variability of TP snow cover, a TPSCI is defined with the snow cover anomalies averaged over TP and the adjacent region (25°–50°N, 90°–105°E) (red box in Fig. 4). To see whether the snow cover anomaly over this elevated land area has a persistent linkage with the first leading mode, Fig. 5 shows the lead–lag correlations between PC1 and the TPSCI time series from June to August [JJA(0)] to February to April [FMA(1)]. Here “0” denotes the simultaneous year and “1” denotes the following year. JJA(0) leads the simultaneous winter December–February [D(0)JF(1)] by 6 months. The correlation coefficients between TPSCI and PC1 become significant from August to October [ASO(0)] and persist throughout the following winter D(0)JF(1). Thus, the TP snow cover has a strong and persistent connection with the NA T’s leading mode, which makes it of seasonal prediction value.

The autocorrelation of TPSCI between SON and DJF is 0.46, which passes the 99% confidence level. How can the TP snow cover anomalies sustain from autumn through winter? To shed light on its possible mechanism, we take the autumn TPSCI as a reference and compute the lead–lag correlation with 200-hPa geopotential height fields. The results are shown in Fig. 6. The absolute values of the correlation coefficients continuously increase from a 1-month lead through to a 3-month lag. In the 1-month lead (Fig. 6a), before the TP snow cover becomes excessive (reduced), negative (positive) H anomalies emerge over the TP. It indicates that the atmospheric anomalies are responsible for the underneath TP snow cover anomalies, which lead to excessive (reduced) snow cover anomalies. From a 0-month lead through to a 3-month lag (Figs. 6b–e), low-pressure anomalies are enhanced (weakened) over the TP and the adjacent areas, which indicates that the TP snow cover anomalies have a significant positive feedback to the atmosphere. Thus, the persistence of the autumn snow cover anomalies is likely due to a positive feedback between the atmospheric circulation and the snow cover underneath. The possible physical mechanisms can be interpreted as follows. An excessive (reduced) snow cover usually would favor low (high) H anomalies because of reduced (increased) solar radiation absorption at the surface and suppressed (enhanced) sensible heat and radiative fluxes from the ground. On the other hand, the low (high) H anomalies and cold (warm) upper-tropospheric temperature anomalies induced by the snow cover anomaly would produce increased (decreased) upper-tropospheric potential vorticity (Hoskins et al. 1985; Shaman and Tziperman 2005). Anomalous surface convergence (divergence) as well as anomalous vertical motions would follow that would lead to increased

FIG. 4. Correlation coefficient map between PC1 and prior autumn snow cover over the TP and adjacent areas. The interval of contours is 0.1. The shaded regions represent correlation coefficients exceeding the 95% confidence level. The snow cover anomalies averaged in the black box is defined as a Tibetan Plateau snow cover index. The altitudes of areas within bold curves are higher than 1500 m.

FIG. 5. The lead–lag correlation coefficients between PC1 and the TPSCI from JJA(0) to FMA(1). The dotted line represents the 95% confidence level based on a Student’s t test. The vertical line indicates D(0)JF(1) where the simultaneous correlations between PC1 and TPSCI are shown.
(decreased) cloudiness and increased (decreased) precipitation in the region. This would maintain the excessive (reduced) snow cover anomalies. Such positive feedback may sustain such an air–snow cover anomaly pattern. In addition, the atmospheric anomalies tend to propagate eastward from a 0-month lead to a 3-month lag (Figs. 6b–e). In a 3-month lag, a large area of negative correlation is formed in the northeast Asian and northwestern Pacific region near the westerly jet. This pattern is reminiscent of the remote response to an anomalous TP snow cover as revealed in Wang et al. (2008), who completed a numerical experiment with the ECHAM4 model and found that in summer, the TP warming can induce two distinct Rossby wave trains: one in the tropics that propagates along the low-level monsoon westerly and the other in the extratropics along the upper-level westerly jet.

