Global and Planetary Change 146 (2016) 10–21

Contents lists available at ScienceDirect

Global and Planetary Change

journal homepage: www.elsevier.com/locate/gloplacha

Impact of climate warming on snow processes in Ny-Ålesund, a polar maritime site at Svalbard

J.I. López-Moreno [a,*], J. Boike [b], A. Sanchez-Lorenzo [a], J.W. Pomeroy [c]

[a] Instituto Pirenaico de Ecología, CSIC, Zaragoza, Spain
[b] Alfred Wegener Institute, Helmholtz Center for Polar and Marine Research, Telegrafenberg A43, 14473 Potsdam, Germany
[c] University of Saskatchewan, Saskatoon, Canada

ARTICLE INFO

Article history:
Received 14 March 2015
Received in revised form 9 September 2016
Accepted 14 September 2016
Available online 15 September 2016

Keywords:
Snow
Sensitivity
Climate warming
Cold regions hydrological model (CRHM)
Arctic
Svalbard

ABSTRACT

The impact of observed changes in air temperature and precipitation from 1969 to 2013 and climate projections for 2050 and 2080 at Ny-Ålesund, an arctic research station on Spitzbergen Island in the Svalbard Archipelago on snow hydrological processes, were analyzed using snow accumulation and ablation algorithms in the physically based Cold Regions Hydrological Modelling platform (CRHM). The climate projections were obtained from phase 5 of the Coupled Model Intercomparison Project (CMIP5), with a focus on the snow-dominated period (October to June). To identify the potential effects of increasing temperature and precipitation, a model sensitivity analysis (1 °C to 5 °C), with and without a 25% increase in precipitation, was run on CRHM snow processes. The results indicated that the greatest observed warming was during the early snow season (October–February), with increases of 0.8 and 0.9 °C decade^{-1} for maximum (Tmax) and minimum (Tmin) temperatures, respectively. There was also a significant increase in annual and winter precipitation (24 mm decade^{-1}). The late snow season (March–June) also had a marked increase in temperature (0.5 and 0.69 °C decade^{-1} for Tmax and Tmin respectively), but no significant change in precipitation. These changes lead to a significant increase in the number of days with rainfall rather than snowfall. The sensitivity analysis indicated that mean snow water equivalent snowpack will decrease by 10.2% (early snow season) and 11.1% (late snow season) per degree of increased air temperature. For each degree of temperature increase, the modelled peak snow-water-equivalent (SWE) declined by 6.9%, duration of snowpack declined 11 days, and the number of days with rain increased 43% for the early snow season and 12.8% for the late snow season. A warmer climate also leads to markedly decreased surface snow sublimation and the fraction of snowfall eroded and transported by blowing snow. For most snowpack parameters analyzed, the response to warming accelerates with increased warming, especially above 3 °C. A 25% increase in precipitation partially counteracted the response to warming, with the greatest effect on peak SWE.

1. Introduction

Most studies of the arctic climate have reported increased precipitation and temperature in recent decades, and that this region has warmed more than any other global region in the last three decades (Zhang et al., 2008; Graversen et al., 2008; Screen and Simmonds, 2010; Maturilli et al., 2013). This “arctic amplification” is mostly explained by modification of air mass circulation, decreases in snow and sea ice albedo, and increased radiative forcing (Serreze et al., 2000; Langen and Alexeev, 2007; Miller et al., 2010).

These changes also occurred in Ny-Ålesund, an arctic research station on Spitzbergen Island in the Svalbard Archipelago of the Norwegian Sea (Fig. 1). Snow dominates the landscape of Svalbard during 8–10 months of the year (Winther et al., 2003). By maintaining a surface temperature at 0 °C or below these surfaces have a high albedo, providing low thermal conductivity, storing water and covering surface features (Pomeroy and Brun, 2001). This snow affects local climate, large scale atmospheric circulation, permafrost regime, mass balance of glaciers, avalanches, ground transportation, and the phenology of plants and animals and the annual carbon budget (Boike et al., 2003; Eckerstorfer and Christiansen, 2011; Callaghan et al., 2011a, 2011b; Derksen and Brown, 2012; Lüers et al., 2014). The Svalbard airport and Ny-Ålesund have some of the best climatic and meteorological records in the Arctic (Førland et al., 1997). In both locations, the changes in climatic variables depend on the study period, but the marked recent increases in temperature cannot be simply explained by shifts in atmospheric circulation patterns (Førland and Hanssen-Bauer, 2000; Nordli et al., 2014; Maturilli et al., 2014). This warming trend also occurred in permafrost, and was even detected at a depth of 60 m (Isaksen et al., 2007). At the permafrost borehole in Janssonhaugen, reconstructed

* Corresponding author.
E-mail address: nlopez@ipe.csic.es (J.I. López-Moreno).

http://dx.doi.org/10.1016/j.gloplacha.2016.09.006
0921-8181/

Fig. 1. Location of Ny-Alesund and Bayelva station.

near-surface ground temperature warmed 1–2 °C over the last eight decades (Isaksen et al., 2000).

This region has also experienced more cloudiness, precipitation, and cyclones in recent years, especially during the wintertime (Hanssen-Bauer and Førland, 1998; Sepp and Jaagus, 2011). However, a part of the reported increases in precipitation could be explained by a progressive decrease of the precipitation undercatch due to the decreased percentage of solid precipitation caused by climate warming (Førland et al., 1997; Førland and Hanssen-Bauer, 2000). Equally, the increased cloudiness has led to increased incoming longwave radiation in Ny-Ålesund, which is the main change in the winter radiative budget in this region (Maturilli et al., 2014).

Climate simulations predict an acceleration of the trends recently observed in Svalbard for the next several decades, with a projected increase in temperature three times greater than observed during the last century. These simulations also project an increase in annual precipitation of 5–20% in southwest of Spitsbergen up to year 2100, and >40% in northeastern parts of the archipelago (Hanssen-Bauer, 2002; Haugen et al., 2000; Førland et al., 2011).

The documented climatic changes in Svalbard should have seriously impacted the duration and magnitude of the snowpack, but long-term information on the distribution and interannual evolution of the snowpack in this region is limited because of the great spatial variability due to wind redistribution by blowing snow (Førland and Hanssen-Bauer, 2000). Remote sensing data and the observed decrease in spring albedo showed that the date of complete ablation now occurs earlier in the season, and mean daily temperature below 0 °C starts later in September (Maturilli et al., 2014). Moreover, the reported increase of rain-on-snow events, even in mid-winter (Førland and Hanssen-Bauer, 2000), lead to warmer snowpack (Putkonen and Roe, 2003), which can substantially warm the permafrost soil underneath (Westermann et al., 2011) and reduce snow accumulation, as reported for the Bayelva glacier in the region of Ny-Ålesund (Nowak and Hodson, 2013).

This paper presents an updated report of the climate in Ny-Ålesund from 1969 to 2013, with a focus on the months dominated by snow cover, projects the climate of this region for the coming decades, and assesses the impact of increasing temperature on snow accumulation and snowpack duration in this region. We also pay particular attention to the effect of dominant snow processes in the region (wind redistribution and rain-on-snow events), and to parameters that determine the duration and thickness of the snowpack. Many meteorological variables affect the interannual evolution of the snowpack (López-Moreno et al., 2013), but sensitivity analysis has not yet been used to assess the effect of warmer temperatures on snow accumulation and snowpack duration in Svalbard or any other polar maritime region. Previous climate studies have used such sensitivity analysis in studies of the Washington Cascades (Casola et al., 2009), the Alps (Beniston et al., 2003), the Spanish Pyrenees (López-Moreno et al., 2013, 2014) and Yukon (Rasouli et al., 2014), and other locations of the Northern Hemisphere (Brown and Mote, 2009). Sensitivity analysis is a valuable technique that allows assessment of the sensitivity of the snowpack to different climatic parameters, and provides important information on how the snowpack responded to recent warming and how it may respond to future warming.

