Snow and Albedo Climate Change Impacts across the United States Northern Great Plains

S.R. Fassnacht[1,2,3*], M.L. Cherry[1,4], N.B.H. Venable[5], F. Saavedra[5]
[1] ESS-Watershed Science, Colorado State University, Fort Collins, Colorado 80523-1476, USA
[2] Cooperative Institute for Research in the Atmosphere, Fort Collins, Colorado 80523-1375 USA
[3] Geospatial Centroid at CSU, Fort Collins, Colorado 80523-1019 USA
[4] now with: Department of Geography, University of Victoria, David Turpin Building B203, 3800 Finnerty Road (Ring Road), Victoria, BC V8P 5C2, Canada
[5] EASC-Watershed Science, Colorado State University, Fort Collins, Colorado 80523-1482, USA

* corresponding author:
S.R. Fassnacht
ESS-Watershed Science
Colorado State University
Fort Collins, Colorado 80523-1476, USA
<*steven.fassnacht@colostate.edu*>
phone: 001.970.491.5454
fax: 001.970.491.1965

Abstract

In areas with a seasonal snowpack, a warmer climate could cause less snowfall, a shallower snowpack and a change in the timing of snowmelt, all which could reduce the winter albedo and yield an increase in net shortwave radiation. Trends in temperature, precipitation (total and as snow), days with precipitation and snow, and winter albedo were investigated over the 60-year period from 1951 to 2010 for 20 meteorological stations across the Northern Great Plains. This is an area where snow accumulation is shallow but persistent for most of the winter (November through March). The most consistent trends were minimum temperature and days with precipitation, both of which increased at a majority of the stations. Among the stations included, a decrease in the modeled winter albedo was more prevalent than an increase. There was substantial spatial variability in the climate trends. For most variables, the period of record used influenced the magnitude and sign of the significant trends.

Keywords: trends, temperature, precipitation, snow, albedo

1 INTRODUCTION

While global annual temperatures have increased by 0.74 °C in the past century and 1.3 degrees °C in the past 50 years, these temperature increases are not consistent across the globe (IPCC, 2007). In some locations temperatures are increasing much more rapidly than the global average rate, and increases are not uniform for annual maximum and minimum temperatures. Trends in annual minimum temperatures are important to snowpack properties. Snowfall and snowpack trends are an important issue in semi-arid to arid climates where water demand already surpasses supply (Stewart, 2009).

In non-polar regions, an overall warmer climate could yield less snowfall, a shallower snowpack, and a change in the timing of the snowmelt (Stewart, 2009). The Western United States has already seen earlier snowmelt and peak discharge in snow dominated river systems of up to 20 days earlier (Stewart, 2009). The Western United States has also seen widespread declines in springtime snow water equivalent (SWE) from 1925 to 2000 (Mote et al., 2005).

Numerous studies have examined the snow cover and its variability across the Northern Hemisphere, with most highlighting a decrease over the period of record. Station data for North America have shown that the end date of snow cover has not changed from 1980 to 2006, although some temperatures have been warming for the same region (Peng et al., 2013). Using up to eight sources, including station and modeled data, the March and April snow cover extent (SCE) were constant or slightly decreasing from about 1960 to the early 1980s, then a decrease through the 1980s, followed by an increase in the early 1990s (Brown and Mote, 2011). After that the March SCE was relatively stable through 2010 but decreasing in April (Brown and Mote, 2011).

There has also been a decrease in precipitation as snow across the Western United States (Knowles et al., 2006). Decreases are amplified when the average wintertime temperatures remain around 0 degrees Celsius (Knowles et al., 2006). The ratio of precipitation as snow has also been decreasing across the Northeastern United States and the contiguous United States region (Huntington et al., 2004; Feng and Hu, 2007, respectively). Over some of the previously studied regions, there has been both an overall decrease in the amount of snowfall and an increase in the amount of annual precipitation. These trends are correlated to winter temperature increases and are a cause for concern as snow cover acts a control for summer soil-water storage and without long periods of snow cover, crop lands like those found across the Northern Great Plains will become drier (Feng and Hu, 2007; Stewart, 2009). Somewhere between 90-120 snow covered days in the Northern Great Plains, with a thin maximum SWE of between 20-40 mm. Brown and Mote (2009) used a number of climate models to simulate snowpack changes and found that the largest decrease in the future was for snow covered days, maximum SWE decreasing only a small amount.

Snowcover has been directly linked to seasonal snow cover is an important part of the energy budget due to its low thermal conductivity and high albedo (Stewart, 2009). Similarly, the snow albedo feedback is a concept used in climate models to define the sensitivity of the atmospheric energy balance to net solar radiation due to a change in snowpack albedo (Qu and Hall, 2006). Reductions in the albedo of a snowpack surface, such as due to the presence of dust, can drastically increase rate of melt and for deep snowpack result in snow-free conditions a month or more earlier than clean snow conditions (e.g., Painter et al., 2007). The increased absorption of solar radiation due to a decreased albedo is much more important in the melting of snow than the small increases in longwave radiation due to temperature increases (Painter et al., 2010).

Changes in the amount of precipitation as snowfall and the number of days with snowfall are both important to the water and energy budget of an area. Reductions in overall amount of precipitation as snow should likely lower the depth of the snowpack and could possibly decrease the persistence of snowcover which could decrease the overall winter albedo. The decreased albedo would increase the absorption the incoming solar radiation. A reduction in the number of days with snowfall would reduce the amount of fresh snow and therefore reduce the overall albedo of the snowpack resulting in an increase in the amount of radiation that is absorbed.

The objectives of this paper are to determine 1) if the amount of precipitation falling as snow is changing, 2) if the number of days with snowfall is changing, and 3) if these changes impact modeled albedo over the winter period, which defined as November through March. Annual precipitation and temperature changes will also be evaluated. This investigation will also focus on two stations in close proximity to one another to examine small-scale spatial changes and the influence of length of record.

2 DATA

For this work the Northern Great Plains (NGP) area is defined as western parts of Kansas, Nebraska and South Dakota as well as the eastern parts of Colorado and Wyoming (Figure 1). This area represents a transition from the more water abundant East to the more arid region bordering the Front Range of the Rocky Mountains, and as such plays an important role in crop production in the United States <*http://www.globalchange.gov/*>. Snowpack accumulation is shallow but persistent for most of the winter. The Köppen-Geiger Classification for the area is mostly semi-arid (BSK) with small parts of the humid continental (DFA) classification (Peel et al., 2007).

Twenty stations (Figure 1) were identified for this region that had less than 30% of all years excluded due to missing data, where missing years were those with more than 14 days of missing data (Reek et al., 1992). The stations were located from 39.4 to 44.7 degrees north latitude and 99.8 to 105 degrees longitude with elevation ranging from 733 to 1880 meters above sea level. Daily data, including maximum temperature, minimum temperature, precipitation, and snowfall, were retrieved for the period from 1951 to 2010 from the National Climatic Data Center (NCDC) <*www.ncdc.noaa.gov*>, and summarized for the annual analysis.

The stations were all part of the NCDC cooperative (COOP) network. Often such stations are moved, especially over a long period of record, such as the 60 years examined here (e.g., Groisman and Legates 1994; Groisman et al. 1996; Peterson et al., 1998). Such moves can created discontinuities. As such, nine of the stations (denoted by an asterisk in Figures 1-3) were part of the US Historical Climatology Network (USHCN) that are considered a high quality data set of basic meteorological variables (Menne et al., 2015). The station metadata were reviewed to examine station moves and thus possible discontinuities in the record. All nine of the USHCN stations were moved over the 60 year of investigation but these moves were usually less than several kilometers. Three of the remaining stations (Chadron, North Platte and Yuma) were not moved and thus considered consistent. Six stations (Belle Fourche, Benkelman, Goodland, Lexington, Newell and Ogallala) were moved less than 0.5 km and most only 150 m. Edgemont was moved in 2009 at the end of the period of investigation, thus this was not considered to bias the trend analysis. Some of the station moves appeared to be back to or in close proximity of previous locations. This appears to be the case for Sterling that was moved in 1983 and back in 2004.

