North American Temperature, Snowfall, and Snow-Depth Response to Winter Climate Modes

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ABSTRACT

The snowpack is an important seasonal surface water storage reservoir that affects the availability of water resources during the spring and summer seasons in mid–high latitudes. Not surprisingly, interannual variations in snow cover extent and snow water equivalent have been extensively studied in arid regions such as western North America. This study broadens the focus by examining snow depth as an alternative snowpack metric, and considers its variability over different parts of North America. The authors use singular value decomposition (SVD) in conjunction with linear and partial correlation to show that regional snow-depth variations can be largely explained by the winter North Atlantic Oscillation (NAO) and the Pacific–North American (PNA) modes of atmospheric variability through distinct mechanistic pathways involving regional winter circulation patterns and hydrologic fluxes. The high index phase of the NAO generates positive winter air temperature anomalies over eastern parts of North America, causing thinning of the winter snowpack via snowmelt. Meanwhile, the high index phase of the PNA generates negative winter snowfall anomalies across midlatitudinal areas of North America, which also serve to thin the snowpack. Positive PNA anomalies have also been shown to increase temperatures and decrease snow depths over western North America. The PNA influence extends across the continent, whereas the NAO influence is limited to eastern North America. The winter snow-depth variations associated with all of these pathways exhibit seasonal persistence, which ultimately yield regional-scale spring snow-depth anomalies throughout much of North America.

1. Introduction

Snow cover modulates many components of the energy balance over terrestrial surfaces (Namias 1985; Ellis and Leathers 1999; Grundstein 2003; Zhang 2005). The high solar reflectivity of snow covered versus snow-free surfaces is traditionally considered to be the most influential factor. However, this reflectivity also partially depends on the diffusive properties and hence the mass (i.e., thickness) of the snowpack (Armstrong and Brun 2008). Baker et al. (1991) showed that a discernible layer of snow is required for albedo to reach 0.7; this depth varies from 5 cm for bare soil to 15 cm for alfalfa cover. Other energy fluxes such as sensible heat conduction between the ground, snowpack, and surface air, and latent heat via snowmelt and sublimation, are also influenced by both the presence and mass of the snowpack (McFadden and Ragotzkie 1967; Cohen 1994).

The response of snow to climate drivers has been studied over the past several decades. Although snow extent (SNE) has been investigated most extensively, snowpack mass expressed as snow water equivalent (SWE) has also received considerable attention. These SWE studies have
largely been motivated by the need to assess water resource availability for specific regions such as the mountainous western United States (e.g., Cayan 1996; McCabe and Dettinger 2002; Jin et al. 2006; Mote 2006). Collectively they portray a modest and ambiguous SWE response to large-scale teleconnections such as the Pacific–North American (PNA) pattern, El Niño–Southern Oscillation (ENSO), and the Pacific decadal oscillation (PDO), which suggests that our understanding of snow–climate linkages is incomplete. Studies using North American snow depth (SND, the total depth of old and new snow on the ground) are less common, although the recently released Dyer and Mote (2006, hereafter DM06) SND data provide the opportunity for research at continental scales. For example, DM06 reported significant negative trends in spring SND over vast areas of Canada along with a retreat of the deepest (>40 cm) SND. Ge and Gong (2009) reported a link between spring SND over much of North America and Pacific climate variability as expressed through the PNA and the PDO modes. Their results are consistent with the previous SWE based regional studies.

