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Comments

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On the Relationship between Spring NAO and Snowmelt in the Upper Southwestern United States

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ABSTRACT

This study examines the relationship between the North Atlantic Oscillation (NAO) and snowmelt in spring in the upper southwestern states of the United States (UP_SW) including California, Nevada, Utah, and Colorado, using SNOTEL datasets for 34 yr (1980–2014). Statistically significant negative correlations are found between NAO averages in the snowmelt period and timings of snowmelt (i.e., positive NAO phases in spring enhance snowmelt, and vice versa). It is also found that correlations between El Niño–Southern Oscillation and snowmelt are negligible in the region. The NAO–snowmelt relationship is most pronounced below the 2800-m level; above this level, the relationship becomes weaker. The underlying mechanism for this link is that a positioning of upper-tropospheric anticyclonic (cyclonic) circulations over the western United States that are associated with development of the positive (negative) NAO phases tends to bring warmer and drier (colder and wetter) spring weather conditions to the region. The temperature variations related with the NAO phases also strongly modulate the snowfall–rainfall partitioning. The relationship between the NAO and spring snowmelt can serve as key information for the warm season water resources management in the UP_SW.

1. Introduction

Winter snowpack is one of the most important components of water storage and supply in the western United States (WUS) (Serreze et al. 1999; Mote et al. 2005). Therefore, a pronounced trend or variability of the snowpack amount (e.g., decreasing trends in winter snowpack) and snowmelt is of great concern, especially in the upper southwestern United States (UP_SW), which includes California, Nevada, Utah, and Colorado. A number of previous studies have documented decreasing trends in the snow water equivalent (SWE) and the snow cover extent (SCE) accompanied by earlier snowmelt in the spring, causing earlier peaks in snowmelt-driven streamflow (e.g., Cayan et al. 2001; Stewart et al. 2005; Tootle et al. 2005; Clow 2010). These observed trends seem to be attributed to the increasing trend in spring temperatures (Hamlet et al. 2005), despite winter precipitation increases in some regions (Stewart 2009).

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The amount of snowpack (e.g., SWE and SCE) is highly susceptible to cold season weather conditions such as temperature and precipitation. Since the SWE value on 1 April tends to "represent a cumulative, simplified summary of the previous several months' weather" (Mote 2006, p. 6210), it has been widely used in the previous snow-related studies for the WUS region. Several studies found that 1 April SWE is closely related to precipitation and temperature anomalies in the antecedent winter months as well as in early spring (Cayan 1996; Mote 2006; Clow 2010; Pederson et al. 2011). In addition, via influences on precipitation and temperature anomalies, large-scale climate variability such as El Niño–Southern Oscillation (ENSO) and the Pacific decadal oscillation (PDO) tends to affect the spatial and temporal variability of the 1 April SWE in the WUS (Ropelewski and Halpert 1987; Gershunov and Barnett 1998; McCabe and Dettinger 1999; Myoung and Deng 2009; Myoung et al. 2015). ENSO typically exerts an opposite effect on precipitation between the lower southwestern United States (Southern California, Arizona, New Mexico, and western Texas) and the Pacific Northwest; with a dipole pattern, ENSO and precipitation anomalies have positive correlations in the lower southwestern United States and negative correlations in the Pacific Northwest. Similar results are also obtained in the relationship between PDO and precipitation anomalies, due to the high degree of serial correlation between ENSO and PDO ($r = 0.49$ on monthly time scales for 1950–2015). Therefore, the 1 April SWE variability also has similar spatial patterns in the relationship with ENSO and with the PDO (Cayan 1996; McCabe and Clark 2005; Pederson et al. 2011). However, these impacts of ENSO and the PDO are negligible in the transitional areas between the two regions, such as the UP_SW region (Dettinger and Cayan 2014; Myoung et al. 2015). Other climate variability indices such as the Atlantic multidecadal oscillation (AMO), North Pacific Index (NPI), and North Atlantic Oscillation (NAO), have also been examined, mostly on decadal time scales (Mote 2006; Hunter et al. 2006). However, none of the above showed a single critical impact on SWE for the entire UP_SW region.

Having close resemblance in characteristics to the Arctic Oscillation (Thompson and Wallace 1998; Thompson et al. 2000; Thompson and Wallace 2001) and northern annular mode (McAfee and Russell 2008), the NAO has been known to greatly affect the variability of temperatures and precipitation in western Europe and eastern United States (Hurrell and van Loon 1997; Higgins et al. 2000; Wettstein and Mearns 2002; Brown et al. 2008). While the connection between the NAO and weather in the WUS has received less attention, the NAO has been recently found to have a significant relationship with the temperature in the UP_SW region in spring and early summer (McAfee and Russell 2008; Myoung et al. 2015), which turns out to have been strengthened in recent decades, 1980–2009, compared to the earlier decades, 1950–79 (Myoung et al. 2015). This suggests that snowmelt can also be similarly related to NAO variability. Usually, SWE reaches its maximum around 1 April or later in the UP_SW region, whereas 1 April SWE is more closely associated with environmental conditions relevant for snow accumulation than for snowmelt. Therefore, a careful examination is required for the variability of SWE in melting periods and possible impacts of the NAO on snowmelt. This study examines multidecadal climatology and interannual variability of SWE characteristics in the UP_SW (e.g., accumulation date–period, max SWE date–amount, melting date–period, and length of melt period), emphasizing snowmelt and its relations to the NAO. Physical processes explaining the NAO and SWE relations are also examined and described in terms of meteorology.