2. Data and methods

Ny-Ålesund (78°56′ N, 11°53′ E) was formerly a small mining town, and is now a research station on the west coast of Spitsbergen in the Svalbard Archipelago in the Norwegian Sea (Fig. 1). The area is bordered by two mountains, Zeppelinfjellet and Scheteligfjellet, between which the glacial Bayelva River originates from the two branches of the glacier Brøggerbreen. The terrain flattens out to the north of the study site, and about 1 km downstream the Bayelva River flows into the Kongsfjorden and the Arctic Ocean. In the area, sparse vegetation alternates with exposed soil and sand, or rock fields. The long-term (1969–2013) mean temperature in Ny-Alesund was −4.2 °C on an annual basis. From October to February, when there is little or no daylight (hereafter called "early snow season"), mean temperature is −9.2 °C. From March to June, when the sunshine dominates or lasts all day (hereafter called "late snow season"), mean temperature is −1.4 °C. The average annual precipitation was 415.5 mm, with 198.7 mm during the early snow season, 81.8 mm during the late snow season, and the rest from July to September. From 1975 to 1996, 25% of the measured precipitation was rain, 44% was snow, and 31% was sleet or a combination of rain and snow (Førland and Hanssen-Bauer, 2000).

The Norwegian Meteorological Service (Metno) began weather observations there in 1969, and other research installations were also established. Specifically, the Ny-Ålesund radiation station, which is part of the Baseline Surface Radiation Network (BSRN), is operated since 1991 by the Alfred Wegener Institute (AWI) and the French Polar Institute Paul Emile Victor (IPEV) research base, and the Bayelva soil station is operated since 1998 by the AWI. Over the past decade the Bayelva catchment has been the focus of intensive investigations on soil and permafrost conditions (Roth and Boike, 2001; Boike et al., 2008; Westermann et al., 2010, 2011) and the annual surface energy balance (Boike et al., 2003; Westermann et al., 2009). Daily data of temperature and precipitation was used to analyze long-term trends (1969–2013); simulated temperature and precipitation from the phase 5 of the Coupled Model Intercomparison Project (CMIP5) experiment was used to asses projected changes for the time period of 2035–2064 (2050) and 2065–2094 (2080) for three different Representative Concentration Pathways (RCPs). Finally, hourly data from an automatic weather station allowed simulating the energy and mass balance of the snowpack under observed conditions (1998–2013), and under increasing temperature and precipitation.

Specifically, data from the Norwegian Meteorological Service (Metno, http://eklima.met.no/) from Ny-Alesund was used for analyzing long-term evolution of maximum and minimum temperatures, precipitation, and precipitation phase (solid, liquid [rain], or mixed [sleet and rain]) in the region. The lack of long-term climate records in Spitzbergen prevented the use of homogeneity tests with a reference series (Mestre et al., 2013). However, the temporal series were checked against available data from Svalbard airport (Longyearbyen) to identify potential inhomogeneities. These two interannual temperature and precipitation series were highly correlated (coefficients of correlation of 0.92 and 0.81 respectively), and had the same abrupt shifts at the same time (not shown). This suggests that the observed changes were not due to instrumental errors or changes at a specific site. Temporal trends of Tmax and Tmin were computed after a prewhitening procedure (Yue and Wang, 2002) to remove temporal autocorrelation. The Mann-Kendall test was used to determine the significance of the trend and the Thiel-Sen estimator to determine the linear slope of the trend (Hirsch et al., 1982). These non-parametric tests are widely used to determine monotonic trends in data series with different statistical distributions. Trends were calculated for the annual, early and late snow seasons.

Simulations of temperature and precipitation for the 20th and the 21st century, which were developed by 20 modelling groups in the framework of CMIP5 (Taylor et al., 2012), were used to predict climate changes in this region for 2050 and 2080. We used 3 radiative forcing scenarios defined by the RCPs: 2.6, 4.5, and 8.5. These correspond to the lowest, intermediate, and highest level of radiative forcing for the next few decades (Meinshausen et al., 2011). The annual, early and late snow seasons simulated precipitation and mean temperature for the period of 2035–2064 (centered in 2050) and 2065–2094 (centered in 2080) were subtracted from the simulated values for the period of 1980–2010 (control period) to calculate the magnitude of changes.

An automatic weather station at the Bayelva station (Fig. 1) measures hourly solar radiation, net radiation, air temperature, humidity, snow depth (albeit with frequent gaps in the dataset), soil temperature at different depths and liquid precipitation. Hourly precipitation data from the Norwegian Meteorological Institute's automatic precipitation gauge were cross-checked with daily precipitation measurements from a manual gauge (Boike et al., 2003), as well as data from BSRN station was used to fill the few gaps existing in the Bayelva station. A summary of sensors installed in Bayelva stations is provided in Table 1. This data is one of the best records available for polar regions and is sufficient for simulating the energy and mass balance of the snowpack from October 1998 to September 2013. Soil temperature was measured in a single profile at 11 depths (−4.5, −16.5, −17, 5, −33.5, −38.5, −60.5, −66.5, −91.5, −108.5 and −117.5 cm). We used data at −14.5 cm depth and completed the series up to September 2013 using data

Table 1
Sensors installed in Bayelva station.

Variable	Sensor	Period of operation	Height [m]	Unit	Integration method
Air temperature	Rotronic MP103A	Sep. 1998–Aug. 2009	0.5, 2	°C	Avg 1 h
Air temperature	Vaisala HMP45 PT100	Aug. 2009-present	2 0.5	°C	Avg 1 h
Relative humidity	Rotronic MP103A	Sep. 1998–Aug. 2009	0.5, 2	%	Avg 1 h
Relative humidity	Vaisala HMP45	Aug. 2009-present	2	%	Avg 1 h
Wind speed and direction	Young anemometer	Sep. 1998-present	3	m/s Deg	Avg 1 h
Net radiation	Kipp & Zonen NR Lite	Sep. 1998–Aug. 2009	1.6	W/m^2	Avg 1 h
Incoming infrared radiation	Kipp & Zonen Pyrgeometer CG1	Sep. 1998–Aug. 2009	2	W/m^2	Avg 1 h
Four component radiation	Huxeflux NR01	Aug. 2009-present	1.56	W/m^2	Avg 1 h
Precipitation	Young 52,203 unheated Tipping Bucket Rain Gauge	Sep. 1998-present	1.68	mm	Sum 1 h
Snow depth (1)	SCI SR50 ultrasound	Sep. 1998–Dez. 2012	1.45	m	Single 12 h
Snow depth	Jenoptik SHM30 laser distance	Aug 2013-present		m	Single 0.5 h
Soil temperature	32 × PT100	Sep. 1998–Dec. 2009	−0.045 … −1.25	°C	Avg 1 h
Soil temperature	8 × T107	Sep. 2009-present	−0.01 … −0.89	°C	Avg 1 h
Snow depth (2)	SR50	Aug. 2007-present		m	1 h

collected in another profile at −10 cm that was available from 2009 to 2014, after adjustment by linear regression ($r^2 = 0.97$).