To understand the impact of autumn TP snow cover anomalies on the downstream large-scale circulation and the NA T’s in winter, Fig. 7 presents regressions of DJF atmospheric variables with respect to SON TPSCI. For the 500-hPa geopotential height field (Fig. 7a), negative anomaly centers are found over the eastern North Pacific and the southeastern United States, with positive height anomalies over Canada. This PNA-like pattern is very similar to that associated with PC1 (Fig. 3b), confirming that the autumn TP snow cover anomaly leads to an atmospheric circulation anomaly that is associated with the leading mode of NA T’s. In the North Atlantic, the 500-hPa geopotential height anomaly pattern appears to have a negative NAO structure. It is well known that the tropical diabatic heating anomaly of an ENSO event in the equatorial Pacific can excite an extratropical Rossby wave train that takes the form of the PNA (e.g., Wallace and Gutzler 1981). How a snow cover anomaly around TP also generates a PNA-like circulation anomaly is of great interest. One hypothesis is that the PNA is a preferred mode of atmospheric variability that can be generated by an appropriate external forcing that can be either a tropical diabatic heating anomaly of an ENSO event or an extratropical diabatic heating anomaly of TP snow cover. Another

Fig. 6. The lead–lag correlation patterns between 200-hPa \( H \) (H200) and the autumn TPSCI. The TPSCI leads the H200 by (a) \(-1\), (b) 0, (c) 1, (d) 2, and (e) 3 months. The dark (light) shaded areas denote significant positive (negative) correlation exceeding the 95% confidence level. Note that \(-1\), 0, 1, 2, and 3 months correspond to ASO, SON, October–December (OND), November–January (NDJ), and DJF, respectively.
possible mechanism is the barotropic instability of the two-dimensional winter mean flow as proposed in Simmons et al. (1983), where a disturbance induced by an external forcing can grow by extracting kinetic energy from the mean flow in the eastern North Pacific and North American region. A contribution from the feedback of synoptic-scale transient eddies can also help to maintain the PNA-like circulation anomaly (e.g., Held et al. 1989; Lin and Derome 1997).

The regressions of DJF SLP, surface wind, and $T$'s with respect to autumn TPSCI are shown in Fig. 7b. The SLP anomaly has a similar distribution as a 500-hPa geopotential height, reflecting an equivalent barotropic vertical structure of the PNA-like circulation pattern. Over NA $T$'s display a north–warm, south–cold dipole with two warm centers over Alaska and northeastern Canada. In general, the distribution of surface anomalies is similar to that related to PC1 (Fig. 2b), indicating that the main feature of EOF1 is related to a TP snow cover anomaly. Compared to Fig. 2b, Fig. 7b has a weaker warm anomaly over central Canada. It is possible that over central Canada, the PC1-related $T$ anomaly is also contributed by other processes, for example, an interaction between local snow cover and the atmosphere.

The strong eastern North Pacific negative geopotential anomaly center associated with the autumn TP snow cover (Fig. 7a) will shift the local westerly jet southward, in a way similar to that which accompanies a positive PNA (e.g., Lin and Derome 1997). This southward shift of the westerly jet modulates the DJF synoptic eddy activities over the North Pacific (Fig. 8). Here the synoptic eddy activity is defined as the variance of bandpass-filtered daily mean meridional winds at 300 hPa, calculated for every winter season, where a Lanczos filter with 201 weights is used to retain a period of 2–8 days. Following a positive autumn TP snow cover anomaly, there is a southward shift of synoptic-scale eddy activity over the eastern North Pacific in the following winter.

6. Seasonal prediction of NA winter $T$’s

It has been generally accepted that ENSO is perhaps the most important predictability source for seasonal prediction of the global climate, including NA $T$’s (Ropelewski and Halpert 1986). Besides ENSO, the NAO may be another potential predictability source (e.g., Hurrell 1996; Wu et al. 2009). The result in this study suggests that the autumn TP snow cover could provide an additional physically based predictor for NA winter $T$’s.