2. Data and methods

The daily NAO index defined by Barnston and Livezey (1987) is obtained from the National Oceanic and Atmospheric Administration Climate Prediction Center archives (http://www.cpc.ncep.noaa.gov/products/precip/CWlink/pna/nao.shtml). We use 11-day running averaged NAO index for comparability to 11-day running averaged SWE (see below). The weekly ENSO (Niño-3.4) index data (http://www.cpc.ncep.noaa.gov/data/indices/wksst8110.for) are analyzed due to the fact that a daily ENSO index is unavailable. The weekly ENSO index is available from 1990. For atmospheric geopotential heights (HGT), we use the monthly-mean values in the National Centers for Environmental Prediction (NCEP)–National Center for Atmospheric Research (NCAR) reanalyses of $2.5° \times 2.5°$ resolution (http://www.esrl.noaa.gov/psd/data/gridded/data.ncep.reanalysis.html) (Kalnay et al. 1996).

The automated SNOTEL system provides the SWE and meteorological data in the regions of high snow accumulation throughout the WUS to supplement and/or replace snow course data (Serreze et al. 1999). Precipitation (PRCP) measurements at the same stations started in the early 1980s, and measurement of daily surface air temperatures such as average, maximum, and minimum temperature (Tave, Tmax, and Tmin, respectively) started in the late 1980s. We have examined data from 151 SNOTEL stations: 21 in California (CA), 22 in Nevada (NV), 60 in Utah (UT), and 48 in Colorado (CO) that are located in the altitude range between 1500 and 3700 m. Station elevations in CO and UT are generally higher than those in CA and NV. Arizona and New Mexico are excluded in this study because of inconsistent SWE values throughout winter seasons that are caused by the less frequent snowfall of smaller amounts compared to the stations in CA, NV, UT, and CO.

Our analysis period covers 34 yr, from 1980/81 to 2013/14 when the number of SNOTEL stations drastically increased (Serreze et al. 1999) and for which corresponding precipitation observations exist. This period also coincides with the period of enhanced impacts of the NAO on temperatures (Myoung et al. 2015).

As snow accumulation–melting characteristics vary by geography (e.g., elevation, direction, and proximity to the ocean), surface conditions (e.g., topography, vegetation types, and soil types), and seasons, we separate the melting period from the accumulation period. This makes this study unique from other studies that focus on the 1 April SWE (e.g., Cayan 1996; Mote 2006; Pederson et al. 2011). The separation is done by applying 11-day running averages on SWE and then identifying the accumulation start date (Accumul start date), maximum SWE date (Max SWE date), and complete melting date (Zero SWE date) at each station and for each year as illustrated in Fig. 1a. The 11-day running average successfully captures the general trend of SWE increases in winter and SWE decrease in spring, separated by Max SWE dates, not only at high- and low-elevation stations but also for dry and wet years (see Fig. S1 in the supplemental material). We also identify the dates of 50% SWE accumulation (+50% SWE date), 75% SWE accumulation (+75% SWE date), and 50% SWE melt (−50% SWE date). Based on these dates, the melting period is defined as the period from Max SWE date to Zero SWE date. Similarly, the first accumulation period is defined as the period from the accumulation start date to +50% SWE date, and the second accumulation period from +50% SWE date to Max SWE date.

The first and second accumulation periods as well as the melting period vary from year to year on a daily basis at each station (interannually adjusted). Thus, the mean NAO in the melting period refers to the mean value of the 11-day running averaged NAO indices during the interannually adjusted melting period, and the mean NAO indices in the second accumulation period refers to those during the interannually adjusted second accumulation period. For ENSO, since the daily Niño-3.4 index data are unavailable, we select specific weeks that correspond to the interannually adjusted melting periods when calculating ENSO averages in the melting periods. This causes differences of up to 6 days with respect to the daily-based melting periods (mostly from 1.5 to 2 months long), which results in a ~10%–13% variance of the accurate melting periods.

In addition to the interannually adjusted periods, we also calculate fixed periods based on the 34-yr median of the +50% SWE date, +75% SWE, Max SWE date, −50% SWE, and Zero SWE date (fixed, hereafter) at each station. Thus, the mean NAO in the fixed melting period refers to the mean value of the 11-day running averaged NAO indices during the fixed melting period. The same approach is adopted for the mean NAO in the fixed second accumulation period. Meanwhile, in section 3, correlation results based on the interannually adjusted period are primarily shown, if not specified as the fixed period.