The hourly meteorological data from Bayelva station was used as input to the Cold Regions Hydrological Modelling platform (CRHM) (Pomeroy et al., 2007). This model uses a flexible, object-oriented, modular structure (Leavesley et al., 2002) to simulate a range of hydrological processes in mountainous and cold regions (including blowing snow, interception, energy balance snowmelt, and infiltration of rain or melting water into frozen soils). Pomeroy et al. (2012) provide a more comprehensive description of the model and a scheme that illustrates the model structure. The CRHM has been applied to diverse environments, including alpine areas, subalpine areas, forests, and arctic basins (Pomeroy et al., 2007; Dornes et al., 2008; Essery et al., 2009; DeBeer and Pomeroy, 2010; Ellis et al., 2010; Fang et al., 2010; Knox et al., 2012; Rasouli et al., 2014).

Selection of the CRHM modules was mainly based on data availability and the adequacy of the climatic characteristics of the arctic sites, which have frozen soils, snow and strong winds. Evapotranspiration was calculated using the Penman-based equation of Granger and Pomeroy (1997). The layered energy balance snowmelt model (Snobal) (Marks and Dozier, 1992), which simulates the development and ablation of snow cover, was used for simulating snowmelt. An explicit (rather than iterative) energy balance solution and a relatively simple two-layer representation of snow cover (the upper 10 cm and the rest of the snowpack) were used for computational efficiency. This model solves the temperature and specific mass or depth of water-equivalent per unit area for each snow layer. Melt is computed in either layer when the layer temperature reaches or exceeds the melting point (0 °C) and the accumulated energy exceeds the cold content of snow. The phase of precipitation was determined by the Harder and

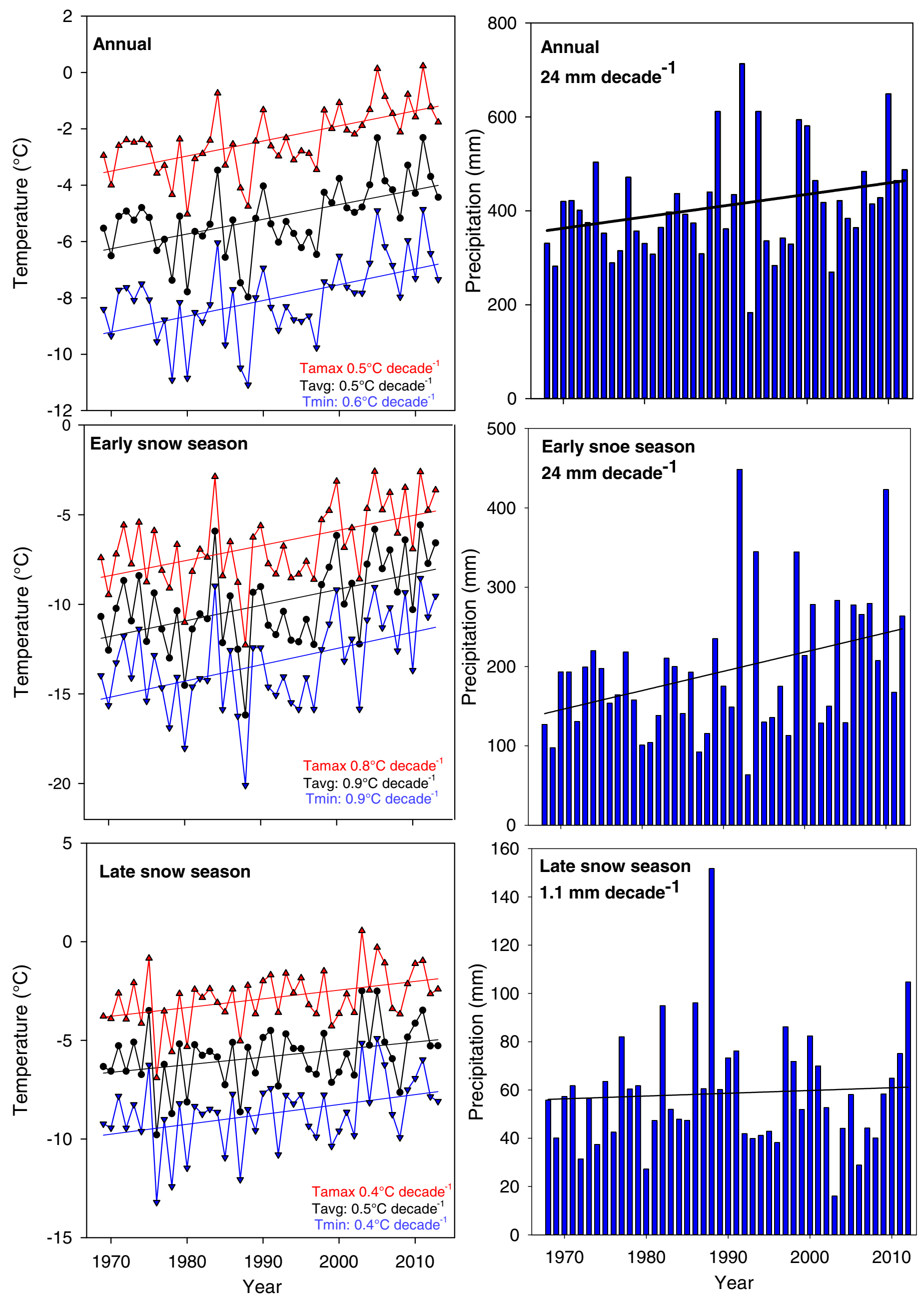

Fig. 2. Temperature and precipitation in Ny-Ålesund from 1969 to 2013 on an annual basis (top), during the early snow season (October–February, middle), and during the late snow season (March–June, bottom) from 1969 to 2013.

Pomeroy psychrometric energy balance procedure (Harder and Pomeroy, 2013), which has the advantage of not using subjective fixed thresholds for determining rain or snow precipitation. An albedo decay routine typically used in land surface schemes (Verseghy, 1991) was used to estimate the decay of snow albedo due to ageing. Wind redistribution and sublimation of blowing snow was estimated using the Prairie Blowing Snow Model module (Pomeroy and Li, 2000), which calculates two-phase transport of blowing snow in saltation and suspension and the sublimation of blowing snow particles as affected by the threshold conditions for snow transport which are a function of snowpack cohesion and ice layers. Solid precipitation undercatch due to wind was estimated by MacDonald and Pomeroy (2008) for Geonor precipitation gauge. Infiltration to frozen soils was estimated using a routine developed for frozen Canadian prairie soils (Gray et al., 1986) that segregates frozen soil infiltrability into three classes, primarily depending on the formation of a basal ice layer at the snow-soil interface.

We used sensitivity analysis to quantify the response of snow accumulation, snowpack duration, and other snow processes (frequency of rain on snow events and snow losses by sublimation and wind blowing) to changing climatic conditions, similarly to Brown and Mote (2009), Rasouli et al., 2014 and López-Moreno et al. (2013, 2014), to assess the effect of temperature change on snowpack characteristics. In particular, we simulated snowpack characteristics under observed conditions, under conditions in which the temperature was 1 °C to 5 °C greater and the precipitation was the same, and under conditions in which the temperature was 1 °C to 5 °C greater and the precipitation was 25% greater. The analyzed warming rates and the considered change in precipitation is the same that used in previous research dealing on snow sensitivity to climate change for other regions, which facilitates the comparability of the obtained results.

3. Results

3.1. Observed and projected changes in precipitation and temperature

Fig. 2 shows the maximum, minimum, and average temperatures (Tmax, Tmin, and Tavg) and precipitation for the annual, early and late snow seasons from 1969 to 2013. These results indicate statistically significant increases in temperature at this site during the last 45 years ($\alpha < 0.05$ for Tmax, Tmin, and Tavg). This temperature increase was particularly strong in the early snow season period, during which Tmax increased by 0.8 °C per decade and Tmin increased by 0.9 °C per decade. For the annual and snow-day periods, the rates of warming were 0.5° per decade for Tmax and 0.6 °C per decade for Tmin. Most of this observed warming was due to changes after the year 2000. In fact, there were no statistically significant changes in temperature from 1969 to 1995, but most years after 2000 had temperatures well above those from 1969 to 1999. There were also significant increases in precipitation for the annual and early snow season periods (24 mm per decade, $\alpha < 0.05$ for both), but not for the late snow season (1 mm per decade, $\alpha > 0.05$).

