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The impact of clouds, land use and snow cover on climate in the Canadian Prairies

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Abstract. This study uses 55 years of hourly observations of air temperature, relative humidity, daily precipitation, snow cover and cloud cover from 15 climate stations across the Canadian Prairies to analyze biosphere-atmosphere interactions. We will provide examples of the coupling between climate, snow cover, clouds, and land use. Snow cover acts as a fast climate switch. With the first snow fall, air temperature falls by 10 °C, and a similar increase in temperature occurs with snow melt. Climatologically, days with snow cover are 10 °C cooler than days with no snow cover in Alberta. However the interannual variability has a larger range, so that for every 10 % decrease in days with snow cover, the mean October to April climate is warmer by 1.4 to 1.5 °C. Snow cover also transforms the coupling between clouds and the diurnal cycle of air temperature from a boundary layer regime dominated by shortwave cloud forcing in the warm season to one dominated by longwave cloud forcing with snow cover. Changing agricultural land use in the past thirty years, specifically the reduction of summer fallowing, has cooled and moistened the growing season climate and increased summer precipitation. These hourly climate data provide a solid observational basis for understanding land surface coupling, which can be used to improve the representation of clouds and land-surface processes in atmospheric models.

1 Introduction

This paper reviews the analysis of the diurnal climatology of the Canadian Prairies (Betts et al., 2013a, b, 2014a, b, 2015; Betts and Tawfik, 2016). There are 15 climate stations with more than 55 years of hourly temperature, relative humidity, wind, precipitation and snow depth, as well as a unique set of hourly estimates of opaque, reflective cloud cover, made by trained observers (Environment Canada, 2013). We have used snow cover and the daily mean opaque cloud cover, $\mathrm{OPAQ_m}$, to stratify this large dataset (660 station-years), and study the diurnal and seasonal climate. The land surface, the boundary layer (BL) and the overlying atmosphere are a tightly coupled system, with two distinct climates above and below the freezing point of water. Below freezing, precipitation falls as snow and the highly reflective snow cover acts as a fast climate switch, which drops the mean surface temperature by 10 °C within days, and changes the climatology (Sect. 2). One result is the mean cold season temperature is linearly related to the fraction of days with snow cover. Another impact of snow cover is to transform the coupling of the diurnal cycle to cloud cover (Sect. 3). There have been large land use changes on the Prairies over the past 30 years, as 5 million hectares of summer fallow has been converted to continuous cropping. Increased transpiration has cooled and moistened the growing season climate and increased precipitation (Sect. 4). These observational studies provide a baseline for model evaluation.

Figure 1 shows the location of the climate stations, Canadian ecozones, regional zones, agricultural regions and boreal forest, with 50 km radius circles around each station. We generated daily means for all variables, such as mean temperature and humidity, T_m and $\mathrm{RH_m}$ and opaque cloud, $\mathrm{OPAQ_m}$, and merged daily total precipitation and daily snow depth. From the hourly data we computed the diurnal temperature range between maximum temperature, T_x, and minimum temperature, T_n, as

$$\mathrm{DTR} = T_x - T_n. \tag{1}$$

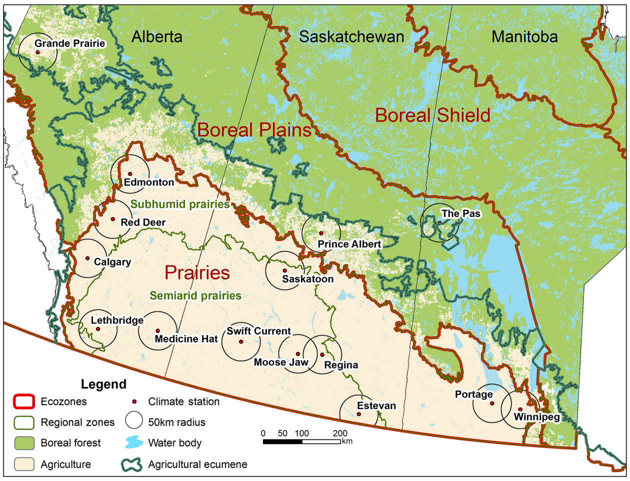

Figure 1. Climate station locations, Canadian ecozones, regional zones, agricultural regions and boreal forest.