The PNA pattern is considered a dominant mode of winter atmospheric variability over North America (Wallace and Gutzler 1981; Barnston and Livezey 1987). It is tied strongly to the surface climate variability via associated regional temperature and precipitation anomalies (Leathers et al. 1991). The PNA pattern is often depicted through four concurrent geopotential height anomalies centered over Hawaii, the North Pacific, western Canada, and the southeast United States. The positive phase of this pattern corresponds to anomalously high ridges over Hawaii and western Canada, and anomalously deep troughs over the North Pacific and the southeastern United States, while the negative phase corresponds to lower ridges and shallower troughs over these same regions. The position of the mean stationary waves determines the spatial location of this teleconnection (Leathers et al. 1991). Recently, Ge et al. (2009; hereafter GGF09) identified two distinct physical mechanisms through which the winter PNA influences spring SND, involving the aforementioned regional temperature and precipitation variations. During the positive phase, the enhanced ridge over western Canada interrupts the cold Arctic airflow into this region, which leads to increased air temperature and melting of the snowpack. At the same time, the blockage of westerly flow from the Pacific and southerly flow from the Gulf of Mexico (due to an enhanced southeast U.S. trough) decreases oceanic moisture influx and thus reduces the precipitation across the midlatitude United States. These regional winter patterns result in a diminished winter SND over much of North America; these anomalies then persist into spring.

The North Atlantic Oscillation (NAO) is another major source of interannual and decadal-scale variability in the winter atmospheric circulation over North America (Walker and Bliss 1932; Wallace and Gutzler 1981; Barnston and Livezey 1987; Hurrell 1995, 1996; Wettstein and Mears 2002). Years with a positive NAO index are identified by an enhanced Icelandic low pressure center and Azores high pressure center. Anomalous southerly flow over the eastern United States and anomalous northerly flow over western Greenland and the Canadian Arctic are typically observed during positive NAO years. Correspondingly, higher temperatures occur over the eastern United States and lower temperatures occur over Quebec (Wettstein and Mears 2002). These anomalies can potentially affect the regional snowpack. Prior studies involving snowfall and SWE have reported only modest relationships with the NAO (Hartley and Keables 1998; Morin et al. 2007; Sobolowski and Frei 2007), but the response of SND has yet to be explicitly considered.

The objective of this study is to develop a physically based understanding of the continental-scale SND response to the major winter climate teleconnection modes over North America, specifically the NAO and the PNA. By considering all of North America, we broaden our focus beyond regional water management in the western United States. This is reasonable since climate teleconnections may not be regionally constrained and can span continents. By evaluating SND, we consider an underutilized snowpack metric, facilitated by a recently available gridded North American SND dataset (see section 2). By considering Atlantic as well as Pacific teleconnections, we build upon recent work by GGF09 that has focused primarily on Pacific drivers, thereby providing a more spatially robust explanation of climate-driven North American SND variability. By evaluating climatic fields and seasonal lags, we develop a mechanistic understanding of the physical processes involved.

Figure 1 shows generalized physical pathways that are investigated through which winter climate modes can potentially affect SND. A temperature pathway (denoted by black arrows) considers the effect of winter temperature anomalies on concurrent SND, while a snowfall pathway (denoted by gray arrows) considers the effect of winter snowfall variability on concurrent SND. Winter anomalies are hypothesized to persist into the spring season. Each step in Fig. 1 is systematically investigated using empirical observations and spatiotemporal statistics including singular value decomposition analysis (SVD; see section 3), to discern the responsible climate modes and their regions of influence. For all analyses, mean December–February (DJF) and March–May (MAM) quantities have been used to represent the winter and
We discuss the datasets used in this project in section 2, and the analytical methodology in section 3. Results are presented in section 4, and conclusions are provided in section 5, including a discussion of how the work presented here complements that of GGF09.

2. Datasets

a. Snow depth and snowfall

We use gridded data described in DM06, which was generated from station observations throughout the United States and Canada. This was gridded following the Shepard (1968) interpolation procedure. These datasets offer a fine spatial (1° × 1°) resolution over all of North America, and temporal (daily) resolution for the last 100 years. The most recent version of these datasets, which are available from the Rutgers University global snow laboratory (http://climate.rutgers.edu/snowcover/), extends from 1900 to 2003. This study is limited to the period 1979–2003, however, since for much of the twentieth century the available station data are sparse and unevenly distributed, resulting in erroneous interpolated gridcell values. This time domain also has the benefit of falling within the period of the satellite remote sensing snow cover record, which serves as an independent data source for validating the DM06 dataset (Ge and Gong 2008). More detailed information on this dataset pertaining to quality control, validation, weaknesses, and applicability to continental-scale studies has been provided in DM06, Ge and Gong (2008), and GGF09 and will not be repeated here. Furthermore, Kluver (2007) explicitly examined the reliability of DM06 snowfall dataset. We present results from SVD analyses (section 4), which show significant correlations over regions characterized by high station densities according to Kluver’s analysis.