Fig. 3 shows the annual number of days with mixed precipitation (snow and rain) and the number of rainy days (liquid precipitation alone) during the early and late snow seasons. These four time series have high interannual variability and statistically significant increases over time ($\alpha < 0.05$ for all). These increases were particularly strong for mixed precipitation during the early snow season (1.4 days per decade) and for rainy days during the late snow season (0.7 days per decade).

Fig. 4 shows the projected average changes in annual, early and late snow seasons precipitation and temperature for 2035–2064 (centered at 2050) and 2065–2094 (centered at 2080) under three different RCPs (2.6, 4.5, and 8.5). The 3 RCPs project by 2050 a temperature increase of 3.5–4.5 °C for the annual period, 4–6 °C for the early snow season, and 2.3–3.9 °C for the late snow season. However, warming was very different for the 3 RCPs by 2080, with ranges of 4–11 °C for the annual period, 4–12.3 °C for the early snow period, and 2.2–10.3 °C for the late snow season. For precipitation, there were similar increases for the 3 RCP by 2050, with ranges of 4–12% for the annual period, 5–12% for the early snow season, and 3–11% for the late snow period. As with temperature, precipitation increase was very different for the 3 RCPs by 2080, and ranged from 3 to 24% for the annual period, 4–26% for the early snow season, and 2–21% for the late snow season. Interestingly, an RCP of 8.5 (highest radiative forcing) led to a smaller increase in precipitation than RCPs of 2.6 and 4.5, suggesting that future evolution of precipitation in this region is not directly related to evolution of greenhouse gas (GHG) emissions.

3.2. Simulation of snowpack from snow energy balance and sensitivity analysis

Fig. 5 shows the observed daily snow depth and snow depth simulated by the CRHM at the Bayelva weather station. Despite some logical biases, CRHM reproduced the annual cycle (accumulation and melting) and inter-annual variability. This suggests that the CRHM can be used to characterize snow depth and snow-water-equivalent (SWE) evolution, and also the cumulative energy balance and the bulk temperature of the snow pack, which drive the onset of melting. Fig. 5B and C show the annual maximum accumulation and the annual duration of the snowpack. In both cases the simulated and observed values were similar for most study years. In particular, the observed and simulated mean maximum snowpack depths were 0.96 m and 0.93 m, respectively, and the snowpack duration was 225 days and 219 days, respectively. The simulated series slightly underestimated the inter-annual coefficients of variation

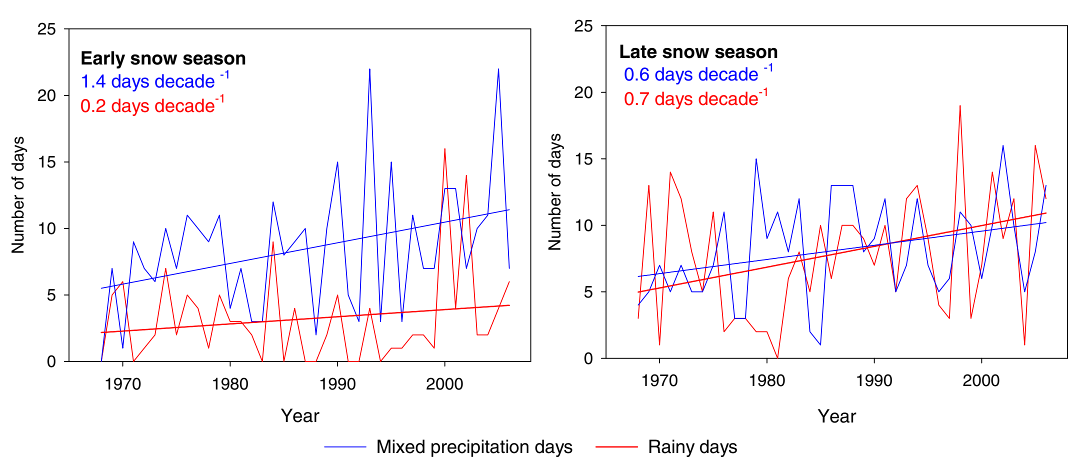

Fig. 3. Number of days with mixed precipitation (snow and rain) and rain alone during the early (left) and late (right) snow seasons in Ny-Ålesund from 1969 to 2013.

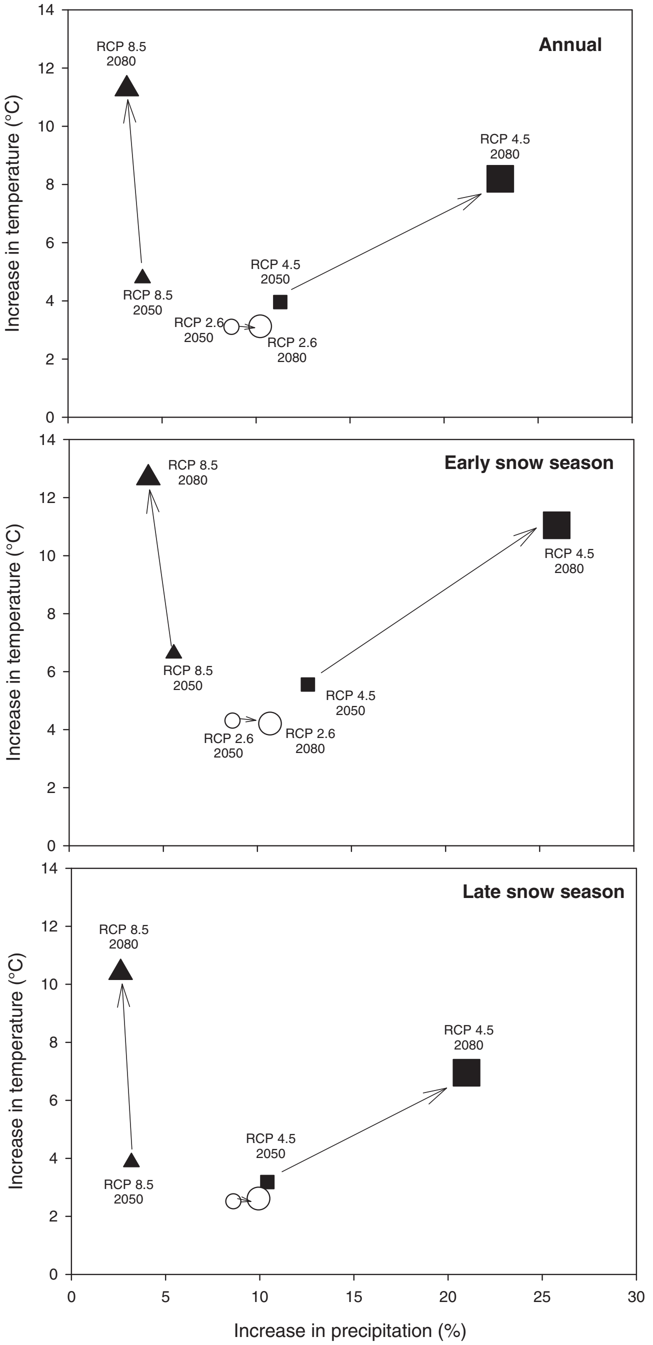

Fig. 4. Projected increase of temperature and precipitation in Svalbard relative to 1980–2010 for the time period of 2035–2064 (2050) and 2065–2094 (2080) for three different Representative Concentration Pathways (RCPs of 2.6, 4.5 and 8.5). Calculations used general circulation model (GCM) ensembles from phase 5 of the Coupled Model Intercomparison Project (CMIP 5).