At the time of maximum temperature, T_x, typically in the afternoon in the warm season, we derived the mixing ratio, Q_{tx}, and the corresponding equivalent potential temperature and pressure-height to the lifting condensation level (LCL), which is close to cloud-base in the warm season (Betts et al., 2013a).

2 Impact of snow cover on climate

Snow/ice albedo feedback is well-known on global scales, but global models show a wide variation in their representation of this key process (Qu and Hall, 2007; Bony et al., 2006; Xu and Dirmeyer, 2013). Early case studies showed that snow cover reduces surface temperatures by about 5 °C on both the short-term timescale and on monthly timescales (Namias, 1960, 1985; Cohen and Rind, 1991; Dewey, 1977; Wagner, 1973). However there has been surprisingly little quantitative observational analysis of the impact of snow on local and regional climate. Betts et al. (2015) found that there are two very distinct Prairie climates, sharply separated by the freezing point of water: one for the cold season with $T_\mathrm{m} < 0$ °C and surface snow cover, and one for the warm season with $T_\mathrm{m} > 0$ °C with no snow cover.

Figure 2 shows the annual mean climatology (black) and the partition for each month into the days with no snow cover (red) and with snow cover (blue). The mean climatology is the weighted mean of the no-snow and snow values, using the number of days in each class as weights. We show three stations in Alberta, which have an average of 36 % of the days in winter with no snow cover. For October to April, we plot the temperature difference ΔT between these snow and no-snow climatologies. There is a clear separation of the climatologies for these cold season months for all three stations with a mean value of

$$\Delta T = -9.8(\pm 0.8)^\circ\mathrm{C}.$$

That is, it is climatologically about 10 °C cooler when there is reflective snow cover with a surface albedo ≈ 0.7 (Betts et al., 2014b). Other stations across the Prairies show a similar cooling of the climatology with snow, although some have too few snow-free days in January and February to make this comparison.

Figure 3 shows the temporal perspective (adapted from Betts et al., 2014b): the climatological transition across the first lasting snow in the fall (left panel) and the final melt of the snowpack in the spring (center panel). The left panel is a composite across the fall transitions from no snow to lasting snow cover for 1955–2004 for six stations in Saskatchewan. The average date of this first snowfall is 15 November, and there is a 10 °C drop of mean temperature within less than a week, as the surface albedo changes from about 0.2 to 0.7 with snow. From a climatological perspective this means that the transition into winter is not a smooth process lasting 4–6 weeks, but it typically occurs in less than a week with the first lasting snow fall. In spring there is a longer reverse transition with a rise of 10 °C in about 10 days as the snowpack melts.

Betts et al. (2014b) described the role of reflective snow cover as a fast climate switch which drops the surface temperature by about 10 °C, as seen in both Figs. 2 and 3. Conceptually, temperatures fall as the solar zenith angle increases with the approach of winter, until precipitation falls as snow, once temperatures fall below freezing in a favorable synoptic situation. With snow cover, two feedbacks cool the surface making it much harder for transient synoptic advection to melt the snow cover. The primary one is the reflection of sunlight by snow, but Betts et al. (2014b) showed that a drop in the incoming long wave radiation is also significant in cooling the surface. Snowmelt in spring is initiated by decreasing solar zenith angle driving warmer temperatures, and with the loss of the feedbacks associated with snow cover, temperatures rise about 10 °C. Snow cover has a further feedback, as it insulates the ground, reducing the ground heat flux, which is typically upward in fall and downward in spring. We will see in the next section that there is also a profound change in the coupling between cloud cover and surface temperature with snow, which may also play a role in the longwave coupling.