3 METHODS

For each station, the annual average maximum and minimum temperature were calculated for each year in degrees Celsius. The total precipitation (mm) and snowfall (mm), from precipitation as snow, as well as the number of days with precipitation and snow were also calculated for each year (Huntington et al., 2004). The amount of precipitation as snow was the total daily precipitation when snowfall was observed. Air temperature was not used as a threshold between rain and snow, as this varies based on climate (Fassnacht and Soulis, 2002; Fassnacht et al., 2013).

We did not attempt to correct for snowfall that melted before being measured (Huntington et al., 2004). This means that snowfall that fell and melted before it was measured was counted as rainfall. This study does not take into consideration days that may have had mixed precipitation (where both rain and solid precipitation fall in the same day). This will overestimate the amount of snowfall since it includes days with both rain and snow entirely as snowfall. However, since precipitation gauges are less efficient at measuring snowfall than rainfall there is an error both underestimating and overestimating the total amount of snowfall. There was no attempt made to quantify the error caused by the assumptions in this study (see Huntington et al., 2004; Knowles et al., 2006). While trends have been observed in wind speed (e.g., Hoover et al., 2014), it was assumed these were limited across the study domain and that there was no trend in the amount of undercatch. In particular, wind induced undercatch (e.g., Yang et al., 1998) was not computed since wind data were not available. Similar assumptions have been applied to studies in other regions with similar climate such as Mongolia (Fassnacht et al., 2011).

Various snow albedo (α_s) models have been created with different data requirements. The simplest models are a three linear segment shallow snowpack model (Gray and Landine, 1987) and a first order decay model (Verseghy, 1991) originally derived by the US Army Corps of Engineers (1956). Other models exist, such as the sub-model used in the CROCUS snow model, where α_s varies as a function of wavelength for three spectral bands and snow particle size (Brun et al., 1992; Vionnet et al., 2012). The Greuell and Konzelmann (1994) formulation considers the density of snow and the interaction of snow and ice albedos. The SNICAR albedo model (Flanner and Zender, 2005; Flanner *et al*., 2007) further considers surface temperature and near-surface temperature gradients (Flanner and Zender, 2006) to simulate changes in snow particle size and shape. However, only precipitation, snowfall, snow on the ground, and temperature data were available at a daily time step for this current study, this the simple α_s decay model was used to determine trends.

Daily albedo was modeled over the winter period (November through March) using meteorological data (temperature, precipitation, and snow on the ground). Since observations of albedo and surface characteristics were not available, a time variant first order decay model was used. This model takes the form:

$$\alpha_{s(t)} = [\alpha_{s(t-1)} - \alpha_{s\text{-}min}]\, e^{-k\Delta t} + \alpha_{s\text{-}min} \qquad (1),$$

where $\alpha_{s(t)}$ and $\alpha_{s(t-1)}$ are the albedo at the current and previous time step, t, $\alpha_{s\text{-}min}$ is the minimum allowable albedo, k is a decay coefficient, and Δt is the time step (Verseghy, 1991). This model is incorporated into the Canadian Land Surface Scheme (Verseghy, 1991) to model snow conditions (Brown et al., 2006). In this paper, the decay (0.01 per hour converted to a daily rate) used by Verseghy (1991) was implemented. It uses three conditions: 1) for fresh snow $\alpha_{s(t)}$ is set at 0.84, 2) for dry snow (no melt), $\alpha_{s(t)}$ decays to an $\alpha_{s\text{-}min}$ of 0.70, and 3) for melting snow, $\alpha_{s(t)}$ decays to an $\alpha_{s\text{-}min}$ of 0.50. Using the same albedo model in the Canadian Land Surface Scheme, Langlois et al. (2014) proposed a threshold of 0.5 cm to reset the albedo to 0.84. Herein, fresh snow was considered to reset the albedo to 0.84 when at least 2.54 cm of snowfall was observed over a day, since this was the resolution of fresh snow depth measurements (1 inch). When no snow was present, a soil albedo of 0.20 was used.

The significance of each climatological trend was determined using the Mann-Kendall test (Gilbert, 1987). This is a robust non-parametric test that does not assume a specific distribution of the data nor is it influenced by extreme values/outliers or missing values. For all stations, trends were analyzed from the entire period of record (1951 to 2010). For the Sterling, Colorado and Kimball, Nebraska stations, the periods from 1951 to 1980 and from 1981 to 2010 were also investigated to see if there was a difference in the changes in trends over a shorter time period (Venable et al., 2012). When a trend was significant, the rate of the change was estimated using the Sen's slope estimator, which is the median slope of all pairs analyzed, giving the overall rate of change. For this analysis the MAKESENS macro developed for Excel spreadsheets was used (Salmi et al., 2002) to identify trends at both the $p<0.05$ and $p<0.10$ level.

To further investigate changes at these two locations over the later 30 years, the North American Regional Reanalysis (NARR) data were also evaluated. Monthly data were acquired from the NARR dataset from the NOAA/OAR/ESRL PSD, Boulder, Colorado, USA, at their website <http://www.esrl.noaa.gov/psd/> (Mesinger et al., 2006). The NARR data are a combination of modeled (the National Center for Environmental Prediction Eta model) and assimilated (Regional Data Assimilation System, RDAS) station data at a spatial resolution of 32 km available for the period from 1979 to present. The point values from the two comparison stations (Sterling and Kimball) were extracted by location. The yearly summary by calendar year for air temperature and precipitation and from November to March from albedo dataset was tested for trend analysis.

Trends are often computed to assess regional climate patterns. The Regional Kendall Test (RKT of Helsel and Frans, 2006) combines a number of stations in close proximity to estimate the Mann-Kendall significance and Sen's slope rate of change. Since Clow (2010) used RKT to assess trends for station groups over 10s to 100s of kilometers in the mountain of Colorado, it is also used here as the terrain in the NGP is less complex that in the mountains and the climate should be less varying.

4 RESULTS

Average annual maximum (Tmax) and minimum (Tmin) temperatures varied by less than 4 °C, ranging from 14.3 to 19.2 and -0.5 to 3.2 °C, respectively, while annual total precipitation (P) varied by almost four-fold from 277 to 1016 mm and precipitation as snow (P as snow) by six-fold from 46 to 349 mm annually (Figure 1). The range in numbers of days with precipitation (P days) and snowfall (snow days) varied similarly, but the average modeled winter albedo only varied from 0.53 to 0.6 (Figure 1).

Climate trends were generally spatially variable with warming of minimum temperatures and more days with precipitation illustrating the most consistent trends across the 20 stations (Figures 2 and 3). Minimum temperatures increased at 9 stations averaging 2.74 degrees Celsius per century, with significant Tmin cooling only at Kimball. Eleven stations saw an average increase of 28.3 days with precipitation per century while only Alliance had fewer precipitation days. Tmax trends were fewer (4 significant and 2 less significant), with four warming and two cooling. Precipitation totals increased at four locations by an average of 182 mm per century and decreased at one. Changes in the amount of precipitation as snow are more prevalent with seven stations increasing by an average of 103 mm per century and five decreasing by an average of 78 mm per century. Four stations had more snow days by an average of 16.6 per century, while seven saw an average of 23.2 fewer snow days per century. Alliance was the only station with a decrease in both P and snow days, and these were the largest changes. Albedo increased at four stations by an average of 6.6% per century and decreased at seven stations by an average of 8.6% per century.