(maximum annual accumulation: 0.19 [observed] and 0.16 [simulated]; snowpack duration: 0.17 [observed] and 0.14 [simulated]).

Fig. 6 shows the evolution of mean daily SWE series simulated for 1998–2013 under observed climatic conditions with temperature increases of 1–5 °C but no change in precipitation (left), and assuming a 25% increase of precipitation at the observed temperature and a temperature increase from 1 to 5 °C (right). The mean annual snow evolution illustrates a marked decrease in the magnitude of peak in SWE (from 235 to 100 mm) with increasing temperature from observed conditions to 5 °C warmer. The timing of peak SWE advanced 32 days as temperature warms 5 °C, but the timing of complete snow ablation only advanced in 17 days. This resulted in an increase in the duration of the ablation period – the time from maximum SWE to snow ablation – which makes physical sense as there is less solar radiation energy available to drive snowmelt if it is initiated earlier in the spring (López-Moreno et al., 2013). Logically, the simulations that considered a 25% increase of precipitation exhibited around 25% higher SWE values, but almost identical patterns of decreasing SWE, earlier peak SWE and complete snow ablation than under stationary precipitation.

Fig. 7A and B show the long-term (1998–2013) averages, with the 10th and 90th percentiles, of SWE for the early and late snow season, under observed conditions (T_{obs}), and under conditions with temperature increases of 1–5 °C with and without a 25% increase of precipitation. Table 2 summarizes the results of the sensitivity analysis. Assuming a temperature increase alone, mean SWE decreased progressively as temperature increased at a mean rate of 10.2% per degree Celsius for the early snow season and 11.1% per degree for the late snow season. For both of these periods, the SWE change per degree was greater as the temperature increase was greater; in other words, the slopes were greater from $T_{+3\ °C}$ to $T_{+5\ °C}$ than from T_{obs} to $T_{+3\ °C}$. The error bars (10th and 90th percentiles) remained very similar and their sizes appeared independent of the intensity of warming, despite the decrease in average snow accumulation. This suggests an increase of interannual variability in snow accumulation as the climate warms. Logically, the mean SWE values were a 20–25% greater with a 25% increase in precipitation in addition to warming. More specifically, with a 25% increase in precipitation, mean SWE values were higher than those under conditions that were about 1 °C colder but with no change in precipitation. However, a 25% increase in precipitation only had a small effect on snowpack duration at different temperatures (9.8% per degree for the early snow season, 11.2% per degree for the late snow season).

The patterns were very similar for peak SWE (Fig. 7C), with an average decrease of 7.1% per degree, and increasing rate of decline for greater warming. As above (Fig. 7A and B), the maximal SWE was greater with a 25% increase in precipitation, although the rate of decline (6.9% per degree) was similar. With a 25% increase of precipitation, maximal SWE values were higher than those under conditions that were 5 °C colder but with no changes in precipitation. The average duration of the snowpack (Fig. 7D) decreased by about 11 days per degree, and the decline was sharper for a temperature increases of 3 °C and greater. A 25% increase in precipitation had little impact on snowpack duration (simulated snowpack lasted between 4 and 7 days more than simulations that did not considered precipitation change).

Fig. 8 shows that the number of days with precipitation as rain increased from 4.6 to 14.6 for the early snow season and from 8.3 to 13.7 for the late snow season as the temperature increased by 5 °C. This corresponds 43.5% more days for the October–February and 12.8% more days for March–June period per degree of warming. In contrast to mean snowpack, peak SWE, and snowpack duration, there was no increase in the rate at which rainfall days increase with temperature as warming increased.

Fig. 9 shows that climate warming leads to a marked decrease in the fraction of snowfall (October–June) that is eroded by blowing snow and then subsequently sublimated from the surface (left) and transported from the site by wind (right). In particular, the CRHM indicated that an average of 41 mm of snow is sublimated from blowing snow per year under observed climatic conditions, and this declines to 17 mm with a 5 °C increase in temperature (~6 mm per degree); as a percentage, this corresponds to a change from 13% to 9% of total snowfall, although in this case there is an accelerated decline after a warming of 2 °C. The rates of these declines differ because the amount of snowfall is altered by climate warming-induced increased precipitation as rainfall. Simulations also indicate a decrease in snowfall transported by blowing snow from 13.2% under observed conditions to 7.6% when the temperature increases by 5 °C. As with blowing snow sublimation,

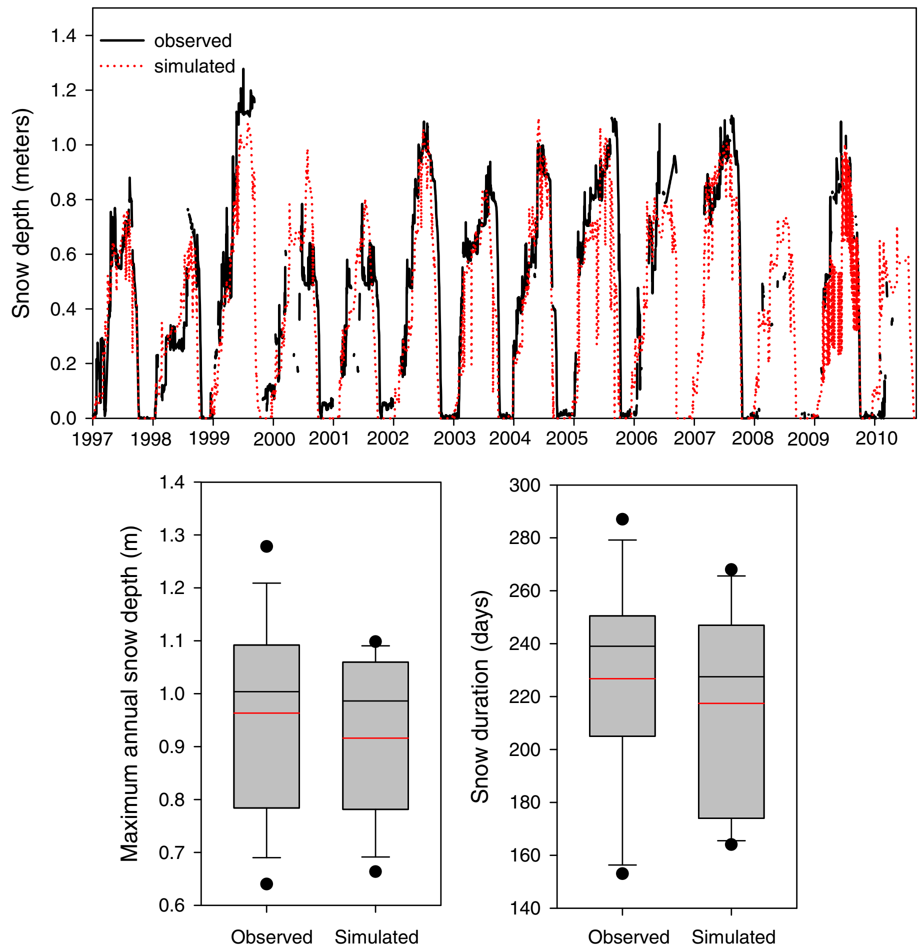

Fig. 5. Comparison of observed and simulated snow depths from 1998 to 2013 (top). Boxplots inform about the interannual variability of maximum reached snow depth and duration of the snowpack: the mean (inner red line), (inner black line), 25th and 75th percentiles (boxes), 10th and 90th percentiles, and 5th and 95th percentiles (dots). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

the decrease in the total snow transported from the site by wind was nearly linear with temperature (−1.5% per degree), but not when these numbers are presented as percentages.