Figure 2 provides a visual representation of the magnitude and variability of SND at each grid point for both winter and spring. First, an interannual SND time series is computed for each season (not shown) using the area-weighted average SND over North America following Ge and Gong (2008). For each season’s time series, the year with the maximum and the year with the minimum value are identified. The seasonal mean SND at each grid point for these years is shown in Fig. 2. Alaska, the Rocky Mountains, and northern Canada exhibit the deepest layers (>80 cm) of snow.

b. Atmospheric temperature fields

In this study, we use global reanalysis datasets from National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) (Kalnay et al. 1996). We extract monthly mean temperatures at the 925-hPa level (hereafter T925) to represent lower atmosphere temperature. We derive seasonal averages from the monthly fields, for the entire extratropical (20°–90°N) Northern Hemisphere, at a spatial resolution of 2.5° × 2.5°. Temperatures at 850- and 1000-hPa levels were also extracted and yielded qualitatively identical results as for the 925-hPa level.

c. Teleconnection indices

A winter NAO index time series is acquired from the National Center for Atmospheric Research (http://www.cgd.ucar.edu/cas/jhurrell/indices.data.html#naostatmon), computed as the difference in normalized sea level pressures (SLP) between Ponta Delgada, Azores, and Stykkisholmur–Reykjavik, Iceland. Results (section 4) using this station-based NAO index hold true when we repeated the analyses with the NAO index from Climate Prediction Center (http://www.cpc.noaa.gov/products/precip/CWlink/pna/nao.shtml#discussion), as computed through rotated principal component analysis of monthly standardized 500-mb height anomalies. A winter PNA index time series is acquired from the National Oceanic and Atmospheric Administration (NOAA) Climate Prediction Center (http://www.cpc.noaa.gov/products/precip/CWlink/pna/pna.shtml), computed from four regionally averaged H500 values following their modified pointwise method.

3. Methodology

a. Singular value decomposition analysis

SVD is an established method for identifying the temporal covariability between two climatic spatial fields (Prohaska 1976; Lanzante 1984; Wallace et al. 1992; Cherry 1997; Rajagopalan et al. 2000). We employ SVD on various combinations of gridded fields during the
winter (DJF) and spring (MAM) seasons. Here, time series at each grid point have been detrended via ordinary least squares regression to remove the influence of linear trends on the explained covariance. Our SVD analysis utilizes the correlation matrix approach since the coupled fields have different units of measurements. It produces pairs of spatial patterns that explain the maximum mean-squared temporal covariance between the two fields (Bretherton et al. 1992); SVD spatial patterns in each field follow orthogonality constraints. It also generates associated pairs of temporal expansion coefficient (TC) time series that describe the temporal variations in the SVD spatial patterns. We present the spatial patterns as heterogeneous correlation (HC) maps. The \( k \)th HC map shows Pearson correlations between the detrended time series at each grid point for one field and the \( k \)th TC of the other field, thus describing how well the spatial pattern of one field can be explained by the \( k \)th TC of another field. Nonparametric Spearman rank correlations were also examined, and the results are consistent with those using Pearson correlations (not shown).