There are some identifiable spatial patterns in the climate trends. For example, in the southwest parts of the NGP Tmax and/or Tmin increased at most stations (Figure 3), but there was a decrease in Tmax at Torrington and in Tmin at Kimball, respectively. Other spatially coherent areas included increasing P totals and days with P in the east, and increasing P as snow, days with P and days with snowfall in the southeast, with some corresponding increases in albedo. However, adjacent stations showed decreasing snow and some decreasing albedo. In the north and north-west parts, snow days and albedo decreased, but Belle Fourche showed the opposite trend. In the south, Sterling, Goodland and Oberlin the modeled winter albedo increased (Figure 3).

Significant trends were observed with the RKT for Tmax (rate of 0.625 deg C/100 years), Tmin (rate of 1.17 deg C/100 years, P amount (rate of 105.3 mm/100years), and days with P (rate of 15.6 /100 years), while days with snowfall decrease (rate of -1.82 /100 years) and albedo decreases (-0.007 /100 years) were only significant at the $p<0.10$ level (Figures 2 and 3). These were similar to the averages of the individual stations. The RKT trends in P as snow were not significant (Figure 2). The magnitude of regional trends tended to be much less than what was observed at individual stations as no systematic trends were observed across the entire domain (Figure 3).

Sterling in northeast Colorado and Kimball in southwest Nebraska are 73 km apart with an elevation difference of 380 meters (highlighted in Figures 2 and 3). Opposite trends in winter albedo were estimated (Figure 4d), in part as a function of differences in temperature (Figure 4a), precipitation as snow (Figure 4b), and days with precipitation and snow (Figure 4c).

Average temperature trends were similar for the station and NARR data (Table 1), while total precipitation was decreasing from the station data yet increasing from the NARR data. Precipitation as snow was decreasing at twice and more than 10 times the rate for the station data compared to the NARR data at Kimball and Sterling, respectively. Albedo trends were of opposite sign and different magnitude between the two datasets. The minimum station temperature and average NARR temperature trends at Sterling were significant; no other variable had a significant trend for both the station and NARR datasets (Table 1).

5 DISCUSSION

The RKT trends showed increases for four variables (at $p<0.05$ level) and decreasing trends ($p<0.10$ level) for two variables (Figure 2) that were similar to the averages of the individual stations. However, since these were regional trends, they tended to be much less than what was observed at individual stations as no systematic trends were observed across the entire domain (Figure 3). Considering the relative terrain homogeneity of the study domain, high levels of spatial and temporal variability suggest a need to understand better climate change across the NGP (Hudson et al., 1983). In a similar terrain and climate, the variability in temperature trends was extensive over a study of eastern Colorado (Pielke et al., 2002). For example, Fort Collins and Fort Morgan saw significant increases in both Tmax and Tmin, while Kimball saw less significant increase in Tmax but a decrease in Tmin (Figure 2 and 3). This highlights the importance of considering the scale of analysis. It is possible that a small increase in the winter temperatures can affect snowfall amounts (Fassnacht et al., 2013). The average annual minimum temperatures for almost all stations in this study remain above freezing (approx. -5 to 10 degrees C). Previous studies have shown that the Northern Great Plains is one of the regions of the United States experiencing increased temperatures as well as a decrease in the amount of annual snowfall (Knowles et al., 2006). The results presented herein include an additional 10 years of data and exhibit less consistency in changes to the amount of snowfall. There has also been a decrease in the ratio of snowfall to rainfall in these regions (Knowles et al., 2006). Feng and Hu (2007) reported a decrease in the ratio of precipitation as snow at stations in the Northern Great Plains, however only three of the stations used in this study showed such a decrease (i.e. Harrison, Kimball, Yuma). Previous work (Feng and Hu, 2007) also found an overall decrease in the amount of snowfall and an increase in total precipitation, however, this study found more variability regardless of an overall trend in the decrease or increase of snowfall or precipitation.

There is some correlation between trends of the different variables, which could be relevant to albedo as some of them are used in the model. For example, there is a weak correlation between days with snow and maximum or minimum temperature with 19% of the variance explained when all 20 trends are used. However, few stations see significant trends among a number of different variables (Figure 2). For example, seven stations had significant trends in amount of precipitation as snow and minimum temperature, with 24% of the variance explained. Trends in temperature and precipitation amounts (total and as snow) were not correlated to albedo trends (see Figure 2).

The length and period of record examined also influences the rates of change and the level of significance (Venable et al., 2012). There was little consistency in the trends over the two 30-year periods (Figure 5). Of the seven variables, only Tmin at Kimball was significantly getting cooler for all time periods (Figure 5b). Conversely there was strong cooling of Tmax early in the period of record but overall there was a weaker degree of warming at Kimball. Similarly, at Sterling, P as snow decreased for the first 30 years then increased, yielding a less significant increase overall (Figure 5). This result may illustrate cyclical trends that become highlighted when analyzing shorter time periods (Chen and Grasby, 2009). These trends also somewhat mirror global patterns of cooling from the 1940s to mid-1970s and warming thereafter.

Most of the stations did move over the 60 years of data collection, as explained in the Study Site section. Stations moves can cause time series discontinuity (Groisman and Legates, 1994; Peterson et al., 1998). While the Sterling station did move about 1.6 km twice (1983 and 2004), these potential discontinuities are not present in the time series (Figure 4i). There was a third move in 2010 of about 100 m. There was no indication of changes in the observation time at this station, as that has also been shown to cause discontinuity in time series (Groisman and Legates, 1994; Peterson et al., 1998). It should also be noted that the Northern Great Plains are relatively flat and thus any change in elevation from a station move is negligible. For example, the Goodland station was moved about 500 meters, but it remained at the airport so the change in elevation could be assumed to be only in the order of meters.

In climate change modeling efforts based on data for the northern prairies in Canada, snow cover duration was less sensitive to temperature changes than in other regions (Brown and Mote, 2009), but trend analyses from the Northern Great Plains for 1910-1993 show high variability in seasonal snow cover durations. Wintertime increases in snow cover duration were linked with significant increases in seasonal snowfall amounts, though no significant changes were seen in total precipitation (Hughes and Robinson, 1996). It may be that these factors sometimes work together and sometimes at cross-purposes. For example, a warmer winter may be likely to have a shorter snow-cover duration because of melt, or a longer snow-cover duration if there were increases in snowfall during what would normally be the colder winter periods.

Albedo is correlated to snowfall amounts (for example, explained variance of 42 and 45% for Kimball and Sterling), but albedo is more strongly correlated to days with snowfall (explained variance of 62 and 81% at the focus sites). This stronger relation is due to the high albedo of fresh snow. While previous studies (e.g. Huntington et al., 2004; Knowles et al., 2006; Feng and Hu, 2007) examined changes in the amount of precipitation as snow, and others examined changes to snow covered area and duration of snow cover (e.g. Brown and Mote, 2009; McCabe and Wolock, 2010), a quantification of changes in the days with precipitation and days with snowfall are important for understanding changes to albedo. Snow-albedo feedbacks are enhanced by increasing air temperatures and increasing total solar radiation as occurs in springtime. Hernández-Henríquez et al. (2015) found generally low magnitude, variably increasing and decreasing fractions of time that mid-latitude sites (such as the NGP) were snow covered, but conclude that snow cover extend trends in late spring and early summer have the greatest impact on the snow-albedo feedback and surface radiation budget input, particularly at higher latitudes. It is thus recommended that days with snowfall be computed as an additional indicator of climate change and that seasonal analysis, particularly of spring datasets be conducted in spatially and temporally variable snow covered regions like the NGP. While the arctic, especially sea ice regions, and the mountains of the western U.S., show an increase in the energy balance for snow and ice based regions, there is a possible negative trend for the Northern Great Plains area (Flanner et al., 2011).