4. Discussion and conclusions

Previous studies reported significant increases in air temperature at Svalbard, including Ny-Ålesund (Forland et al., 2011). However, the magnitude of these changes depends on the selected study period, as for example there were no significant changes from 1975 to 1998 (Przybylak, 2000). The present study uses updated climate data, and has identified a clear acceleration of the warming rate to unprecedented values of 0.5–0.9 °C per decade, depending on the temperature variable (Tmin, Tmax, Tavg), for study? periods ending? after the year 2000. In line with most previous climatology research in the Arctic, there is also a tendency for a wetter climate in Ny-Ålesund, with a notable increase from 2000 to 2013.

This paper focused on the snow-dominated seasons, and distinguished the early snow season (October–February) from the late snow season (March–June). The results indicated that the early snow period was more sensitive to climate change than the late snow season. The March–June period also experienced intense warming, but precipitation remained almost stationary. An important consequence of these trends is an increase in the likelihood of rain-on-snow events. Previous studies reported similar changes (Førland et al., 1997, Førland and Hanssen-Bauer, 2000, Vihma et al., 2016) and an association of these changes with a sudden increase of snowpack temperature, even in mid-winter (Putkonen and Roe, 2003; Westermann et al., 2011). These changes are facilitated by a shallower snowpack as expected under warmer climates (Martín-Moreno and Serrano-Cañadas, 2013).

Long-term studies have not examined snowpack (depth or snow-water equivalent) (Førland and Hanssen-Bauer, 2000), because arctic snowpack has an uneven distribution due to wind in the open tundra (Pomeroy and Li, 2000; Essery and Pomeroy, 2004; King et al., 2008; Iacozza and Barber, 2010; Mott et al., 2010), and point snow depth measurements are not representative (Pomeroy and Gray, 1995). For this reason, there are no detailed assessments about how the observed climate change has affected snow processes such as accumulation, redistribution, ablation, duration, or the timing of the onset of ablation. Nonetheless, some studies based on remote sensing or the interannual evolution of albedo indicated a tendency toward a shallower and shorter-lasting snowpack (Winther et al., 2003; Nowak and Hodson, 2013; Maturilli et al., 2014), and reduced portion of snowfall compared to rain (Bring et al., 2016). These results agree with the conclusions of Brown and Mote (2009) at the hemispheric scale, and with Callaghan et al. (2011a, 2011b)who studied arctic regions. Thus, there is abundant evidence that climate change has greatly affected snowpack in polar maritime regions in recent decades.

Climate models project patterns similar to the observed evolution of temperature and precipitation in recent decades (Hanssen-Bauer, 2002; Haugen et al., 2000). The present study showed that the extent of warming and precipitation increase both depend on the RCP

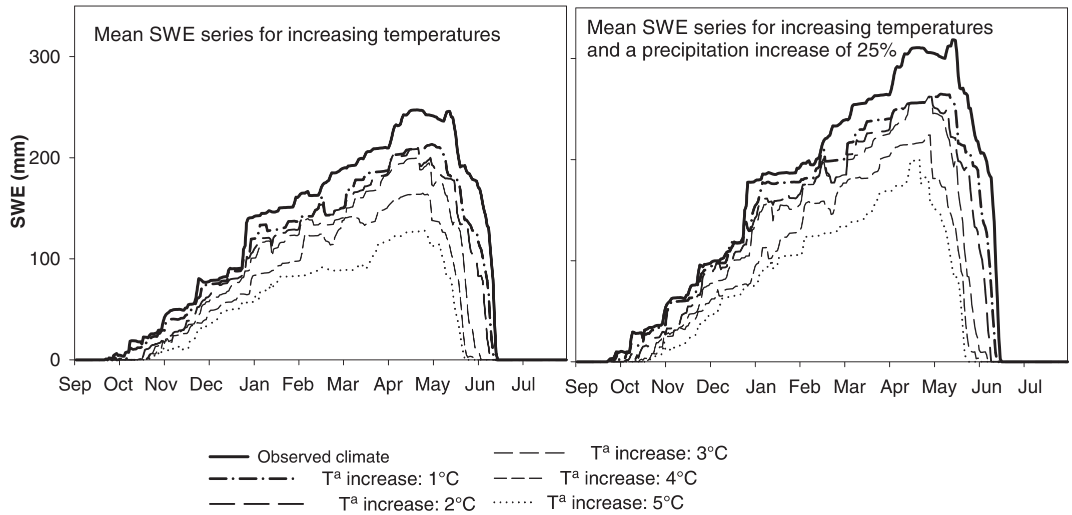

Fig. 6. Annual mean snow water equivalent (SWE) at the observed temperature and with increasing temperature but no change in precipitation (left). Mean snow water equivalent (SWE) assuming a 25% increase of precipitation at the observed temperature and with increasing temperature (right).