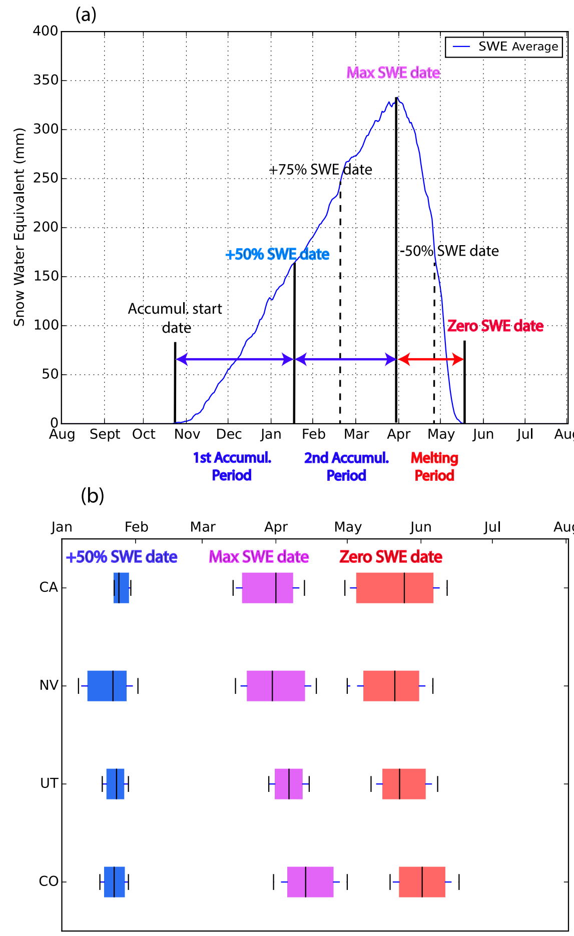

FIG. 1. (a) The 34-yr median daily SWE (11-day running mean) over the stations in California and definitions of various SWE variables used in this study. (b) The 34-yr median and variations of the +50% SWE date (blue), Max SWE date (purple), and Zero SWE date (red) in each state: California, Nevada, Utah, and Colorado.

To examine the impacts of the NAO on the variations in SWE, HGT, surface temperature, precipitation, and snowfall ratio, we use a simple linear correlation analysis. Here, the snowfall ratio is defined as the snowfall amount divided by the total precipitation amount. SNOTEL observations do not distinguish snowfall from rainfall; thus, precipitation is assumed to be in the form of snow when a daily temperature (e.g., Tave) is below a threshold value (e.g., −3°, 0°, 3°, 6°, or 9°C). In this analysis, Tmin and Tmax as well as Tave are also tested. Several previous studies have examined the snow and rainfall ratio across rain–snow transition regions using daily temperature at surface (Auer 1974; Kienzle 2008; Dai 2008). The daily temperature has been generally used to determine the precipitation type at the surface due to the lack of an atmospheric boundary layer dataset that could provide accurate precipitation phase.

The resulting Pearson correlation coefficients indicate the strength of a linear relationship between two fields. Assuming independent, normally distributed data, ± 0.34 is the 95% confidence level for a nonzero correlation for the 34-yr samples with $N^{-2} = 32$ degrees of freedom. Note that the period of the correlation analysis using temperature variables is 24 yr, 1990–2014, due to lack of reliable temperature data before 1990 at the SNOTEL stations. In this case, ± 0.40 is the 95% confidence level. While the NAO and ENSO indices do not show noticeable long-term trends in spring, some variables such as temperature and snowmelt date show weak trends at some stations. To examine the significance of these trends, correlation coefficients are calculated both before and after removing the long-term linear trends from all variables. Differences in correlation coefficients between these two calculations, less than 0.1, are not significant; we only present the correlations calculated with raw data (i.e., the long-term trend is not removed) in section 3.

3. Results

a. Springtime SWE characteristics and their links to the NAO

The 34-yr median and variation of the SWE dates for each state are displayed in Fig. 1b. The median and variation are determined by, first, calculating the annual average SWE dates for each given state and, second, computing median and variation of the annual-mean SWE dates. While the +50% SWE dates occur in January in all of the states, Max SWE dates and Zero SWE dates vary by states, with the earliest in CA and the latest in CO. Snowmelt usually occurs from mid-March to early June. The melting period (~48 days) is generally shorter than the second accumulation period (~71 days). The interannual variations of Max SWE date and Zero SWE date are larger than those of the +50% SWE date.

Figure 2a shows the correlations between the mean NAO in the second accumulation period and the +75% SWE date. The correlations between the mean NAO in the melting period and the −50% SWE date (Fig. 2b), and then with the Zero SWE date (Fig. 2c), are also shown. Regarding Fig. 2a, although negative correlations are pervasive, they are not statistically significant at the 95% confidence level except for a few stations in NV, UT, and CO. This suggests that the +75% SWE dates are not well correlated with the NAO during the second accumulation period. However, for the melting period in Fig. 2b, not only strong but also statistically significant negative correlations prevail over most of the study region, indicating that the higher the NAO index value, the earlier the melting date. The correlation of the mean NAO in the melting periods with Zero SWE date (Fig. 2c) shows a similar pattern but even stronger relationships although the changes in magnitudes are not obviously recognizable in the current format of the figure. Thus, the spring snowmelt in the UP_SW region is strongly negatively correlated with the NAO (i.e., earlier snowmelt in the years of strongly positive NAO in spring). These results are consistently found even if we use the fixed second accumulation periods and the fixed melting periods (not shown).