The NARR dataset was evaluated as it includes albedo data for the second 30-year period of analysis. However, the trends in albedo were of opposite sign and different magnitude (Table 1), but no trends were significant. There was greater interannual variability in the NARR albedo than the station data, but the NARR winter albedo (averaging about 0.31 for the 32 km pixels about each site) is much less than the modeled station albedo (averaging 0.59 at Kimball and 0.55 at Sterling). For the Northern Great Plains, the maximum surface albedo from the dataset compiled by Robinson and Kukla (1985) was from 0.71 to 0.80; these data are used to model the NARR dataset. The maximum surface albedo was derived at a 1 degree (latitude and longitude) resolution from satellite data; there can be a large difference between point measurements of albedo and satellite-based estimated due to inhomogeneities such as partial snow cover (Arola et al., 2003). The low values of albedo during snow conditions is consistent among numerous climate/land surface models, even in non-forested areas where the canopy does not need to be considered (Qu and Hall, 2007). Further, the decrease is surface albedo is much more dominated by the melting and thus disappearance of snow, rather than the metamorphism and the addition of meltwater to the surface of the snowpack (Qu and Hall, 2007). New additions to the snow albedo model in the land surface scheme used to generate the NARR dataset uses a modification of the albedo decay model presented in equation 1 (Livneh et al., 2010).

The constant soil albedo of 0.2 can create problems when snow free conditions persist in the winter months. In the Community Land Model version 4.0, the albedo of soil is a function of color, wetness, and wavelength, such that it can vary between 0.04 in the visible for saturated dark soil to 0.61 in the near infrared for dry, light soil (Oleson et al., 2013). The identification of soils and vegetation at the 20 stations was not undertaken.

For the Northern Great Plains, the winter albedo of snow free areas will vary over space, but not necessarily over time due to the dormant nature of the vegetation. In the prairie regions, the grasses can be up to 50 cm tall yet will lie down during snow accumulation yielding 10 to 20 cm high vegetation. Thus 10 to 20 cm of snow is required to completely cover such vegetation. However, the dataset used herein are collected at airports and near towns with grass areas that are landscaped, thus the vegetation is only 3 to 5 cm high which allows it to be buried much quicker than native, non-landscaped prairie vegetation. Formulations do exist to consider the burial of vegetation by snow (e.g., Wang and Zeng, 2009), but this is beyond the scope of this paper.

Using simple snow albedo models is not uncommon; recently the European Centre for Medium-Range Weather Forecasts (ECMWF) was using the snow albedo model presented herein for melting conditions and a linear model for non-melting conditions, based on the Météo-France climate model (Douville et al., 1995). There are limitations in the Verseghy (1991) albedo model since the decrease in α_s is a function of time and air temperature that dictates the state (accumulating or melting) of the snowpack. From the one available dataset at the time, Flanner and Zender (2006) illustrated that the default albedo decay coefficient used by Verseghy (1991) was too large; a slower decrease in albedo computed using a smaller decay ($k < 0.01$ /h) would yield a higher winter albedo when snow was present. Other models, however, could not be used, as snow specific data were not available for the 60-year period of analysis. Such data include snow particle size or shape, the density of snow, surface temperature and near-surface temperature gradients. This paper illustrates the influence of changing amounts of snow and the occurrence of snowfall using the available meteorological data. The latter variable has not been shown in the literature yet.

In the future, more snow data will be available from *in situ* and remote sensing sources that will allow for the use of more physically-based albedo estimates and models. Additional information to improve snow albedo estimates will include the presence of light absorbing particulates, such as black carbon, dust, and needle litter in forested regions (Flanner et al., 2007; Painter et al., 2007; Boon, 2009]. Each of these constituents can play a relevant role in lowering the snow albedo.

6 CONCLUSION

The amount of precipitation falling as snow is changing at a majority of the stations analyzed. Trends derived from the Regional Kendall Test for all twenty stations showed the dominant trends among the stations but not the local variability. There were more stations with a decrease in the days with snowfall than an increase.

The number of days with snowfall was most important in the modeling of albedo, while the amount of precipitation as snow was less important. Albedo trends mirrored the direction of number of snow days trends at most stations, with albedo decreasing significantly at 7 stations and increasing at 4 others. This affects the winter energy balance.

Temperatures are warming at some of the sites. However, climate trends are not consistent over space or over different time periods, illustrating the relevance of the scale of analysis. There is substantial spatial variability in trends across the study domain, with opposite trends in nearby stations. The period of analysis also influence the significance and rates of change.

Acknowledgements

Thanks to John Stednick of the Colorado State University (CSU) for his input and to the CSU Honors Program for supporting MLC. Partial funding was provided by the NASA Terrestrial Hydrology Program Project award NNX11AQ66G (PI Dr. Michael Jasinski NASA GSFC). In kind funding was provided by the NSF Dynamics of Coupled Natural and Human Systems award BCS-1011801 (PI Maria Fernandez-Gimenez CSU). We would like to thank the handling editor Dr. Philip Marsh and two anonymous reviewers for insightful comments that improved this paper.

References

Arola A, and Coauthors, 2003: A new approach to estimating the albedo for snow-covered surfaces in the satellite UV method. Journal of Geophysical Research 108: 4531, doi:10.1029/2003JD003492.

Boon S (2009) Snow ablation energy balance in a dead forest stand. Hydrological Processes 23: 2600-2610.

Brown R, Mote P (2009) The response of Northern Hemisphere snow cover to a changing climate. Journal of Climate 22(8): 2124-2145.

Brown RD, Robinson DA (2011). Northern Hemisphere spring snow cover variability and change over 1922–2010 including an assessment of uncertainty. The Cryosphere, 5, 219–229.

Brown R, Bartlett P, MacKay M, Verseghy D (2006) Evaluation of snowcover in CLASS for SnowMIP. Atmosphere-Ocean 44: 223-238 [doi:10.3137/ao.440302].

Brun E, David P, Sudul M, Brunot G (1992) A numerical model to simulate snow-cover stratigraphy for operational avalanche forecasting. Journal of Glaciology 38: 13-22.

Chen Z, Grasby S (2009) Impact of Decadal and Century-Scale Oscillations on Hydroclimate Trend Analyses. Journal of Hydrology 365(1-2): 122-133.

Clow D (2010) Changes in the timing of snowmelt and streamflow in Colorado: A response to recent warming. Journal of Climate 23(9): 2293-2306.

Douville H, Royer JF, Mahfouf JF (1995). A new snow parameterization for the Météo-France climate model. Climate Dynamics 12: 21–35.

Fassnacht SR, Soulis ED (2002) Implications during transitional periods of improvements to the snow processes in the Land Surface Scheme - Hydrological Model WATCLASS. Atmosphere-Ocean 40(4): 389-403.

Fassnacht SR, Sukh T, Fernández-Giménez M, Batbuyan B, Venable NBH, Laituri M, Adyabadam G (2011) Local understanding of hydro-climatic changes in Mongolia. Cold Region Hydrology in a Changing Climate (Proceedings of symposium H02 held during IUGG2011 in Melbourne, Australia, July 2011), IAHS 346: 120-129.

Fassnacht SR, Venable NBH, Khishigbayar J, Cherry ML (2013) The Probability of Precipitation as Snow Derived from Daily Air Temperature for High Elevation Areas of Colorado, United States. Cold and Mountain Region Hydrological Systems Under

Climate Change: Towards Improved Projections (Proceedings of symposium H02, IAHS-IAPSO-IASPEI Assembly, Gothenburg, Sweden, July 2013) IAHS 360: 65-70.

Feng S, Hu Q (2007) Changes in winter snowfall/precipitation ratio in the contiguous United States. Journal of Geophysical Research 112: 1-12.

Flanner MG, Zender CS (2005). Snowpack radiative heating: Influence on Tibetan Plateau climate. Geophysical Research Letters 32: L06501 [doi:10.1029/2004GL022076].

Flanner MG, Zender CS (2006). Linking snowpack microphysics and albedo evolution. Journal of Geophysical Research 111: D12208 [doi:10.1029/2005JD006834].