b. Correlation and partial-correlation analysis

Linear Pearson correlations and partial correlations are also applied directly to various combinations of climatic data and SVD extracted TCs; only results that exceed the 95% confidence level have been reported. Partial correlations identify the remaining linear relationship between two variables after the influence of a third variable has been removed, which is useful for discerning the intervening or antecedent role of a variable in a causal link between two other variables (Blalock 1961). The partial correlation between \( X \) and \( Y \), given \( Z \) as a third variable, is expressed as

\[
r_{XY/Z} = \frac{r_{XY} - r_{XZ}r_{YZ}}{\sqrt{(1 - r^2_{XZ})(1 - r^2_{YZ})}}.\]

This value approaches zero if \( Z \) is responsible for nearly all correlation between \( X \) and \( Y \), and it approaches the basic Pearson correlation value \( r_{XY} \) if the correlation between \( X \) and \( Y \) is unrelated to \( Z \).

4. Results

Two analogous series of analyses are performed in sections 4a and 4b to identify distinct temperature and snowfall pathways, each associated with different climate modes.

a. Temperature pathway

1) Covariability between winter temperature and snow-depth fields reveals the NAO signal

We start with an SVD analysis between the winter T925 and winter SND fields to evaluate the temperature
The HC maps for the leading mode, which explains 31% of the total covariability, are shown in Figs. 3a,b. The winter SND pattern is mainly centered over eastern North America along with a second, less coherent region of covariability over northwestern North America. The winter T925 pattern exhibits regions that vary in phase (out of phase) with winter SND over Greenland, the Bering Sea, and North Africa (east and central North America and northern Eurasia). Over eastern North America, we find the expected local anticontent correlation of temperature and SND.

A similar analysis has been performed between the winter T925 and spring SND. The HC maps for the leading mode, which explains 29% of the total covariability, are shown in Figs. 3c,d. The spring SND covariability pattern (Fig. 3c) shifts northwest to central interior North America following the poleward retreat of the snow line from winter to spring. This central interior SND covariability region is more apparent in spring than winter, since winter T925 anomalies can affect the thermal content of the winter snowpack more than its thickness. Consequently, this may make spring snowpack vulnerable to the positive temperature anomalies and may lead to the faster loss of spring SND. Overall, winter lower atmosphere temperatures exhibit notable structured covariability with both winter and spring SND over North America. Their similarities (Figs. 3b,d) suggest a common climatic driver whose impact on SND is readily visible in winter and spring.

Note that the hemispheric-scale HC1 patterns of T925 (Figs. 3b,d) from both analyses are similar to the spatial pattern of temperature variations associated with the NAO (Hurrell 1996). Furthermore, Tables 1 and 2 indicate that the TC1 time series are highly and consistently correlated with each other and with the winter NAO. Thus an NAO signature appears in the principal covariability between winter T925 and both winter and spring SND fields, whereby the negative NAO anomalies produce negative T925 anomalies and consequently positive SND anomalies over eastern and central parts of North America, and vice versa.

Although the winter NAO appears to be a driving force behind winter T925 and winter–spring SND covariability, it is unlikely to physically extend as far as northwestern...
North America so as to affect the modest SND covariability with T925 during winter for this region indicated in Fig. 3a. Thus, a different climatic driver may also be involved with the covariability between SND and T925. The PNA is a likely candidate, since it is known to exert a strong climatic influence over northwestern North America (NA).

We also present the correlation coefficients associated with second SVD mode in Tables 1 and 2. TC2 time series for winter SND and winter temperature are highly correlated with each other and with the winter PNA (Table 1). Table 2 does not show any TC2 relationship with the winter PNA. However, the TC1 time series for spring SND exhibits a significant relationship with the winter PNA as well as with the winter NAO. Note that, for the first SVD mode between winter temperature and spring SND, the spring SND region of covariability (Fig. 3c) is located northwest of the winter SND region (Fig. 3a), approaching the region of known PNA influence, suggesting that a PNA signal may be present in this first TC.