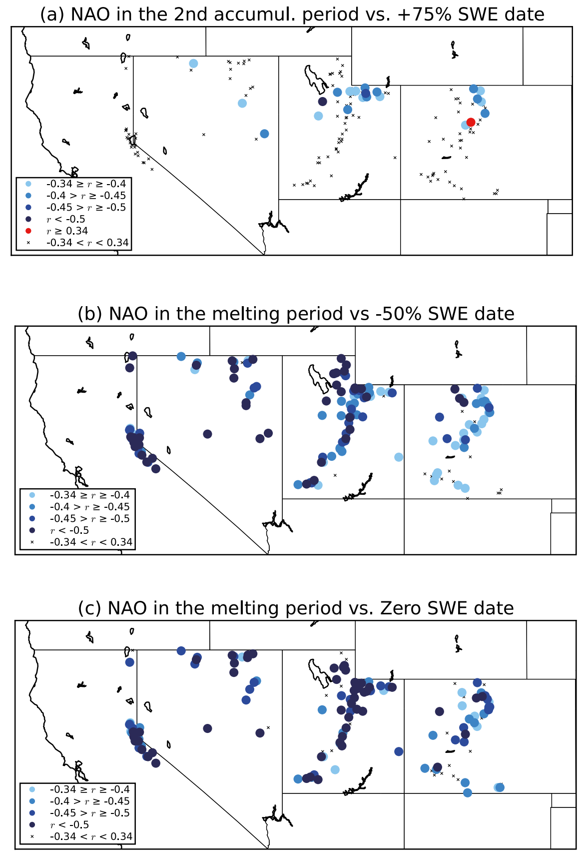

FIG. 2. (a) Correlation map between the NAO average in the second accumulation period and the +75% SWE date for 34 yr (1981–2014). (b),(c) Correlation maps between the NAO average in the melting period and the −50% SWE date and between the NAO average in the melting period and Zero SWE date, respectively. Statistically significant (insignificant) correlations at the 95% confidence level are indicated by colored circles (small crosses). Negative (positive) correlations are shown by blue (red) color shades.

b. *Underlying mechanisms of the impacts of the NAO on snow melting*

The physical reasons for the link between the NAO and the snowmelt dates seem to be due to the impacts of the NAO on surface weather conditions in spring, mainly temperature and precipitation. For example, the strong positive impact of the NAO on spring temperatures has been documented previously (e.g., McAfee and Russell 2008; Myoung et al. 2015), and the impact of the NAO has been strengthened in the most recent 30 years compared to that in the previous 30 yr (Myoung et al. 2015). According to Myoung et al. (2015), an upper-level anticyclonic circulation tends to be developed over the WUS during a positive NAO phase, which is similar to an atmospheric blocking, a prominent feature in the Northern Hemisphere in recent decades (Carrera et al. 2004; Häkkinen et al. 2011; Francis and Vavrus 2012). This anticyclonic circulation is likely to increase the surface temperature in UP_SW by reducing cloud cover and by changing wind direction (warm temperature advection). The increase in surface air temperatures in positive NAO springs may induce 1) accelerated snowmelt and/or 2) reduced snowfall ratio; both are related with higher spring temperature and can result in earlier snowmelt–Zero SWE dates. In addition, the NAO can affect snowmelt via precipitation amounts in melting periods because the upper-level anticyclonic circulation in the positive NAO phase tends to suppress surface disturbances, and thus precipitation (and vice versa). This is important as well as the impacts of temperature because precipitation during the melting period, particularly from snowfall, can increase SWE and then delay snowmelt.

To test the above hypotheses, we compute the correlations between the NAO and Tave and between the NAO and PRCP (Fig. 3) during the melting period. The prevalent positive NAO–Tave correlations (Fig. 3a) and negative NAO–PRCP correlations (Fig. 3b) indicate that a positive NAO phase tends to increase temperatures and reduce precipitation. The NAO–Tave correlations are strong and statistically significant at most of the stations. Surprisingly, precipitation is also significantly correlated with the NAO at most stations in CA and NV as well as at a number of stations in UT and CO. These results suggest that the links among the NAO, SWE, and snowmelt described above occur via the impacts of the NAO not only on surface temperatures but also on precipitation. The fact that the correlations between the NAO and the Zero SWE date (Fig. 2c) are higher than those between the NAO and temperature (Fig. 3a) or between NAO and precipitation (Fig. 3b) also suggests that not only the NAO–temperature link

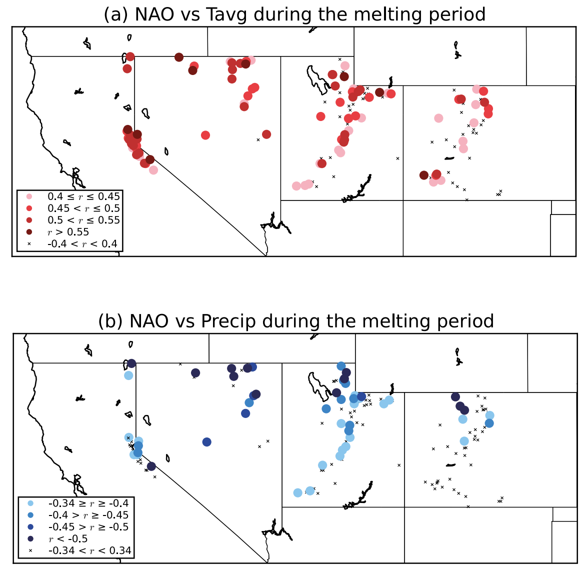

FIG. 3. Correlation map (a) between the NAO and daily surface air temperature for 24 yr (1991–2014) and (b) between the NAO and precipitation for 34 yr (1981–2014) in the melting period. Statistically significant (insignificant) correlations at the 95% confidence level are indicated by colored circles (small crosses). Negative (positive) correlations are shown by the blue (red) color shades.

but also the NAO–precipitation link may jointly strengthen the NAO–snowmelt link. Specifically, for UT and CO, the correlations between the NAO and the Zero SWE date are statistically significant at most of the stations despite weak relationships between the NAO and Tave and/or between the NAO and PRCP. We also estimate correlations after removing the long-term trend from each variable to find only negligible changes (i.e., slightly higher correlations).