Flanner MG, Zender CS, Randerson JT, Rasch PJ (2007). Present day climate forcing and response from black carbon in snow. Journal of Geophysical Research 112: D11202 [doi: 10.1029/2006JD008003].

Flanner MG, Shell KM, Barlage M, Perovich DK, Tschudi MA (2011). Radiative forcing and albedo feedback from the Northern Hemisphere cryosphere between 1979 and 2008. Nature Geoscience 4, 151-155.

Gilbert RO (1987) Statistical Methods for Environmental Pollution Monitoring. Van Nostrand Reinhold, New York, NY, 336pp.

Gray DM, Landine GP (1987) Albedo model for shallow prairie snowcovers. Canadian Journal of Earth Science 24: 1760-1768.

Greuell JW, Konzelmann T (1994) Numerical modeling of the energy balance and the englacial temperature of the Greenland ice sheet: calculations for the ETH-Camp location (West Greenland, 1155ma.s.l.). Global Planetary Change 9(1-2): 91-114.

Groisman P Ya, Legates DR (1994). The Accuracy of United States Precipitation Data. Bull. Amer. Meteor. Soc., 75, 215–227.

Groisman PY, Karl TR, Knight RW (1994). Observed impact of snow cover on the heat balance and the rise of continent spring temperature. Science 263(14):198–200.

Groisman P Ya, Easterling DR, Quayle RG, Golubev VS, Krenke AN, Mikhailov A Yu (1996). Reducing biases in estimates of precipitation over the United States: Phase 3 adjustments. Journal of Geophysical Research 101, 7185.

Helsel DR Frans LM (2006) Regional Kendall test for trend. Environmental Science and Technology 40(13): 4066-4070.

Hernández-Henríquez MA, Déry SJ, Derksen C (2015) Polar amplification and elevation-dependence in trends of Northern Hemisphere snow cover extent, 1971–2014. Environmental Research Letters 10:044010 [doi: 10.1088/1748-9326/10/4/044010].

Hoover JD, Doesken N, Elder K, Laituri M, Liston GE (2014) Temporal Trend Analyses of Alpine Data Using North American Regional Reanalysis and In Situ Data: Temperature, Wind Speed, Precipitation, and Derived Blowing Snow. Journal of Applied Meteorology and Climatology 53: 676-693 [doi: 10.1175/JAMC-D-13-092.1].

Hudson PD, Miller G, Jenifer D, Hudson E (1983) Coptic Times. Track 1 on Rock for Light, PVC Records Number 8917, New Jersey.

Hughes MG, Robinson DA (1996) Historical Snow Cover Variability in the Great Plains Region of the USA: 1910 through to 1993. International Journal of Climatology 16:1005–1018 [doi: 10.1002/(SICI)1097-0088(199609)16:9<1005::AID-JOC63>3.0.CO;2-0].

Huntington TG, Hodgkins GA, Keim BD, Dudley RW (2004) Changes in the proportion of precipitation occurring as snow in New England (1949–2000). Journal of Climate 17: 2626-2636.

IPCC (2007) In Climate Change 2007: Synthesis Report. Contribution of working groups I, II and III to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change. Pachauri RK, Reisinger A. IPCC: Geneva: 104pp.

Knowles N, Dettinger MD, Cayan DR (2006) Trends in snowfall versus rainfall in the Western United States. Journal of Climate 19: 4545-4559.

Langlois A, Bergeron J, Brown R, Royer A, Harvey R, Roy A, Wang L, Thériault, N (2014). Evaluation of CLASS 2.7 and 3.5 simulations of snow properties from the Canadian Regional Climate Model (CRCM4) over Québec, Canada. Journal of Hydrometeorology 15: 1325-1343, [doi: http://dx.doi.org/10.1175/JHM-D-13-055.1].

Livneh B, Xia Y, Mitchell KE, Ek MB, Lettenmaier DP (2010). Noah LSM Snow Model Diagnostics and Enhancements. Journal of Hydrometeorology 11: 721–738 [doi: http://dx.doi.org/10.1175/2009JHM1174.1].

McCabe G, Wolock DM (2010) Long-term variability in Northern Hemisphere snow cover and association with warmer winters. Climatic Change 99(1-2): 141-153.

Menne MJ, Williams CN Jr, Vose RS (2015). United States Historical Climatology Network (USHCN) Daily Dataset. National Climatic Data Center, National Oceanic and Atmospheric Administration. Available at <http://cdiac.ornl.gov/epubs/ndp/ushcn/ushcn.html>, last accessed 2015-11-13.

Mote P, Hamlet AF, Clark MP, Lettenmaier DP (2005) Declining mountain snowpack in Western North America. Bulletin of the American Meteorological Society 86: 39-49.

Oleson KW, Lawrence DM, Bonan GB, Flanner MG, Kluzek E, Lawrence PJ, Levis S, Swenson SC, Thornton PE, Dai A, Decker M, Dickinson R, Feddema J, Heald CL, Hoffman F, Lamarque J-F, Mahowald N, Niu G-Y, Qian T, Randerson J, Running S, Sakaguchi K, Slater A, Stockli R, Wang A, Yang Z-L, Zeng Xi, Zeng Xu (2010). Technical Description of version 4.0 of the Community Land Model (CLM). NCAR Technical Note NCAR/TN-478+STR, National Center for Atmospheric Research, Boulder, CO, 257 pp.

Painter TH, Barrett AP, Landry CC, Neff JC, Cassidy MP, Lawrence CR, McBride KE, Farmer GL (2007). Impact of disturbed desert soils on duration of mountain snow cover. Geophysical Research Letters 34: L12502 [doi:10.1029/ 2007GL030284].

Painter TH, Deems JS, Belnap J, Hamlet AF, Landry CC, Udall B (2010). Response of Colorado River runoff to dust radiative forcing in snow. Proceedings of the National Academy of Sciences 107(40): 17125-17130 [doi:10.1073/pnas.0913139107].

Peel MC, Finlayson BL, McMahon TA (2007) Updated world map of the Köppen-Geiger climate classification. Hydrology and Earth System Science 11: 1633-1644 [doi:10.5194/hess-11-1633-2007].

Peng S, Piao S, Ciais P, Friedlingstein P, Zhou L, Wang T (2013). Change in snow phenology and its potential feedback to temperature in the Northern Hemisphere over the last three decades. Environmental Research Letters 8, 014008.

Peterson TC, Easterling DR, Karl TR, Groisman P, Nicholls N, Plummer N, Torok S, Auer I, Boehm R, Gullett D, Vincent L, Heino R, Tuomenvirta H, Mestre O, Szentimrey T, Salinger J, Førland EJ, Hanssen-Bauer I, Alexandersson H, Jones P, Parker D (1998). Homogeneity adjustments of in situ atmospheric climate data: a review. Int. J. Climatol., 18: 1493–1517.

Pielke RA Sr, Stohlgren T, Schell L, Parton W, Doesken N, Redmond K (2002) Problems in evaluating regional and local trends in temperature: an example from eastern Colorado, USA. International Journal of Climatology 22(4): 421-434.

Qu X, Hall A (2006). Assessing snow albedo feedback in simulated climate change. Journal of Climate 19: 2617-2630 [doi: http://dx.doi.org/10.1175/JCLI3750.1].

Qu X, Hall A (2007). What Controls the Strength of Snow-Albedo Feedback? Journal of Climate 20(15): 3971-3981.

Reek T, Doty SR, Owen TW (1992) A Deterministic Approach to the Validation of Historical Daily Temperature and Precipitation Data from the Cooperative Network. Bulletin of the American Meteorological Society 73: 753-762 [doi: 10.1175/1520-0477]

Salmi T, Määttä A, Anttila P, Ruoho-Airola T, Amnell T (2002) Detecting trends of annual values of atmospheric pollutants by the Mann-Kendall Test and Sen's Slope Estimates. Finnish Meteorological Institute, Helsinki, Publications on Air Quality #31.