In Alberta in the lee of the Rocky Mountains the winter snowpack is often transient, so there is a wide interannual variation in the number of days with snow cover. The right panel plots mean temperature for the period October to April, against the fraction of days with snow cover for five Alberta stations, each with almost fifty years of data. There is a strong linear dependence (with $R^2 = 0.79$) of the interannual variation in the cold season mean temperature on the fraction of days with snow cover. The temperature range of 14.6 ± 0.6 °C from zero to 100 % snow cover is larger than

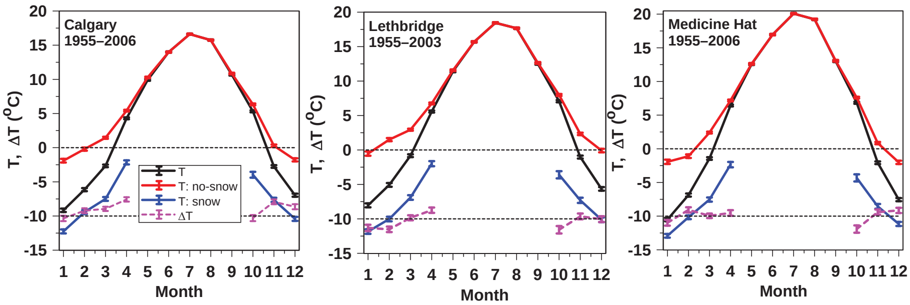

Figure 2. Annual Climatology for Calgary, Lethbridge and Medicine Hat, with the separation into days with and without snow cover.

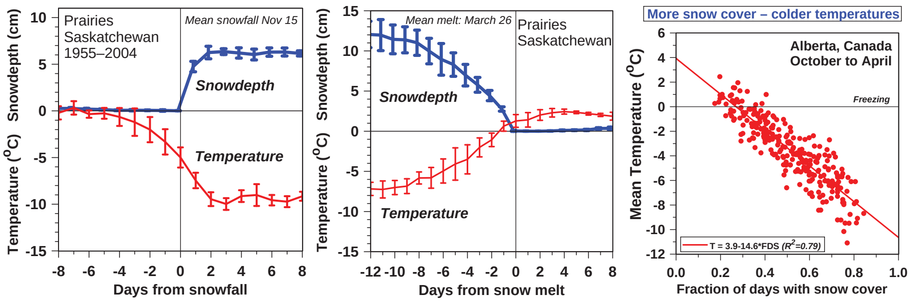

Figure 3. Fall (left panel) and spring (center panel) snow transition for six Saskatchewan stations, and (right panel) relation of cold season temperature to fraction of days with snow cover for five Alberta stations.

the local change with snow cover of order 10 °C. This apparent amplification suggests that there might be a further positive feedback between the local impact of the snowpack and the larger-scale regional climate.

3 Impact of snow cover and cloud cover on the diurnal cycle

The coupling between cloud cover, temperature and the diurnal cycle is a central feature of the Earth's climate (Betts, 2015). However, the representation in global models is a challenge because of uncertainties in their representation of clouds (Groisman et al., 2000). So observational studies play a key role (Dai et al., 1999). The long time-series of opaque reflective cloud data at climate stations on the Canadian Prairies used for our analyses has been transformative, because it can be calibrated against baseline surface radiation measurements to give the longwave and shortwave cloud forcing (Betts et al., 2015).