Here we focus on the seasonal prediction of PC1, which represents the variability of the dominant mode of NA $T$’s. To assess the contribution of the autumn TP snow cover to the seasonal prediction skill of PC1, an empirical seasonal prediction model is developed using a linear regression method for the period of 1972–2009, which is expressed as

$$y = a_0 + a_1 x_1 + a_2 x_2 + a_3 x_3 + a_4 x_4,$$

where $x_1$ denotes the autumn TPSCI defined above, and $x_2$, $x_3$, and $x_4$ refer to the autumn normalized SSTAs averaged in tropical Indian Ocean (TIO; 20°S–20°N, 50°–85°E), Pacific Ocean ([10°S–10°N, 155°E–80°W] plus [50°–60°N, 170°E–140°W]), and North Atlantic Ocean ([0°–30°N, 70°–20°W] plus [55°–65°N, 65°–20°W] minus [25°–35°N, 95°–70°W]), respectively, that are selected according to correlations between the global autumn SST and PC1 (boxes in Fig. 9). The SSTA in the Pacific can be considered as an ENSO-related predictor, whereas
that in the North Atlantic can be regarded as the NAO-related predictor, namely, a so-called tripole SST anomaly pattern, which was documented in previous studies (e.g., Rodwell et al. 1999; Wu et al. 2009). There is evidence from previous studies that the Indian Ocean SSTA may provide useful information for long-term Northern Hemisphere extratropical circulation variability (e.g., Hoerling et al. 2004). Therefore, although it is not independent of the Niño-3.4 index (their correlation coefficient being 0.54), we consider it not negligible, and it is used as a complement to the other predictors. All these predictors show a significant correlation with PC1 (Fig. 10). The correlation coefficient with the autumn TP snow cover is calculated for the 1972–2009 period because of snow cover data limitation, while the rest are calculated for the 1957–2009 period. If we use the same period of 1972–2009, then the correlation coefficients between PC1 and $x_2$, $x_3$, and $x_4$ are 0.39, 0.36, and 0.56, respectively. Thus, the relative importance of the predictors to the overall model is ordered by the TP snow cover (0.57), the North Atlantic tripole SST anomaly (0.56), the Indian Ocean SSTA, and the Pacific SSTA in terms of the partial correlations in parentheses.

To test the predictive capability of the above empirical model, the cross-validation method is performed to hindcast PC1 for the 1972–2003 period (Wu et al. 2009). To warrant the robustness of the hindcast result, we choose a leaving-eight-out strategy. The procedure is as following: the cross-validation method systematically deletes 8 yr from the period 1972–2003 (namely, years left out in four blocks of 1972–79, 1980–87, 1988–95, and 1996–2003), derives a forecast model from the remaining years, and tests it on the deleted cases. Note that the choice of “leaving eight out” is not random. Randomly choosing 20%–30% of the data to be in the test data and the remainder as the training set can prevent overfitting or data wasting (e.g., Wu et al. 2009). For PC1, 25% of the whole hindcast period (32 yr) equals 8 yr.

The cross-validated estimates of PC1 are shown in Fig. 11. The correlation coefficient between the observations (black line in Fig. 11) and the cross-validated estimate of the empirical model (red line in Fig. 11) reaches 0.55 and the statistical significance test $P$ value equals 0.0012, exceeding the 99.9% confidence level. If we established the prediction model with the TP snow cover excluded and did the same hindcast (green line in Fig. 11), then the correlation coefficient between the observations and the 32-yr cross-validated estimate would drop to 0.43, and the $P$ value increases to 0.0135. This indicates that the autumn TP snow cover does significantly improve the seasonal prediction skill of the empirical model. For comparison, we also presented the observations and the MME hindcast (1969–2002) of the four dynamic models (GEM-CLIM, GCM2, GCM3, and SEF) from the Canadian Meteorological Centre. The MME hindcast of DJF seasonal mean 2-m air temperature initialized from 1 December of each year is projected onto the observed first mode. The hindcast of the new empirical model and that of the MME (blue line in Fig. 11) are both in general agreement with the observations. The correlation coefficient between the observations and the 34-yr MME hindcast is 0.32. The root-mean-square errors for the empirical model and the MME are 0.59 and 0.70, respectively. Therefore, the empirical model has a better prediction skill than the MME as far as PC1 is concerned.