Stewart IT (2009) Changes in snowpack and snowmelt runoff for key mountain regions. Hydrological Processes 23: 78-94.

US Army Corps of Engineers (1956) Snow Hydrology: Summary Report of the Snow Investigations. North Pacific Division, Portland, OR, 437 pp. (Available through the US Department of Commerce, Washington, DC).

Venable NBH, Fassnacht SR, Adyabadam G, Tumenjargal S, Fernández-Giménez M, Batbuyan B (2012) Does the Length of Station Record Influence the Warming Trend That is Perceived by Mongolian Herders near the Khangai Mountains? Pirineos 167: 71-88 [doi:10.3989/Pirineos.2012.167004].

Verseghy DL (1991) CLASS - A Canadian Land Surface Scheme for GCMs: I. Soil model. International Journal of Climatology 11: 111-133.

Vionnet V, Brun E, Morin S, Boone A, Faroux S, Le Moigne P, Martin E, Willemet J-M (2012) The detailed snowpack scheme Crocus and its implementation in SURFEX v7.2. Geoscientific Model Development 5: 773-791 [doi:10.5194/gmd-5-773-2012].

Wang A, Zeng X (2009). Improving the treatment of vertical snow burial fraction over short vegetation in the NCAR CLM3. Adv. Atmos. Sci. 26:877-886 [DOI:10.1007/s00376-009-8098-3].

Yang D, Goodison BE, Metcalfe JR, Golubev VS, Bates R, Pangburn T, Hanson CL (1998) Accuracy of NWS 8” standard nonrecording precipitation gauge: Results and application of WMO intercomparison. Journal of Atmospheric and Oceanic Technology 15: 54-68.

List of Figures

1. Location map and average climate from 1951 through 2010 for the 20 study stations in the Northern Great Plains. The annual climate summary includes the temperature (temp) as daily maximum (max) and minimum (min), precipitation (precip) as the sum of rain and snow shown separately, days with precipitation (days with P) as the sum of days with rain and days with snowfall shown separately, and the winter albedo. The precipitation as snow is the amount of precipitation when fresh snow was observed, as defined by Huntington *et al*. (2004); the days with snowfall are the days when fresh snow was observed. The winter albedo is modeled. The nine stations that are part of the U.S. Historical Climate Network (HCN) are identified with an asterisk.

2. Significant climatic trends at the 20 stations and for the seven variables summarized in Figure 1 for the period from 1951 through 2010 shown per century. Significant trends are presented at the $p<0.05$ level, except for trends shown with a dashed border which are at the $p<0.10$ level. Trends from the Regional Kendall Test (RKT) are shown at the right. The Sterling, Colorado (yellow) and Kimball, Nebraska (orange) stations are highlighted as they are subsequently compared in Figures 4 and 5. The nine stations that are part of the U.S. Historical Climate Network (HCN) are identified with an asterisk.

3. The spatial distribution of the climatic trends at the $p<0.05$ level for the period from 1951 through 2010 shown per century. Trends shown with a dashed border are at the $p<0.10$ level. Significant RKT trends are shown in the top left. Sterling and Kimball stations are highlighted.

4. Time series of climate at the neighbouring i) Sterling, Colorado and ii) Kimball, Nebraska meteorological stations, illustrating a) temperature (annual maximum and minimum), b) total annual precipitation and precipitation as snow, c) days with precipitation and days with snowfall, and d) winter albedo. Significant trends are shown as solid lines for the $p<0.05$ level and as dashed lines for the $p<0.10$ level. The Kimball station is 73 km northwest of the Sterling station.

5. Significant rates of change for the seven variables (a-g) for the two focus stations over varying lengths of record (60 years: 1951-2010 and 30 years: 1951-1980, 1981-2010). There were no significant precipitation trends (c).

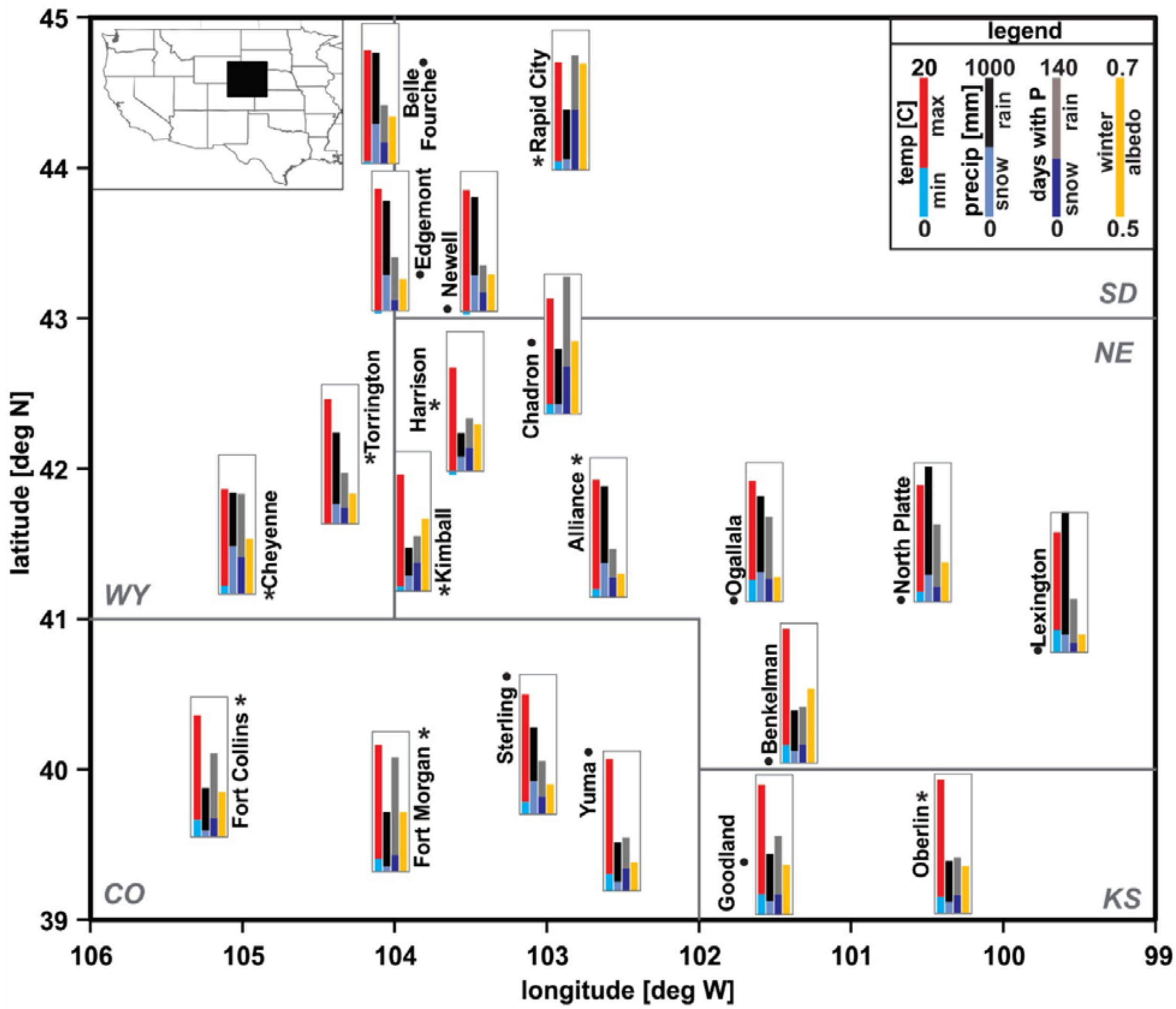

Figure 1. Location map and average climate from 1951 through 2010 for the 20 study stations in the Northern Great Plains. The annual climate summary includes the temperature (temp) as daily maximum (max) and minimum (min), precipitation (precip) as the sum of rain and snow shown separately, days with precipitation (days with P) as the sum of days with rain and days with snowfall shown separately, and the winter albedo. The precipitation as snow is the amount of precipitation when fresh snow was observed, as defined by Huntington *et al*. (2004); the days with snowfall are the days when fresh snow was observed. The winter albedo is modeled. The nine stations that are part of the U.S. Historical Climate Network (HCN) are identified with an asterisk.