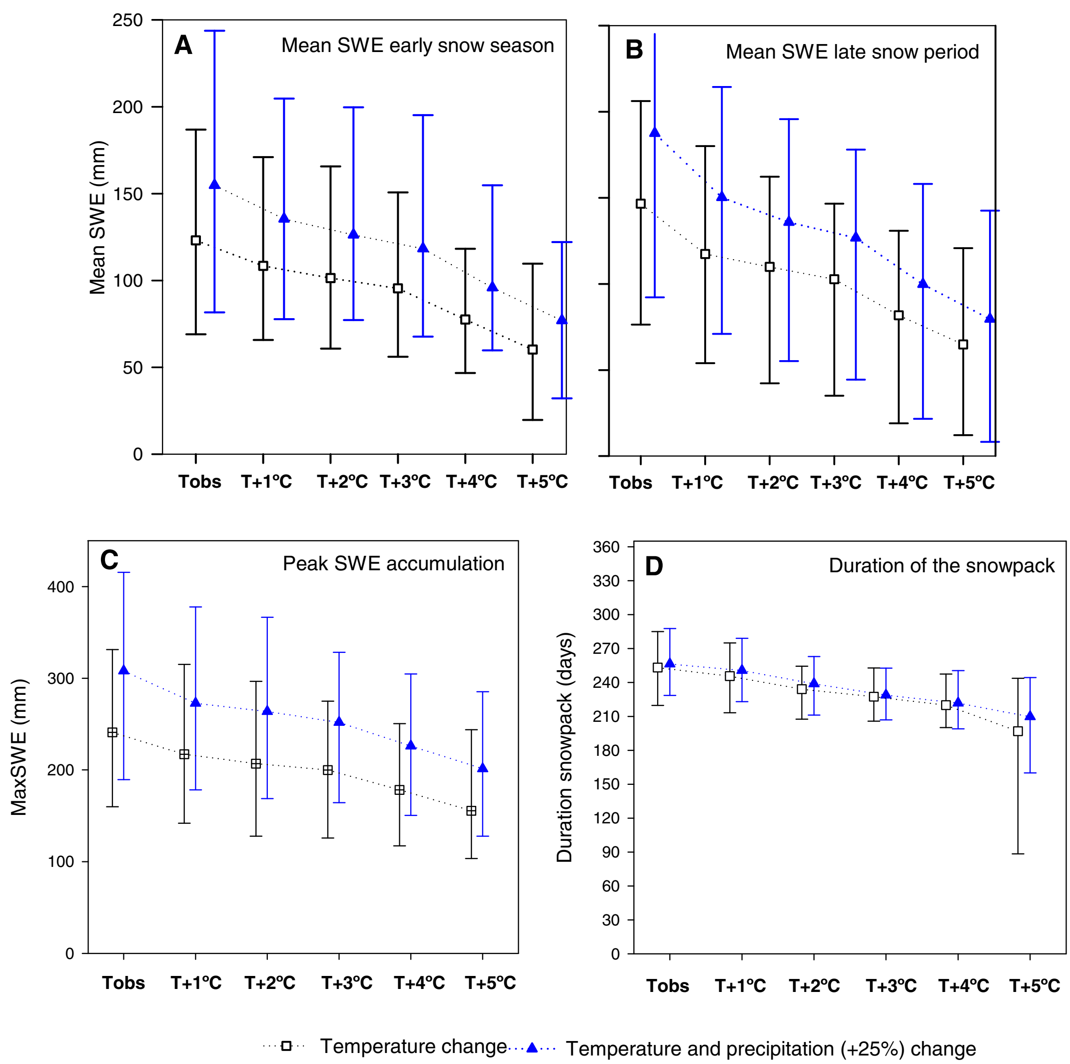

Fig. 7. Mean SWE at the observed temperature (T_{obs}) and with different levels of warming for the early (A) and late (B) snow seasons. Annual maximum SWE (C) and snowpack duration (D) at the observed temperature (T_{obs}) and with different levels of warming.

Table 2
Summarizes the results of the sensitivity analysis.

	Mean SWE: early snow season	Mean SWE +25% precipitation: early snow season	Mean SWE: late snow season	Mean SWE +25% precipitation: late snow season	Peak SWE	Peak SWE +25% precipitation	Duration SWE	Duration SWE +25% precipitation	Days with liquid precipitation: early snow season	Days with liquid precipitation: late snow season	% of snowfall sublimated	% of snowfall drifted out by wind
Units	cm	cm	cm	cm	cm	cm	Days	Days	Days	Days	%	%
T observed	123.1	158.0	146.6	192.6	241.4	308.2	252.7	256.5	4.6	8.3	13.2	9.9
T + 1 °C	108.4	138.7	117.5	155.4	217.4	272.9	245.1	250.7	6.7	9.5	13.0	8.9
T + 2 °C	101.4	129.5	109.9	141.0	207.2	263.8	233.5	238.9	8.8	10.9	13.4	7.8
T + 3 °C	95.5	121.6	102.7	132.0	200.3	252.0	226.9	228.9	10.6	12.1	11.2	6.0
T + 4 °C	77.6	99.1	81.8	105.1	178.5	226.3	219.5	221.8	12.9	13.0	8.9	4.2
T + 5 °C	60.3	80.1	64.9	85.0	155.8	201.4	196.3	209.9	14.6	13.7	7.6	3.0
Mean change per °C	−12.5	−15.6	−16.4	−21.5	−17.1	−21.4	−11.3	−9.3	2.6	1.1	−1.1	−1.4
% change per °C from observed temperature	−10.2	−9.8	−11.1	−11.2	−7.1	−6.9	−4.5	−3.6	43.5	12.8	−8.5	−14.5

considered. All three considered RCPs project an annual temperature increase of 3.4–4.5 °C for 2050 and an increase that could exceed 10 °C by 2080 under the most pessimistic pathways (RCP 4.5 and 8.5). The October–February period is projected to warm slightly more (~1 °C) than the March–June period. Precipitation is also projected to increase, but the magnitude of this change is highly variable depending on the RCP (5–26%). This is in line with the results of Hanssen-Bauer (2002), who projected that annual precipitation will increase significantly up to 2050 mainly because of increased spring precipitation. Dynamical downscaling (Haugen et al., 2000) projects an even greater precipitation increase (~2% per decade) at the west coast of Spitsbergen. Despite the importance of snow in this environment, no published research has yet assessed the effect of increased temperature or precipitation on the snowpack in Svalbard.

This sensitivity analysis has indicated a general decrease of snowpack thickness, water equivalent and duration as the climate warms. In particular, the mean SWE will decrease by 10.2% per degree increase during the early snow season and by 11.1% per degree increase during the late snow season. For each degree increase of temperature, the peak SWE declines by 6.9%, the duration of snowpack declines by 11 days, and the number of days with rainfall increases by 43.5% (early snow season) or 12.8% (late snow season). A warmer climate also leads to marked decreases in blowing snow sublimation (−8.5% per degree) and the percentage of snowfall transported from the site by blowing snow (−14.5% per degree). The decreased snow transport occurs because warm snow is more often wet snow, which is not as readily eroded as is cold and dry snow (Li and Pomeroy, 1997; López-Moreno et al., 2013). For most of the analyzed snowpack parameters, the response to climate warming accelerates as the simulated warming is greater, especially above 3 °C. A similar effect also occurs in the Spanish Pyrenees in southern Europe, but the threshold is 2 °C (López-Moreno et al., 2014).

A 25% increase of precipitation partially compensates for the reduced snowpack due to warming, and this is more evident for the peak SWE. Thus, the peak SWE with a 5 °C increase of temperature and a 25% increase of precipitation is similar to that simulated under the observed climate. However, peak SWE occurs earlier in the season, has a much warmer snowpack, is rapidly followed by the onset of snow ablation, and is associated with a reduction in snow duration. Indeed the rates of decrease in snowpack duration per degree are similar with and without a 25% increase in precipitation, except for a temperature increase of 5 °C. Similarly, previous sensitivity analyses for the whole northern hemisphere indicated that peak SWE is likely to remain stationary or even increase as result of increased precipitation, but the duration of the snow season will be drastically reduced (Brown and Mote, 2009). Previous studies have highlighted the importance of changes in precipitation for estimating the impact of climate change in particular areas (Dyer and Mote, 2006; Räisänen, 2007; Brown and Mote, 2009), and indicated that high uncertainty in simulated precipitation hampers to ability to obtain reliable projections of future snowpack (Lopez-Moreno et al., 2009).

Despite the notable sensitivity of snow accumulation (indicated by snowpack) in Ny-Ålesund to climate warming, snowpack reduction per degree increase of temperature is clearly lower than the 11–20% reduction for the Pyrenees (López-Moreno et al., 2013, 2014), the 20% reduction for the Washington Cascades (Casola et al., 2009), and the 15% reduction for the Swiss Alps (Beniston et al., 2003). In an Arctic catchment in Yukon, Rasouli et al. (2014) found a lower sensitivity than previous reported studies with an approximate reduction of 7% of peak SWE per degree increase of temperature. The lower sensitivity of Yukon and Ny-Ålesund to warming is likely due to the colder temperatures in this arctic region than in the other study areas, because sensitivity to warming increases as the temperature is closer to 0 °C. In agreement with this interpretation, mountain areas at lower elevations respond more rapidly to climate change (Pomeroy et al., 2003; Lopez-Moreno et al., 2009; Hopkinson et al., 2011). Moreover, snow