To elucidate the large-scale circulation associated with the snowmelt variation and with the NAO, we compute the 34-yr correlation between the domain-averaged Zero SWE date and April–May mean HGT at 500 hPa (Fig. 4a) and between the April–May mean NAO and April–May mean HGT at 500 hPa (Fig. 4b). The months of April and May are chosen because the fixed melting periods generally fall between April and June in the study domain (Fig. 1b). For direct comparison with Fig. 4b, the Zero SWE date and HGT correlations in Fig. 4a are multiplied by −1.0, so that positive correlations indicate that positive height anomalies are associated with early snowmelt. It is notable that similar results are also consistently found when, instead of the domain-averaged Zero SWE date, the state-averaged Zero SWE dates are used (not shown). This feature is primarily due to the fact that the state-averaged Zero SWE dates are strongly correlated with each other among the four states (r ranging from 0.61 to 0.92).

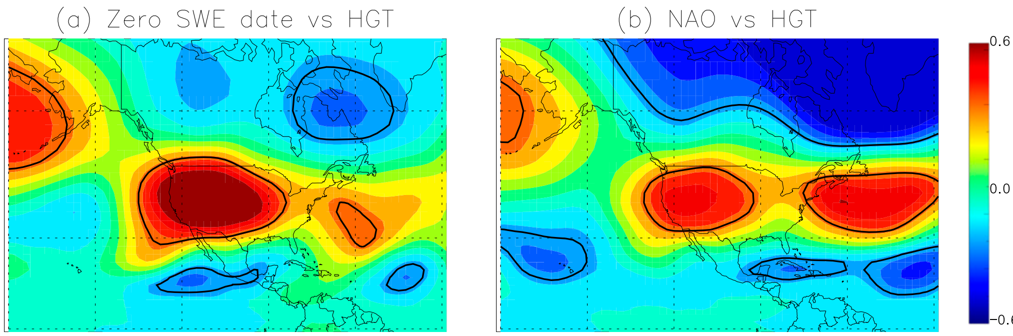

FIG. 4. Correlation map (a) between the domain-averaged Zero SWE date and April–May mean geopotential height (HGT) at 500 hPa and (b) between the April–May mean NAO and April–May mean HGT at 500 hPa. Areas with statistically significant correlations at the 95% confidence level are enclosed in thick solid lines. For direct comparison with (b), the Zero SWE date and HGT correlations in (a) are multiplied by −1.0.

The snowmelt–HGT relationship in Fig. 4a highlights three positive and negative correlation regions that are statistically significant at the 95% level. The thee positive regions appear over the northern Pacific around 60°N, the WUS, and the midlatitude North Atlantic (30° and 45°N); the negative regions occur over the northeastern Canada–south Greenland area and the subtropical Pacific and Atlantic. The spatial pattern of this snowmelt–HGT relationship (e.g., locations of the three positive correlation centers and negative centers in the North Atlantic) is almost analogous to that of the NAO–HGT relationship (Fig. 4b), despite some differences in magnitudes (e.g., tight relationships in North Atlantic for the NAO–HGT link vs tight relationships in North Atlantic for the snowmelt–HGT link). The three positive centers are speculated to be characteristics of Rossby wave propagation from the Pacific to Atlantic Ocean when positive NAO phases are developed in spring (Benedict et al. 2004; Myoung et al. 2015). These results in Fig. 4 suggest that the NAO–snowmelt connection is through development of anomalous upper-level circulations over the WUS region. More specifically, in positive NAO phases, anticyclonic circulations over the WUS region bring warm and dry condition into the UP_SW region and then enhance snowmelt. The opposite is also valid; that is, cyclonic circulations in negative NAO phases bring cold and wet conditions into the region and then delay snowmelt.

The critical role of NAO-related temperature variations on snowmelt is expected to be amplified at lower-elevation stations where surface temperatures often rise above the freezing point in the spring. Figure 5 shows the NAO versus Zero SWE date correlations with respect to the station elevation (y axis). While the correlations below 2800 m are all statistically significant except for one station in UT, one-quarter of the stations in CO, which are located above 2800 m, do not show significant correlations. Such low sensitivity of snowmelt to the NAO at high-elevation stations is due to the fact that the NAO-related temperature fluctuations rarely result in the temperature reaching above the freezing point at these high-elevation stations. This result is consistent with Mote (2006) in addressing the weakened negative relationship between the 1 April SWE and November–March temperature in CA at the stations above 2700–2800 m.

c. *Impact of the NAO on snowfall ratio*

Interestingly, the present study shows that the snowfall ratio is also affected by the NAO. Figure 5a shows the negative correlations of the NAO with the snowfall ratio computed based on the 0°C threshold of Tave. Statistically significant correlations are concentrated in the UT and CO stations. This suggests that the higher the

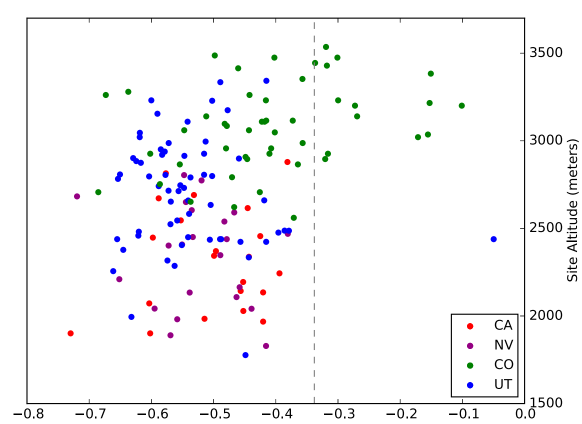

FIG. 5. Correlations (x axis) between the NAO average in the melting period and the Zero SWE date with elevations of SNOTEL stations (y axis). The dashed vertical line indicates the statistically significant level of the correlations at the 95% confidence level. Different colors were used for the stations of each state.