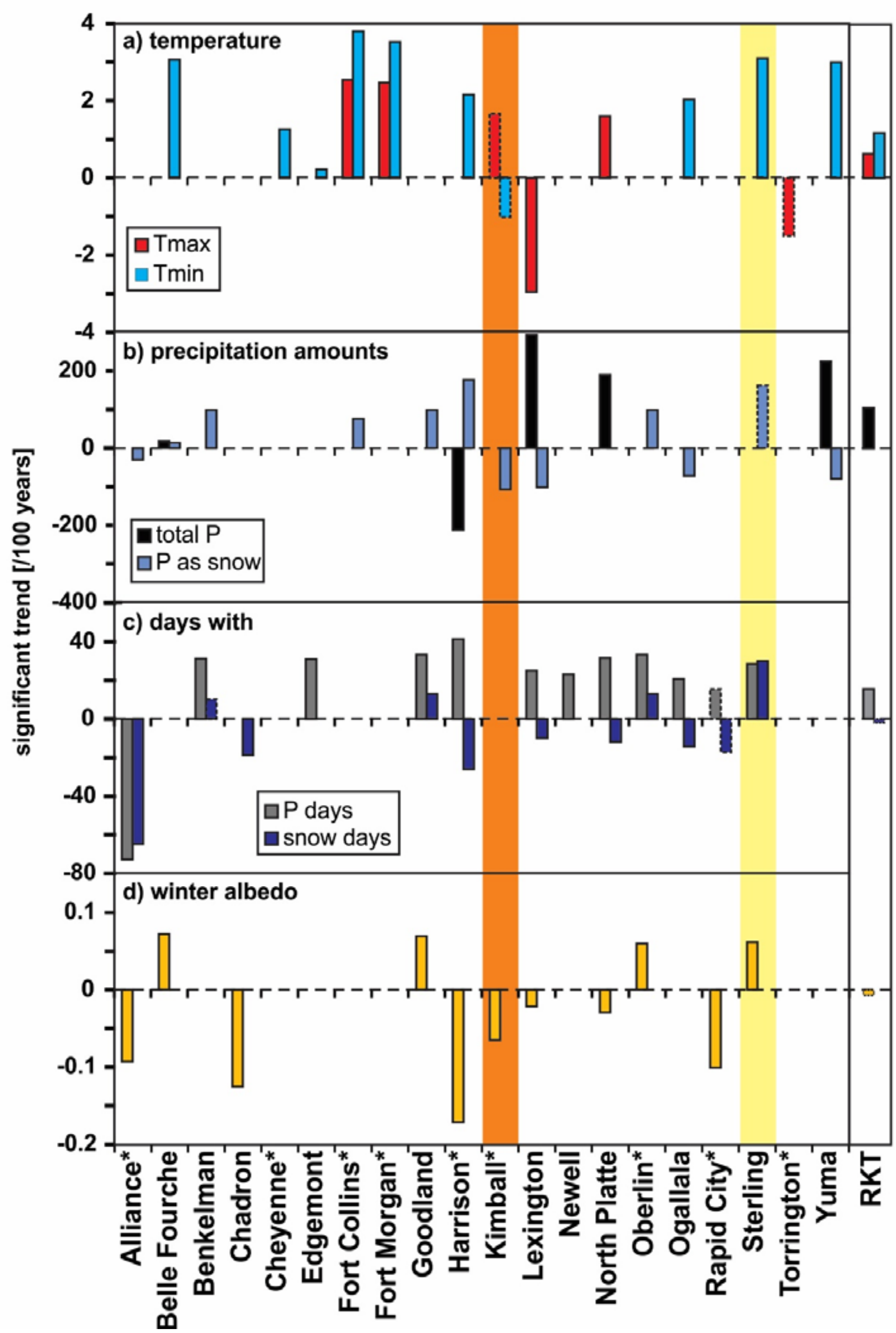

Figure 2. Significant climatic trends at the 20 stations and for the seven variables summarized in Figure 1 for the period from 1951 through 2010 are shown per century. Significant trends are presented at the $p<0.05$ level, except for trends shown with a dashed border, which are at the $p<0.10$ level. Trends from the Regional Kendall Test (RKT) are shown at the right. The Sterling, Colorado (yellow) and Kimball, Nebraska (orange) stations are highlighted as they are subsequently compared in Figures 4 and 5. The nine stations that are part of the U.S. Historical Climate Network (HCN) are identified with an asterisk.

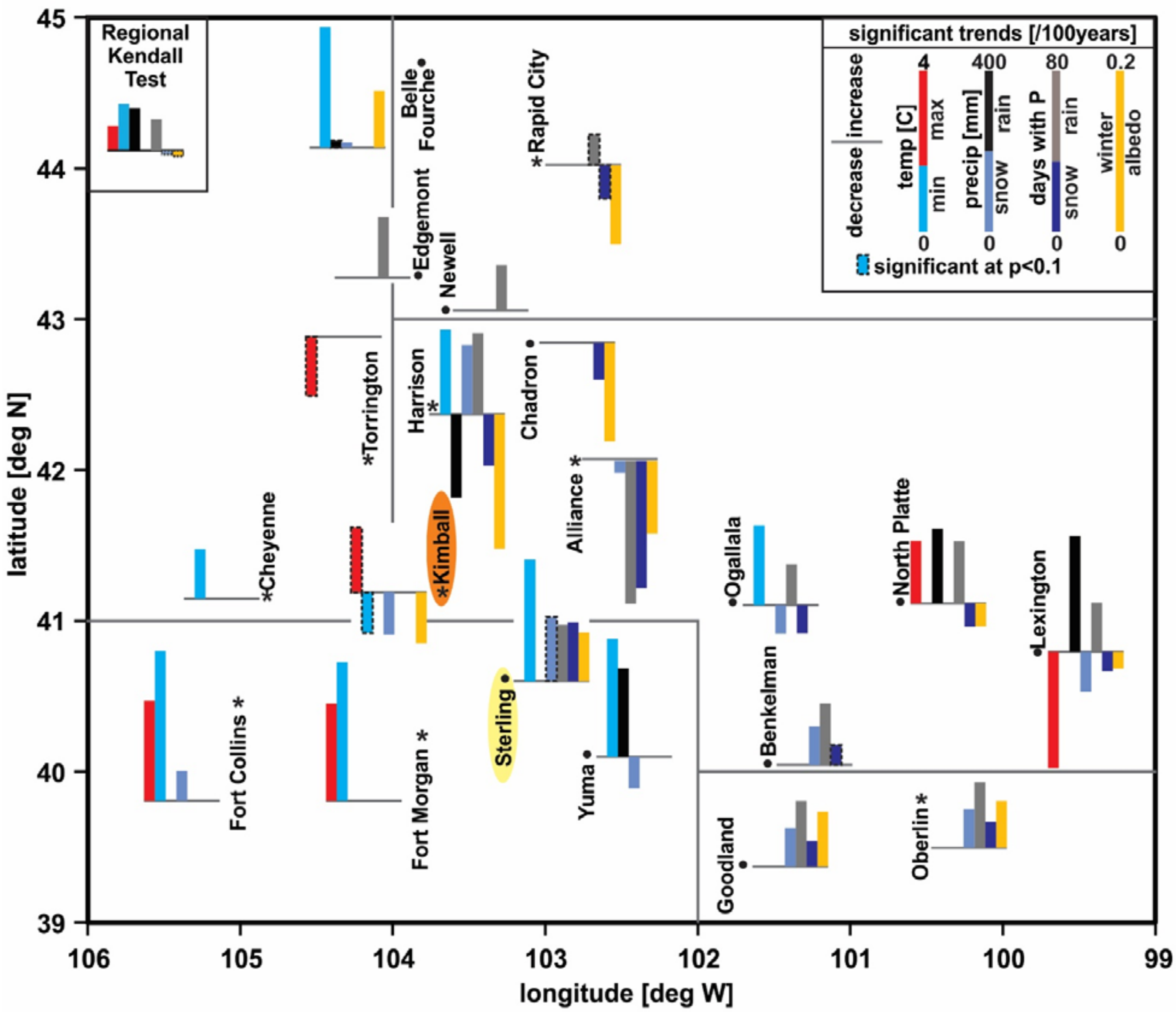

Figure 3. The spatial distribution of the climatic trends at the p<0.05 level for the period from 1951 through 2010 are shown per century. Trends shown with a dashed border are at the p<0.10 level. Significant RKT trends are shown in the top left. Sterling and Kimball stations are highlighted.