Thus, in addition to the primary NAO signature revealed in the first mode of our SVD analyses of winter T925 versus winter–spring SND, a secondary PNA signature is also apparent. This result is consistent with GGF09, which derived a component of NA SND variability associated with the PNA mode and identified a physical pathway involving temperatures over northwestern NA. For this study, it appears clearly in the second SVD mode when winter SND is considered (Table 1) but rather vaguely in the first SVD mode with spring SND considered (Table 2). Note that SVD analysis identifies mathematically orthogonal modes of variability, but these modes may not necessarily represent physically distinct phenomena. The same physical relationship may be reflected in more than one mode, and conversely a single mode may reflect more than one physical pathway. Similarities and differences between this study and GGF09 are discussed in section 5.

2) CONFIRMING THE NAO–TEMPERATURE PATHWAY

To confirm that the dominant winter pattern of covariation between the temperature and SND fields is associated with the NAO, we perform additional correlation and partial-correlation analyses. First, gridpoint winter SND is correlated directly against the winter NAO index, as shown in Fig. 4a. Significant negative correlations occur mainly over the eastern United States, largely collocated with the region of highest covariability in the HC map between winter SND and T925 (Fig. 3a), which is also correlated with the NAO (Table 1). Thus the direct linear relationship between winter SND and the NAO is consistent with the temperature pathway discerned from SVD analysis, and the eastern North America center of covariability is reasonable since the NAO has its centers of action in the North Atlantic sector.

The results thus far indicate that the winter NAO mode anomalies result in winter SND anomalies over eastern North America via collocated winter temperature anomalies. To confirm that temperature is a physical mechanism linking the NAO and SND, a partial-correlation analysis is performed to determine the remaining correlation between winter gridpoint SND and the winter NAO after removing the effect of winter temperature. We use the TC1 time series of winter T925 resulting from its SVD analysis with winter SND (denoted as TC1 DJF T925_{SND}) to represent winter temperature, since this component of temperature variability is associated with both SND and the NAO as indicated in Table 1. The partial correlation can be represented symbolically as $r(DJF\ SND, DJF\ NAO|TC1\ DJF\ T925_{SND})$. The result is shown in Fig. 4b, which indicates that, after removing TC1 DJF T925_{SND}, there is essentially no remaining correlation between the winter NAO and winter SND. Thus, the portion of winter T925 variability that relates to winter SND and the winter NAO is responsible for the direct statistical relationship between them. This partial-correlation analysis confirms that, during the winter season, SND over eastern North
b. Snowfall pathway

1) Covariability between winter snowfall and snow-depth fields reveals the PNA signal

An SVD analysis is performed between the winter snowfall (SNF) and winter SND fields to evaluate the snowfall pathway shown in Fig. 1. The HC maps for the leading mode, which explains 27% of the total covariability, are shown in Figs. 5a,b. Both patterns are centered over a broad midlatitudinal band across North America, although the winter SND pattern is more widespread in the west than the winter SNF pattern. The two patterns vary in phase, which is reasonable since snowfall is expected to be directly related to SND.

A similar analysis has been performed between the winter SNF and spring SND. The HC maps for the leading mode, which explains 23% of the total covariability, are shown in Figs. 5c,d. Both patterns are similar to the HC1 patterns of winter SNF and winter SND (Figs. 5a,b) in phase, magnitude, and location. The broad and seasonally consistent spatial patterns of these HC maps suggest that the driver of this observed covariability between SND and SNF occurs on a continental scale, and its influence on SND persists from winter into spring.

Tables 3 and 4 indicate that the TC1s for both fields are significantly correlated with each other and with the winter PNA, but the TC2s are not significantly correlated with the winter PNA. Hence a PNA signature appears prominently in the principal covariability between winter SNF and winter–spring SND fields, whereby the negative PNA anomalies produce positive SNF anomalies and consequently positive SND anomalies across much of North America, and vice versa. Note that the winter PNA appears to influence snowfall variability over the eastern sector of the continent as well as the western sector. This is consistent with previous literature (e.g., Coleman and Rogers 2003; Sheridan 2003), which reported a PNA influence on snowfall over eastern North America via moisture flux from the Gulf of Mexico.