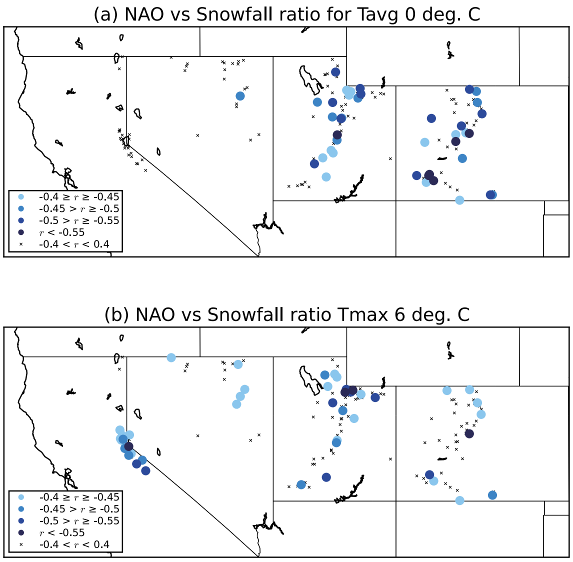

FIG. 6. Correlation map for 24 yr (1991–2014) between the NAO average in the melting period and the snowfall ratio that was computed based on the threshold of (a) Tave < 0°C and (b) Tmax < 6°C. Statistically significant (insignificant) correlations at the 95% confidence level are indicated by colored circles (small crosses).

NAO values, the lower snowfall ratio in the UP_SW. Also, when using a different threshold for the snowfall definition, 6°C of Tmax, significant correlations are observed in most of the study domain, including CA (Fig. 6b). It should be noted that the March–May mean Tmax is higher than the March–May mean Tave by 6°C over the studied stations, which implies that the 6°C threshold with respect to Tmax is somewhat equivalent to the 0°C threshold with respect to Tave. Although the reasons for the different sensitivity of the NAO–snowfall relationships on the snowfall definitions between CA and UT–CO are unclear, the results in Fig. 6 suggest that positive NAO phases tend to enhance snowmelt by reducing the snowfall ratio as well.

4. Summary and conclusions

A number of previous studies showed long-term trends of declining snow accumulation and earlier snowmelt in past decades (Mote 2003; Mote et al. 2005; Stewart et al. 2005), which are serious problems in water resource management in the WUS. For the 34-yr research period (1980–2014) in this study, we also find weak decreasing long-term trends for both the dates of SWE maximum and Zero SWE, but their interannual variability is found to be much greater than their long-term trends (not shown). Our study shows the important role of the NAO on the interannual variability of the snowmelt dates in the UP_SW. As shown in Fig. 2, the NAO is highly correlated with the dates related with snowmelt (i.e., the −50% SWE date and the Zero SWE date) rather than those related with snow accumulation (i.e., +75% SWE). This suggests that snowmelt is strongly affected by NAO phases. The NAO has dominant periods ranging from a few weeks to years and decades; a strong positive tendency of the NAO occurred in the 1980s and 1990s (Yiou and Nogaj 2004; Delworth et al. 2016), which is included in this study.

The relationship between the NAO and snowmelt appears to be the effects of the NAO on temperature and precipitation and/or their combinations through positioning of an upper-level anticyclone (cyclone) over the WUS in positive (negative) NAO phases (Fig. 4). In this study, during the melting periods, the NAO is strongly correlated with the surface air temperature with positive signs (Fig. 3a) and less strongly with precipitation, but still substantially with negative signs (Fig. 3b). These results indicate that both warmer and drier (colder and wetter) conditions under positive (negative) NAO periods tend to result in early (late) snowmelt. Interestingly, the joint effect of temperature and precipitation on snow melting are also represented in the negative correlations between the NAO and the snowfall ratio (Fig. 6). In result, all of these effects seem to strengthen the relationship of the NAO with snowmelt dates. The impact of the NAO on snowmelt through temperature variation is significant below the 2800-m altitude level (Fig. 5); above this level, temperatures largely remain below freezing (Mote 2006) despite the fluctuations according to NAO variability.

Although our study highlights how the NAO affects snowmelt more than snow accumulation (Figs. 2a,b), the NAO can also affect snow accumulation in early spring via rainfall–snowfall partitioning. We have found that the Max SWE date is also strongly correlated with the NAO at most stations (i.e., the larger the NAO, the earlier the Max SWE date) and that the amount of Max SWE is significantly correlated with the NAO at a number of stations (Fig. S2). However, the latter relationships are not as robust as the links of the NAO with snowmelt shown in Fig. 2. One of reasons for the stronger impact of the NAO on snowmelt rather than on snow accumulation may be because surface temperatures during the accumulation periods rarely exceed the freezing point in the highly elevated regions where SNOTEL stations are located, which requires further analysis.