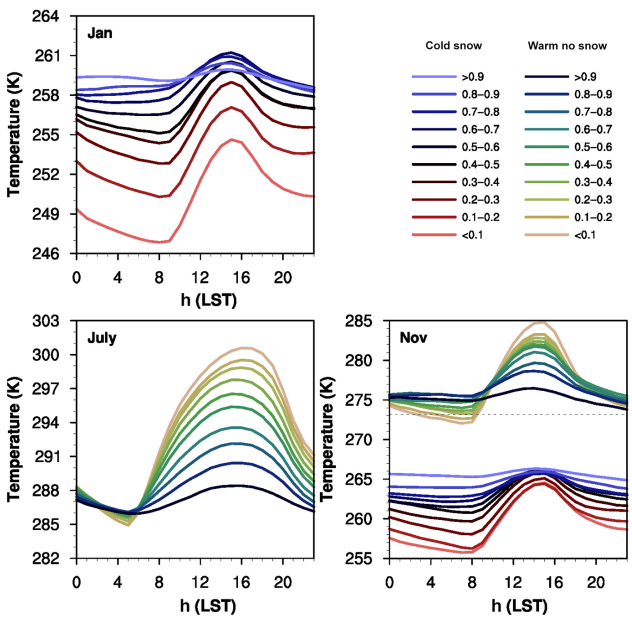

Figure 4. Diurnal cycle of temperature stratified by opaque cloud for January, July and November (adapted from Betts and Tawfik, 2016).

Figure 4 shows the climatology of the diurnal cycle for January, July and November stratified by opaque cloud fraction and by surface snow cover for all the Canadian Prairie data (adapted from Betts and Tawfik, 2016). In January, which is representative of all the winter months with snow (not shown), mean air temperature increases with opaque cloud cover, and sunrise minimum temperatures plunge under clear skies. The amplitude of the diurnal cycle increases as cloud cover decreases in response to the daytime solar forcing. This winter stable boundary layer regime is dominated by long-wave cloud forcing (Betts et al., 2015), for which cloud cover reduces the cooling of the surface to

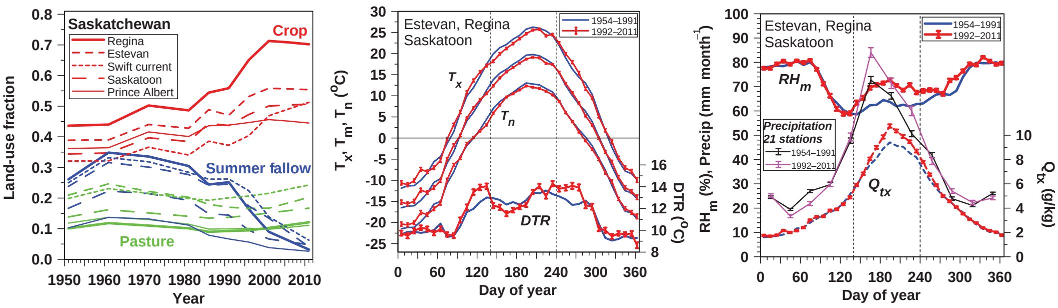

Figure 5. Long term trends in total cropland, pasture, and summer fallow around five climate stations in Saskatchewan (left panel); mean changes in annual cycle of DTR, T_x, T_m and T_n for Saskatoon, Regina and Estevan (center panel) and $\mathrm{RH_m}$, Q_{tx} and mean precipitation for 21 stations in southern Saskatchewan (right panel).

space. In contrast, in July and indeed all months with no snow cover (not shown), minimum temperatures vary little with cloud cover, and the diurnal range to the afternoon maximum is largest under clear skies. This is because in the warm season, the shortwave reflection by clouds dominates (Betts et al., 2015), so that mean temperatures are warmest under clear skies. This is characteristic of an unstable daytime convective BL (Betts et al., 2013a). For November (as well as March and April, not shown), there is sufficient data to show both climatologies, with and without snow, and the clear separation in temperature between them that increases as cloud cover falls (Betts and Tawfik, 2016). We see that snow cover completely transforms the coupling between opaque cloud and the diurnal cycle of temperature.

4 Impact of land use on climate

In the past thirty years there has been a major change in land use across the Canadian Prairies, specifically the conversion of more than five million hectares of summer fallow to continuous cropping. The large increase in the area of cropland has increased summer transpiration, which has reduced maximum temperatures in the growing season over the Prairies (Gameda et al., 2007). Other analyses of US Midwest summer temperature maxima also show a cooling from land-use change to cropland (Bonan, 2001) and cropland intensification (Mueller et al., 2016).