The TC2s in Tables 3 and 4 do not exhibit a clear and consistent relationship with the winter PNA mode, which suggests that the PNA signature is effectively captured by the first mode of covariability between SND and SNF. Likewise, a relationship with the winter NAO mode is not apparent in either SVD mode. Isolated statistically significant correlations to winter NAO appear in SVD2 (Table 3) and SVD1 (Table 4), but they are not coherent and likely arise as artifacts of the SVD analysis. SVD identifies mathematically distinct modes of covariability that are not necessarily physically distinct, while the general regions of known PNA and NAO influence overlap over eastern North America.

2) Confirming the PNA-snowfall pathway

Analogous to our approach for confirming the NAO-temperature pathway in section 4a, here we also correlate gridpoint winter SND directly against the winter PNA index (Fig. 6a). Strong negative correlations occur over western North America, which is reasonable since the PNA pattern is primarily a Pacific sector phenomenon. This region also partially overlaps with the covariability regions in the HC map between winter SND and SNF (Fig. 5a) that are significantly correlated with the winter PNA (Table 3). Figures 5a and 6a are not expected to look exactly alike, since factors other than PNA can contribute to the winter SND and SNF covariability shown in Fig. 5a, and PNA–SND relationship shown in Fig. 6a can occur via mechanisms other than
snowfall (e.g., as shown in GGF09). The relationship between PNA, SNF, and SND is clearly not a simple one. Nevertheless, the direct statistical relationship between PNA and SND in Fig. 6a provides additional circumstantial evidence that the broad covariability signal between SNF and SND shown in Fig. 5a is associated in part with the PNA pattern.

The results thus far support the snowfall pathway illustrated in Fig. 1, more specifically that the winter PNA mode anomalies result in winter SND anomalies over portions of North America, via collocated winter snowfall anomalies. To confirm that snowfall is a physical mechanism linking the PNA and SND, a partial-correlation analysis is performed to determine the remaining correlation between winter gridpoint SND and the winter PNA after removing the effect of winter snowfall. We use the TC1 time series of winter SNF resulting from its SVD analysis with winter SND (denoted as TC1 DJF SNFDJF SND) to represent winter snowfall, since this component of snowfall variability is associated with both SND and the PNA as indicated in Table 3. The partial correlation can be represented symbolically as

\[
\begin{align*}
\text{Corresponding SVD field} & \quad 0.96 & \quad 0.96 \\
\text{DJF PNA} & \quad -0.58 & -0.56 & 0.32 & 0.16 \\
\text{DJF NAO} & \quad -0.17 & -0.22 & 0.4 & 0.39 \\
\end{align*}
\]

Fig. 5. Heterogeneous correlation maps of the first SVD mode between winter snow depth vs winter snowfall: (a) SVD1 for winter snow depth; (b) SVD1 for winter snowfall. Same between spring snow depth vs winter snowfall: (c) SVD1 for spring snow depth; (d) SVD1 for winter snowfall. Correlation coefficients (r) significant at the $\approx 95\%$ confidence level are shown.

\[
\begin{align*}
\text{Corresponding SVD field} & \quad 0.93 & \quad 0.94 \\
\text{DJF PNA} & \quad -0.69 & -0.63 & 0.08 & 0.13 \\
\text{DJF NAO} & \quad -0.28 & -0.41 & -0.25 & -0.15 \\
\end{align*}
\]
The result is shown in Fig. 6b, which indicates that after removing TC1 DJF SNF DJF SND, the only remaining significant and coherent correlation is found over the coastal mountain ranges of British Columbia and southern Alaska. Thus, the portion of winter SNF variability that relates to winter SND and the winter PNA is responsible for most of the direct statistical relationship between them. This partial-correlation analysis confirms that during the winter season, SND over much of North America responds to the PNA mode via variations in snowfall.