Some possible reasons for the negligible relationship between the NAO and SWE–streamflow over the UP_SW region in the previous studies (e.g., Tootle et al. 2005; Hunter et al. 2006) are attributed to 1) different time scales studied, 2) the relationship with the NAO being stronger in the melting period than in the accumulation period as described above, and 3) the amplification of the impacts of the NAO on weather conditions in specific regions, including the UP_SW, since 1980 (Sun and Wang 2012; Myoung et al. 2015). Consistent to the previous findings in other studies (Clark et al. 2001; Hunter et al. 2006), we have not found statistically significant correlations between ENSO and SWE variables during the snowmelt periods in the UP_SW (Fig. S3). As previously highlighted, our focus region is located in a gray zone where neither ENSO nor the PDO strongly affects local–regional precipitation (Dettinger and Cayan 2014; Myoung et al. 2015). With this in mind, the results in our study are critical in finding the key factor (i.e., the NAO) in modulating weather and snowmelt in the UP_SW region during spring. Within the UP_SW region, the atmospheric circulation patterns and cloud physics that are responsible for wintertime precipitation and snowfall in CA and NV are different from those in UT and CO (Cayon 1996). This contrast may be responsible for the slightly different correlation patterns of the NAO with the weather and snow variables (e.g., Figs. 3 and 6). Despite this, homogenous relationships of the NAO with snowmelt dates suggest the important role of the NAO with regard to spring weather, snowpack, and water resources in the UP_SW in addition to the global warming effect and associated decreasing trends of springtime SWE.

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REFERENCES

Auer, A. H., 1974: The rain versus snow threshold temperature. *Weatherwise*, **27**, 67, doi:10.1080/00431672.1974.9931684.

Barnston, A. G., and R. E. Livezey, 1987: Classification, seasonality and persistence of low-frequency atmospheric circulation patterns. *Mon. Wea. Rev.*, **115**, 1083–1126, doi:10.1175/1520-0493(1987)115<1083:CSAPOL>2.0.CO;2.

Benedict, J. J., S. Lee, and S. B. Feldstein, 2004: Synoptic view of the North Atlantic Oscillation. *J. Atmos. Sci.*, **61**, 121–144, doi:10.1175/1520-0469(2004)061<0121:SVOTNA>2.0.CO;2.

Brown, S. J., J. Caesar, and C. A. T. Ferro, 2008: Global changes in extreme daily temperature since 1950. *J. Geophys. Res.*, **113**, D05115, doi:10.1029/2006JD008091.

Carrera, M., R. Higgins, and V. Kousky, 2004: Downstream weather impacts associated with atmospheric blocking over the northeast Pacific. *J. Climate*, **17**, 4823–4839, doi:10.1175/JCLI-3237.1.

Cayan, D. R., 1996: Interannual climate variability and snowpack in the western United States. *J. Climate*, **9**, 928–948, doi:10.1175/1520-0442(1996)009<0928:ICVASI>2.0.CO;2.

——, S. A. Kammerdiener, M. D. Dettinger, J. M. Caprio, and D. H. Peterson, 2001: Changes in the onset of spring in the western United States. *Bull. Amer. Meteor. Soc.*, **82**, 399–415, doi:10.1175/1520-0477(2001)082<0399:CITOOS>2.3.CO;2.

Clark, M. P., M. C. Serreze, and G. J. McCabe, 2001: Historical effects of El Niño and La Niña events on the seasonal evolution of the montane snowpack in the Columbia and Colorado River basins. *Water Resour. Res.*, **37**, 741–757, doi:10.1029/2000WR900305.

Clow, D. W., 2010: Changes in the timing of snowmelt and streamflow in Colorado: A response to recent warming. *J. Climate*, **23**, 2293–2306, doi:10.1175/2009JCLI2951.1.

Dai, A., 2008: Temperature and pressure dependence of the rain–snow phase transition over land and ocean. *Geophys. Res. Lett.*, **35**, L12802, doi:10.1029/2008GL033295.

Delworth, T. L., F. Zeng, G. A. Vecchi, X. Yang, L. Zhang, and R. Zhang, 2016: The North Atlantic Oscillation as a driver of rapid climate change in the Northern Hemisphere. *Nat. Geosci.*, **9**, 509–512, doi:10.1038/ngeo2738.

Dettinger, M., and D. R. Cayan, 2014: Drought and the California delta—A matter of extremes. *San Francisco Estuary Watershed Sci.*, **12**, 1–6.

Francis, J. A., and S. J. Vavrus, 2012: Evidence linking Arctic amplification to extreme weather in mid-latitudes. *Geophys. Res. Lett.*, **39**, L06801, doi:10.1029/2012GL051000.

Gershunov, A., and T. Barnett, 1998: Interdecadal modulation of ENSO teleconnections. *Bull. Amer. Meteor. Soc.*, **79**, 2715–2725, doi:10.1175/1520-0477(1998)079<2715:IMOET>2.0.CO;2.

Häkkinen, S., P. B. Rhines, and D. L. Worthen, 2011: Atmospheric blocking and Atlantic multidecadal ocean variability. *Science*, **334**, 655–659, doi:10.1126/science.1205683.

Hamlet, A. F., P. W. Mote, M. P. Clark, and D. P. Lettenmaier, 2005: Effects of temperature and precipitation variability on snowpack trends in the western United States. *J. Climate*, **18**, 4545–4561, doi:10.1175/JCLI3538.1.

Higgins, R. W., A. Leetmaa, Y. Xue, and A. Barnston, 2000: Dominant factors influencing the seasonal predictability of U.S. precipitation and surface air temperature. *J. Climate*, **13**, 3994–4017, doi:10.1175/1520-0442(2000)013<3994:DFITSP>2.0.CO;2.

Hunter, T., G. Tootle, and T. Piechota, 2006: Oceanic–atmospheric variability and western U.S. snowfall. *Geophys. Res. Lett.*, **33**, L13706, doi:10.1029/2006GL026600.