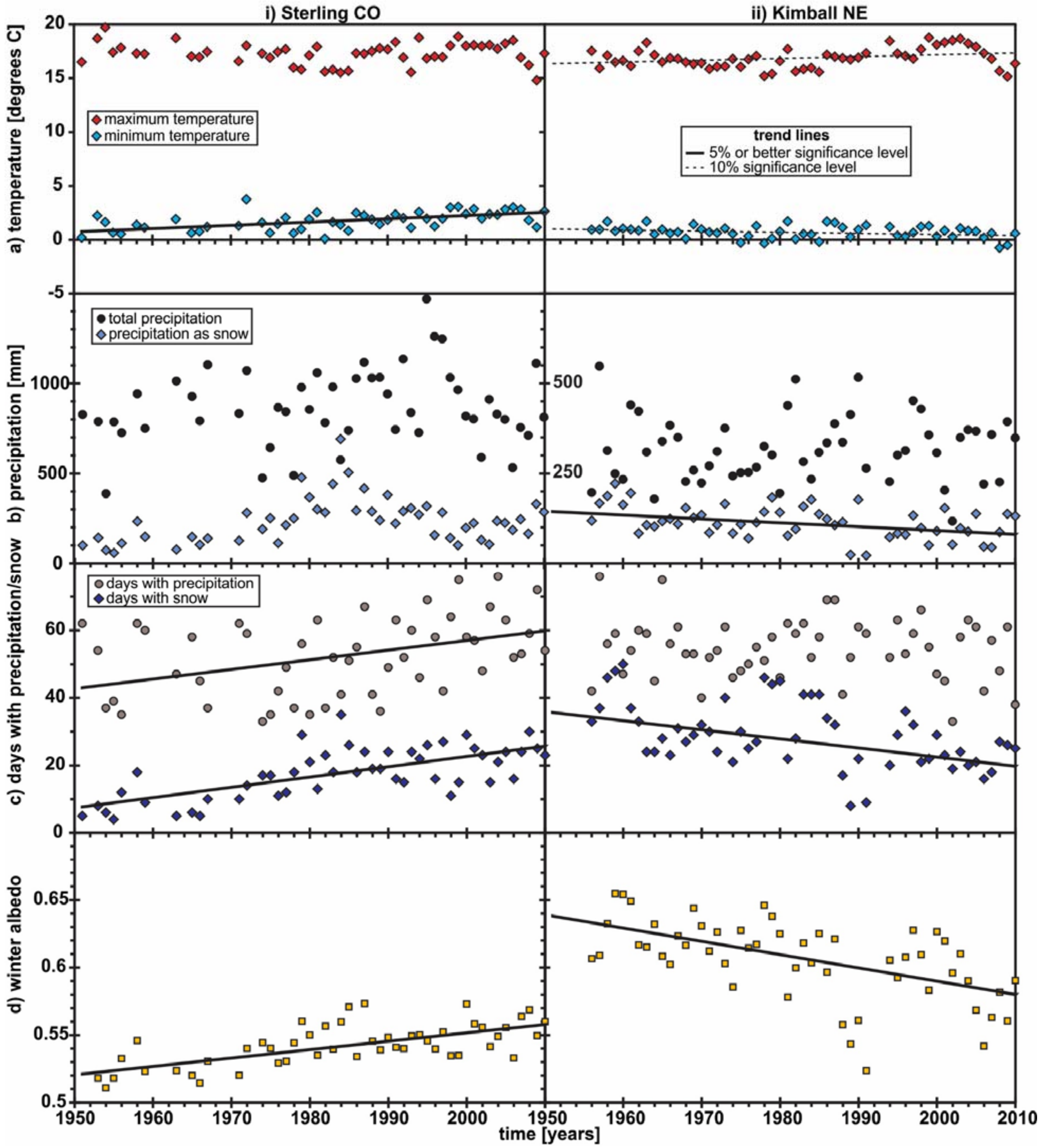

Figure 4. Time series of climate at the neighbouring i) Sterling, Colorado and ii) Kimball, Nebraska meteorological stations, illustrating a) temperature (annual maximum and minimum), b) total annual precipitation and precipitation as snow, c) days with precipitation and days with snowfall, and d) winter albedo. Significant trends are shown as solid lines for the $p<0.05$ level and as dashed lines for the $p<0.10$ level. The Kimball station is 73 km northwest of the Sterling station.

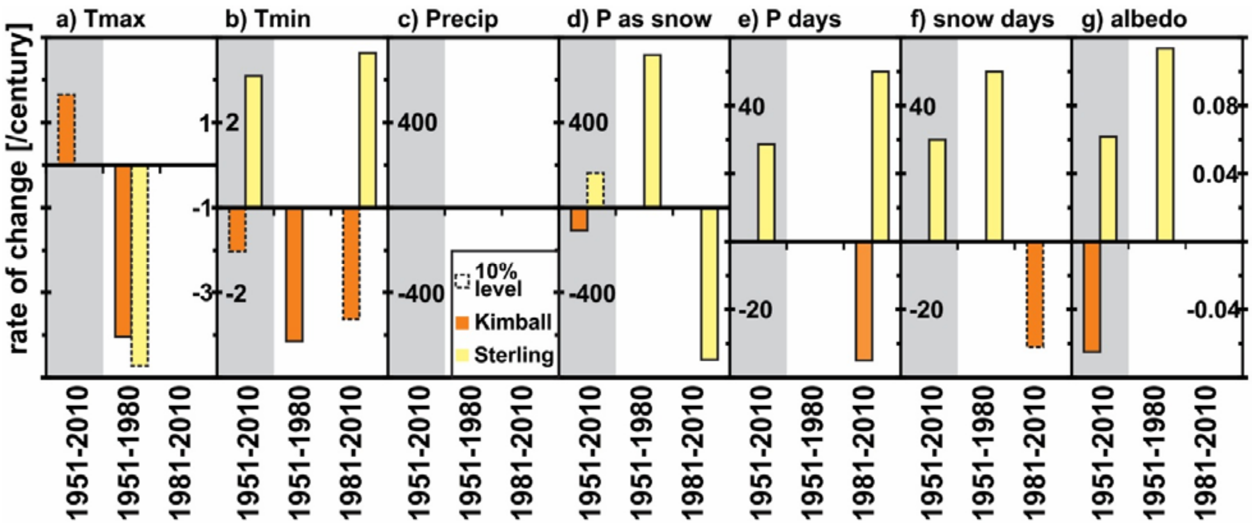

Figure 5. Significant rates of change for the seven variables (a-g) for the two focus stations over varying lengths of record (60 years: 1951-2010 and 30 years: 1951-1980, 1981-2010). There were no significant precipitation trends (c).