Note that this PNA–snowfall pathway occurs across much of North America according to the SVD analysis shown in Fig. 5, with an emphasis on western North America according to the linear and partial-correlation analyses shown in Fig. 6. In contrast, the NAO-temperature pathway occurs over eastern North America, according to Figs. 3 and 4. Hence, there appear to be distinct physical mechanisms throughout North America over geographical regions that are consistent with the domain of their respective climatic drivers.

c. Confirming the independence of two pathways

To demonstrate the distinction between the two pathways identified, a final set of partial correlations is performed. Figure 7a shows $r$(DJF SND, DJF PNA(TC1 DJF SNF DJF SND)), which represents the remaining correlation between winter gridpoint SND and winter SNF after removing the effect of winter T925. Once again, we use component of T925 variability that covaries with SND as the intervening variable. Figure 7a exhibits the same pattern of covariability as Fig. 5a, which indicates that T925 has no bearing on SVD1 covariability between SND and SNF. In other words, the PNA-snowfall pathway is maintained after removing the contribution of the NAO–temperature pathway.

We use the same technique to remove the effect of winter SNF instead of winter T925. Figure 7b shows $r$(DJF SND, TC1 DJF T925 SNP DJF SND/TC1 DJF SNF DJF SND), which represents the remaining correlation between winter gridpoint SND and winter T925 after removing the effect of winter SNF. Once again, we use component of SNF variability that covaries with SND as the intervening variable. Figure 7b exhibits the same pattern of covariability as Fig. 3a, which indicates that SNF has no bearing on SVD1 covariability between SND and T925. In other words, the NAO-temperature pathway is maintained after removing the contribution of the PNA–snowfall pathway.

Note also that, when the PNA-snowfall pathway is removed in Fig. 7b, a separate region of covariability between SND and T925 begins to emerge prominently over northwestern North America. This is very similar to Fig. 3a, though Fig. 7b exhibits a much stronger and more coherent signal. GGF09 revealed the PNA-driven SND anomalies over North America that involved both precipitation and temperature. Their PNA-precipitation pathway spanned midlatitude North America and is entirely consistent with the PNA-snowfall pathway identified here. Their PNA-temperature pathway was centered over northwest of North America and is consistent with the secondary covariability region between SND and air temperature revealed in Fig. 3a and especially in Fig. 7b. Hence, our interpretation of Figs. 3a and 7b is that, when the winter snowfall is removed as an intervening variable, the SVD covariability between winter SND and air temperature is refined and reveals two mechanistic pathways: one involving the NAO over eastern North America and another involving the PNA over northwestern North America.
d. Persistence of winter snow-depth anomalies into spring reveals signatures of atmospheric modes

Finally, an SVD analysis between the winter and spring SND fields is performed to identify the dominant spatiotemporal patterns of its seasonal persistence. This relationship represents the common final step in both pathways illustrated in Fig. 1. Spatial patterns for the first mode (HC1; Figs. 8a,b) explain 29% of the total covariability between these two fields. Broad regions with significant covariability upward of $r = 0.9$ are centered over northwest North America. As indicated in Table 5, the corresponding TC1 time series are highly correlated with each other. The spatial location and the temporal variability of the first mode suggest a possible signature of the PNA on the covariability between winter SND and spring SND. However, Table 5 shows that linear correlations between winter PNA and the SND TC1s are not significant at the 95% confidence level, although they are significant at the 90% confidence level.

The spatial patterns for second SVD mode (HC2; Figs. 8c,d) are shifted slightly to the southeast, across southern Canada and the northern United States, with broad regions of significant covariability upward of $r = 0.88$ that explain 17% of the total covariability. Table 5 indicates that the corresponding TC2 time series are highly correlated with each other, significantly correlated with the winter PNA, the winter NAO, and the winter NAO. The signature of PNA on the covariability between winter SND and spring SND is more apparent in this second SVD component of SND persistence than the first mode, and an NAO signature has also emerged, with negative correlations indicating inverse relationships. Thus the first two modes of seasonal SND persistence over North America span most of the continent and appear to be statistically related to major winter climate teleconnections; although the first mode is correlated with PNA at only 90% confidence level. Note that the PNA and NAO signals emerge from this analysis, which includes only SND observations, with no input from any other fields.