Hurrell, J. W., and H. van Loon, 1997: Decadal variations in climate associated with the North Atlantic Oscillation. *Climatic Change*, **36**, 301–326, doi:10.1023/A:1005314315270.

Kalnay, E., and Coauthors, 1996: The NCEP/NCAR 40-Year Reanalysis Project. *Bull. Amer. Meteor. Soc.*, **77**, 437–471, doi:10.1175/1520-0477(1996)077<0437:TNYRP>2.0.CO;2.

Kienzle, S. W., 2008: A new temperature based method to separate rain and snow. *Hydrol. Processes*, **22**, 5067–5085, doi:10.1002/hyp.7131.

McAfee, S. A., and J. L. Russell, 2008: Northern annular mode impact on spring climate in the western United States. *Geophys. Res. Lett.*, **35**, L17701, doi:10.1029/2008GL034828.

McCabe, G. J., and M. D. Dettinger, 1999: Decadal variations in the strength of ENSO teleconnections with precipitation in the western United States. *Int. J. Climatol.*, **19**, 1399–1410, doi:10.1002/(SICI)1097-0088(19991115)19:13<1399::AID-JOC457>3.0.CO;2-A.

——, and M. P. Clark, 2005: Trends and variability in snowmelt runoff in the western United States. *J. Hydrometeor.*, **6**, 476–482, doi:10.1175/JHM428.1.

Mote, P. W., 2003: Trends in snow water equivalent in the Pacific Northwest and their climatic causes. *Geophys. Res. Lett.*, **30**, 1601, doi:10.1029/2003GL017258.

——, 2006: Climate driven variability and trends in mountain snowpack in western North America. *J. Climate*, **19**, 6209–6220, doi:10.1175/JCLI3971.1.

——, A. F. Hamlet, M. P. Clark, and D. P. Lettenmaier, 2005: Declining mountain snowpack in western North America. *Bull. Amer. Meteor. Soc.*, **86**, 39–49, doi:10.1175/BAMS-86-1-39.

Myoung, B., and Y. Deng, 2009: Interannual variability of the cyclone activity along the U.S. Pacific Coast: Influences on the characteristics of winter precipitation in the western United States. *J. Climate*, **22**, 5732–5747, doi:10.1175/2009JCLI2889.1.

——, S. H. Kim, J. Kim, and M. Kafatos, 2015: On the relationship between the North Atlantic Oscillation and early warm season temperatures in the southwestern United States. *J. Climate*, **28**, 5683–5698, doi:10.1175/JCLI-D-14-00521.1.

Pederson, G. T., S. T. Gray, T. Ault, W. Marsh, D. B. Fagre, A. G. Bunn, C. A. Woodhouse, and L. J. Graumlich, 2011: Climatic controls on the snowmelt hydrology of the northern Rocky Mountains. *J. Climate*, **24**, 1666–1687, doi:10.1175/2010JCLI3729.1.

Ropelewski, C. F., and M. S. Halpert, 1987: Global and regional scale precipitation patterns associated with the El Niño/Southern Oscillation. *Mon. Wea. Rev.*, **115**, 1606–1626, doi:10.1175/1520-0493(1987)115<1606:GARSPP>2.0.CO;2.

Serreze, M. C., M. P. Clark, R. L. Armstrong, D. A. McGinnis, and R. S. Pulwarty, 1999: Characteristics of the western United States snowpack from Snowpack Telemetry (SNOTEL) data. *Water Resour. Res.*, **35**, 2145–2160, doi:10.1029/1999WR900090.

Stewart, I. T., 2009: Changes in snowpack and snowmelt runoff for key mountain regions. *Hydrol. Processes*, **23**, 78–94, doi:10.1002/hyp.7128.

——, D. R. Cayan, and M. D. Dettinger, 2005: Changes towards earlier streamflow timing across western North America. *J. Climate*, **18**, 1136–1155, doi:10.1175/JCLI3321.1.

Sun, J., and H. Wang, 2012: Changes of the connection between the summer North Atlantic Oscillation and the East Asian summer rainfall. *J. Geophys. Res.*, **117**, D08110, doi:10.1029/2012JD017482.

Thompson, D. W. J., and J. M. Wallace, 1998: The Arctic Oscillation signature in the wintertime geopotential height and temperature fields. *Geophys. Res. Lett.*, **25**, 1297–1300, doi:10.1029/98GL00950.

——, and ——, 2001: Regional climate impacts of the Northern Hemisphere annular mode. *Science*, **293**, 85–89, doi:10.1126/science.1058958.

——, ——, and G. C. Hegerl, 2000: Annular modes in the extratropical circulation. Part II: Trends. *J. Climate*, **13**, 1018–1036, doi:10.1175/1520-0442(2000)013<1018:AMITEC>2.0.CO;2.

Tootle, G., T. Piechota, and A. Singh, 2005: Coupled oceanic–atmospheric variability and U.S. streamflow. *Water Resour. Res.*, **41**, W12408, doi:10.1029/2005WR004381.

Wettstein, J. J., and L. O. Mearns, 2002: The influence of the North Atlantic–Arctic Oscillation on mean, variance, and extremes of temperature in the northeastern United States and Canada. *J. Climate*, **15**, 3586–3600, doi:10.1175/1520-0442(2002)015<3586:TIOTNA>2.0.CO;2.

Yiou, P., and M. Nogaj, 2004: Extreme climatic events and weather regimes over the North Atlantic: When and where? *Geophys. Res. Lett.*, **31**, L07202, doi:10.1029/2003GL019119.