To examine the real forecast skill of the empirical model, we use it to predict PC1 for 2004–09. The prediction equation is established based on training data of
all the years prior to 2004. It is worth mentioning that the extremely abnormal NA 2009/10 winter is realistically forecast (shown by green circle in Fig. 11). Since all these predictors can be readily monitored in real time, this empirical model provides a real-time forecast tool and may operationally facilitate the seasonal prediction of the NA winter climate.

7. Conclusions and discussion

The seasonal prediction of anomalous NA winter $T'$s is of great importance. Many previous studies have shown that NA winter $T'$ variations are closely related to ENSO and the NAO (e.g., Ropelewski and Halpert 1986; Hurrell 1996). This paper, for the first time, presents observed evidence that the autumn TP snow cover anomalies can make a significant impact on the abnormal winter $T'$s over North America. A positive TP snow cover anomaly may induce a pattern similar to the positive phase of the PNA. This usually favors the occurrence of warm–north, cold–south winter over the NA continent. When a negative TP snow cover anomaly occurs, the situation tends to be opposite.

Since the autumn TP snow cover shows only a weak linkage with ENSO and the tropical Indian Ocean SSTA (their correlation coefficient being 0.2), it may serve as a new predictability source for the dominant mode of NA winter $T'$s. Based on the above results, an empirical model is established to predict PC1 of the NA winter $T'$s using a combination of the ENSO-related predictors, the NAO-related predictors, and the TP snow cover anomalies in prior autumn. A hindcast experiment is performed for the 1972–2003 period, which shows a better prediction skill than the MME hindcast of the four dynamical models. Since all these predictors can be readily monitored in real time, this empirical model provides a real-time forecast tool and may facilitate the seasonal prediction of the NA winter climate.

In this study we focused on the seasonal prediction of the leading mode of NA $T'$s (EOF1), which explains about 30% of the seasonal mean variance of surface temperature variability over NA. As the EOF was done over a large area covering the whole continent and some adjacent water, the percentage of variance explained refers to an average of the whole analysis area. For local regions, such as the anomaly centers over Canada and the southeastern United States, the variance explained by this mode would be much larger. The contribution of TP snow cover on NA temperature seasonal prediction is also evidenced in the skill distribution of $T'$s forecast at each grid point. In Fig. 1d, it appears that there is an upward trend in the time series of PC1 and TPSCI. The trend, however, is not significant, as can be seen from the spectrum analysis (Fig. 1g), which shows that PC1 is dominated by interannual variability. To test how much of the skill on the interannual time scale could be contributed by this trend, the hindcast experiment with the empirical model is repeated after removing the linear trend. The skill turns out to be very similar, and there is no change to the conclusion.

Our results indicate that the contribution of the preceding TP snow cover anomalies should be correctly taken into account in a dynamical model. In general, the MME hindcast of four dynamical models used here has a realistic performance in capturing the first mode of NA winter $T'$s (Fig. 12). The meridional seesaw pattern over the NA continent is well reproduced, and the correlation coefficient between observed and hindcast PC1 is 0.41,
exceeding the 95% confidence level. However, our empirical model performs even better than the MME of the dynamical models. One possible reason is that the TP snow cover influence may not have been adequately represented in the models. Our results thus suggest a future direction for the dynamical model improvement.

The empirical seasonal prediction model in this study is based on the seasonal mean time scale, and it is assumed that PC1 is stable. However, many climate extreme events occur on a subseasonal time scale. Does a similar relationship between the TP snow cover and NA T’s exist that can be used to predict the subseasonal events? If not, what are the sources of predictability for the subseasonal extreme events? If the leading mode changes with time, then the predictors and relevant mechanisms may also change, correspondingly. These are still open questions for future investigations. In addition, the hypotheses concerning the origins of the first leading mode call for further numerical and theoretical studies, which will be discussed in the future.

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REFERENCES