5. Discussion and conclusions

In this study, two prominent modes of atmospheric circulation, that is, the NAO and the PNA, have been identified as major drivers of SND variability over North America. Lower atmospheric temperature and snowfall are used in conjunction with SND to identify spatially coherent patterns of covariability within North America, which portray physically based linkages between the climate modes and North American SND. These atmospheric linkages result in winter SND anomalies over North America that subsequently persist into the spring season. Furthermore, the SVD analysis of winter versus spring SND reveals the signature of the NAO and PNA teleconnections on SND observations: no corollary field (e.g., atmospheric circulation), and no a priori assumptions, are required to identify the PNA and NAO signatures in the SND observations.

Figure 9 shows a refined version of the generalized pathways in Fig. 1 that incorporates the results of our investigation. The snowfall pathway (denoted by gray arrows) is attributed to the winter PNA mode, where the positive phase of the winter PNA is associated with below-normal snowfall across midlatitude North America and hence shallower winter–spring SND. Results of this study attribute the temperature pathway (denoted by black arrows) to the winter NAO mode, where the positive phase of the winter NAO is associated with above-normal temperatures over eastern North America and hence a thinner
winter–spring snowpack. A separate temperature pathway attributed to the winter PNA mode (denoted by dashed arrows) was shown previously by GGF09 and is also supported by results presented here, where the positive phase of the winter PNA is associated with above-normal temperatures over northwestern North America and hence a thinner winter–spring snowpack. Taken together, these three pathways explain much of the SND variability throughout North America.

Note that SVD analysis simply identifies coherent covariability between two specified variables and does not necessarily identify physically distinct modes of variability over multiple variables. Hence the SVD spatial pattern resulting from one mode may arise from partially overlapping phenomena, and a phenomenon may be evident in more than one SVD pattern or in different SVD modes across different variables. This helps to explain why, for example, the NAO–temperature pathway is clearly evident in the first mode of winter SND versus T925 covariability but obscured within the second mode of winter versus spring SND covariability. This may also explain why the second covarying pattern is not. Nevertheless, the clear, coherent and consistent patterns that result from the SVD analyses presented here provide compelling evidence for the aforementioned pathways.

Since the results presented here are consistent with the results of GGF09, it is informative to discuss the similarities and differences between these two studies. Both apply the DM06 dataset over the full North American spatial domain, use the same PNA index time series, and consider the winter–spring season. However, GGF09 used a one-parameter EOF filtering procedure on SND in order to extract its specific relationship to the PNA and the associated atmospheric pathways. In contrast, here...
we use two-parameter SVD analyses to identify the specific patterns of covariability between various combinations of variables along the pathways, and statistically link them with climate teleconnections without a priori identifying the PNA or any particular signal as a mechanism. Hence our study complements GGF09, by broadening the suite of climate teleconnections which influence winter–spring SND variability over North America. In particular, the PNA-snowfall pathway across midlatitude North America identified here echoes the PNA-precipitation pathway in GGF09. Moreover, a physical linkage with the winter NAO is revealed here that contributes to the second-most dominant mode of seasonal SND persistence and is manifest as an NAO-temperature pathway over eastern North America.

Recent studies (e.g., Feldstein 2000; Rivière and Orlanski 2007) show that the NAO and PNA have time scales of about 10 days and intraseasonal variability cannot be captured through the mean seasonal (winter or spring) index, as applied in this paper. Thus, there is a future scope of research to investigate how the variations in the NAO and PNA affect SND variability at shorter time scale as opposed to interannual time scale.

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