Figure 5 summarizes the long-term climate impact of the reduction of summer fallow in Saskatchewan (Betts et al., 2013b). The left panel shows the land-use trends in total cropland, pasture, and summer fallow around five climate stations. The 50 km radius circles around each station in Fig. 1 were used to generate local averages of the ecodistrict crop data. We split the climate station timeseries into two periods: a longer historic period, 1954–1991, when summer fallow cover was large (although slowly decreasing) and a recent 20-year period, 1992–2011, when summer fallow has fallen rapidly to its present low value. We did not attempt any analysis of decadal trends.

Saskatoon, Regina and Estevan in Saskatchewan have complete datasets and show similar changes between the two time periods, so we averaged them as 10-day means (Betts et al., 2013b). The center panel shows the mean changes in the annual cycle of DTR, T_x, T_m and T_n and the right panel the mean changes in $\mathrm{RH_m}$ and mixing ratio Q_{tx} at the time of the afternoon maximum temperature. We show the standard errors of the difference between the two mean time series as an indication of significance. Precipitation has much more variability than temperature and humidity, so we used the 21 stations in Saskatchewan south of 53.22° N in the second generation adjusted precipitation dataset (Mekis and Vincent, 2011). The right panel shows the difference in the monthly mean time-series of precipitation for the two time periods, with the standard error of each monthly mean.

It is clear that there are significant changes in the growing season climate between the historic period, 1953–1991, and the more recent period since 1992. Betts et al. (2013b) considered the period $140 \leq \mathrm{DOY} < 240$ (20 May–27 August) to be representative of the growing season (where Day of Year is abbreviated DOY). Since 1992, the growing season is significantly cooler, with a drop of (T_x, T_m, T_n) of (-0.93 ± 0.09, -0.82 ± 0.07, -0.68 ± 0.06 °C), and significantly moister with a rise of ($\mathrm{RH_m}$, Q_{tx}) of (6.9 ± 0.2 %, $0.70 \pm 0.04\,\mathrm{g\,kg^{-1}}$). There is a corresponding fall of the LCL of 22.3 ± 1.1 hPa, and a small rise of equivalent potential temperature of 1.1 ± 0.2 K, at the time of afternoon maximum temperature. There is also an increase of summer (June, July and August) precipitation of 25.9 ± 4.6 mm. The change of DTR in the growing season has a more complex structure with a decrease in the early part of the growing season ($140 \leq \mathrm{DOY} < 200$) of -0.60 ± 0.09°C that is coupled to an increase in cloud cover (Betts et al., 2013b).

We conclude that more intensive agriculture has increased transpiration, which has cooled and moistened the growing season climate with a lowered cloud-base, increased equivalent potential temperature, and an increase of summer precipitation.

5 Conclusions

The long time-series of hourly climate data including opaque cloud measurements provides insight into the fully coupled land–atmosphere–cloud interactions in the Canadian Prairies. Reflective surface snow cover acts as a climate switch, since surface albedo changes from around 0.2 to 0.7. The climatology with and without snow cover shows that near-surface temperature drops by 10 °C with snow cover, and the coupling between opaque cloud and the diurnal cycle of temperature is transformed. During the warm season, the shortwave reflection by clouds dominates and temperatures are coolest under cloudy skies. Minimum temperatures vary little with cloud cover and the diurnal range to the afternoon maximum is largest under clear skies. The reverse is true during the cold season with snow cover, when the longwave warming by clouds dominates. The near-surface air is warmest under cloudy skies, while sunrise minimum temperatures plunge under clear skies. This snow-albedo coupling is so strong and rapid that the mean temperatures of the cold season are linearly related to the fraction of days with snow cover. Increased transpiration related to the shift to more intensive agriculture over the past 30 years, from summer fallow to continuous cropping on 5 million hectares of the agricultural land, has cooled and moistened the growing season climate, and increased precipitation. The on-going analyses of these data are providing new understanding of the climate and land–atmosphere–cloud system for northern latitudes. We expect that these observational data sets will also be extremely valuable for testing and evaluating atmospheric models.

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