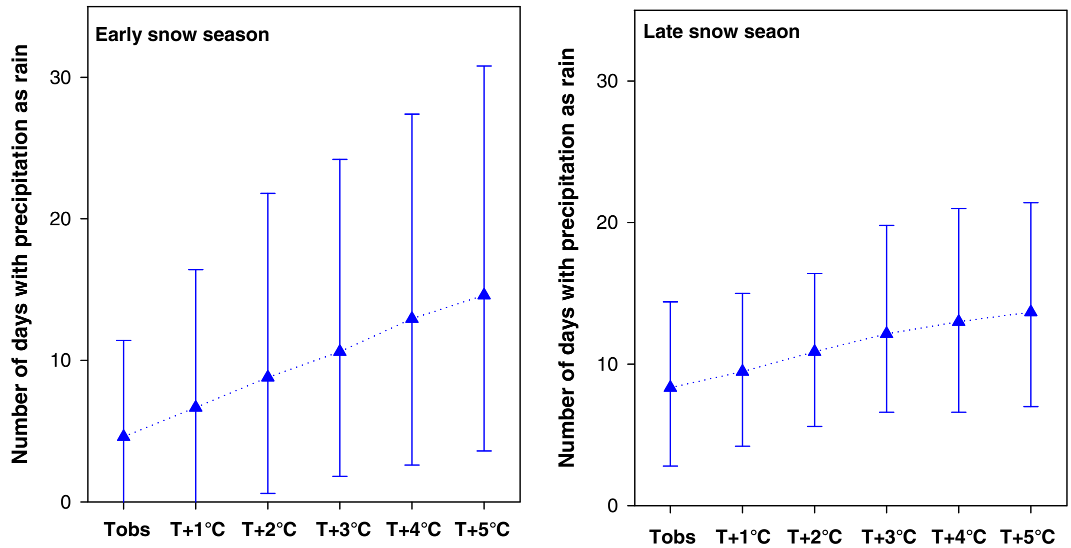

Fig. 8. Simulated number of days with precipitation as rain at the observed temperature (T_{obs}) and with different level of warming during the early (left) and late (right) snow seasons. Dots indicate averages and bars indicate 10th and 90th percentiles.

losses by blowing snow sublimation and transport are very high in the Arctic, and the marked reduction of these processes under warmer conditions may modulate snowpack sensitivity to temperature increase.

Climate simulations using 3 RCPs, even the one with the lowest radiative forcing (2.6), project a temperature increase of 5 °C by the end of this century in the case of the one with the lowest radiative forcing (RCP 2.6), and by the mid this century in the case of the two more pessimistic ones (RCPs 4.5 and 8.5). According to GCMs projections, the mean snowpack would be 51% lower (early snow season) and 55% lower (late snow season), the peak SWE would be 35% lower, and the overall duration of snow cover would be 56.5 days less. However, these numbers also depend on changes in precipitation and other important variables of the snow energy balance (radiative flux, wind speed, and relative humidity) as it was noted by Rasouli et al. (2014) for the Yukon. Nonetheless is undeniable that a shallower and shorter-lasting snowpack is likely to have many environmental and economic consequences, as described by Callaghan et al. (2011a, 2011b). These include significant alterations in hydrological and ecological systems because of changes in surface energy balance (e.g., albedo), water balance (water storage and release, mass balance of the glaciers, shifts from snow- to rain-dominated regimes, etc.), thermal regimes of the soil (insulation), vegetation, fauna, and fluxes of trace gases (Jones et al., 2001; Wheeler et al., 2016). Thus, multiannual trend analyses carried out in Breelva basin (Spitsbergen, Svalbard) revealed that rainfall has a growing importance on outflow compared to snow and ice melting (Majchrowska et al., 2015). The livelihoods and well-being of Arctic residents and the many services they provide to the wider population depend on snow conditions in the Arctic, so these changes may have important consequences. Thus, more research is needed to fully understand the consequences of changes in the snow energy balance in additional polar regions, and the effect of snowpack on permafrost temperature and depth of the active layer (Westermann et al., 2011). Thus, in a recent study, Semenchuk et al. (2016) showed how the impact of climate warming on snowpack in Svalbard and Northwest Territories, Canada, increases growing-season respiration and turn arctic tundra in carbon source. It is also necessary to understand the sensitivity of these parameters to other climatic variables in this region, such as radiative flux, which changes with increasing cloudiness and reduced

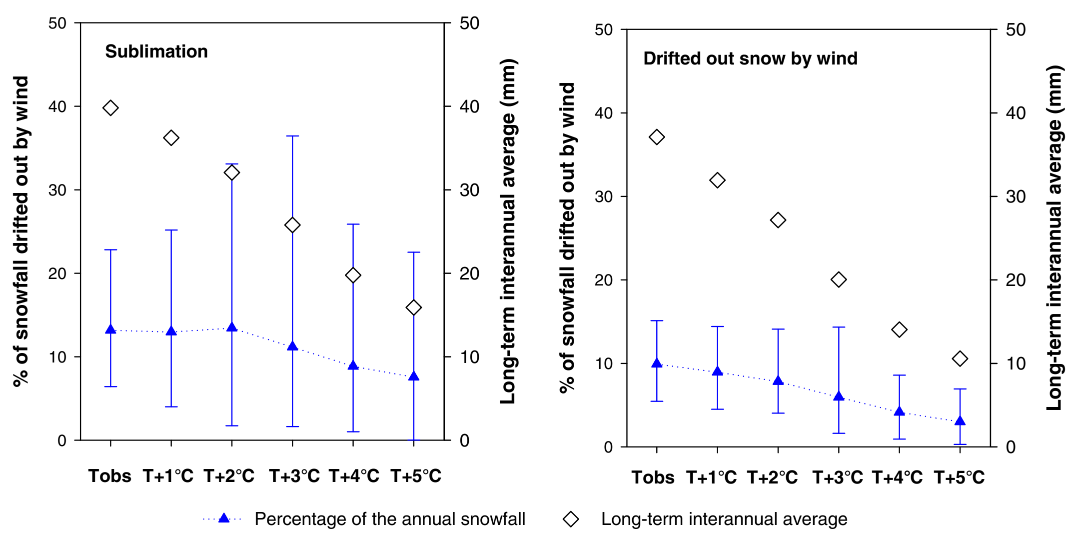

Fig. 9. Snow sublimated (left) and drifted out by the wind (right) at the observed temperature (T_{obs}) and with different levels of warming. Values indicate the percentage of total snowfall and the long-term interannual average. Solid triangles indicate averages and bars indicate the 10th and 90th percentiles.

albedo (Maturilli et al., 2014; Vihma et al., 2016) and may feedback to affect snowpack and other climatic variables.

Acknowledgements

This work was supported by the research projects CGL2011-27536/HID: “Hidrologia nival en el Pirineo central español: variabilidad espacial, importancia hidrológica y su respuesta a la variabilidad y cambio climático” and IBERNIEVE (CGL2014-5259/HID), financed by the Spanish Commission of Science and Technology. Arturo Sanchez-Lorenzo was supported by the postdoctoral fellowship JCI-2012-12508 funded by the Ministry of Economy and Competitiveness.

References

Beniston, M., Keller, F., Koffi, B., Goyette, S., 2003. Estimates of snow accumulation and volume in the Swiss Alps under changing climatic conditions. Theor. Appl. Climatol. 76, 125–140.

Boike, J., Roth, K., Ippisch, O., 2003. Seasonal snow cover on frozen ground: energy balance calculations of a permafrost site near Ny-Ålesund, Spitsbergen. J. Geophys. Res. 108 (D2), 8163. http://dx.doi.org/10.1029/2001JD000939.

Boike, J., Ippisch, O., Overduin, P., Hagedorn, B., Roth, K., 2008. Water, heat and solute dynamics of a mud boil, Spitsbergen. Geomorphology 95, 61–73.

Bring, A., Fedorova, I., Dibike, Y., Hinzman, L., Mård, J., Mernild, S.H., Prowse, T., Semenova, O., Stuefer, S.L., Woo, M.-K., 2016. Arctic terrestrial hydrology: a synthesis of processes, regional effects, and research challenges. J. Geophys. Res. Biogeosci. 121. http://dx.doi.org/10.1002/2015JG003131.

Brown, R.D., Mote, P.W., 2009. The response of northern hemisphere snow cover to a changing climate. J. Clim. 22, 2124–2145. http://dx.doi.org/10.1175/2008JCLI2665.1.

Callaghan, T.V., Johansson, M., Brown, R.D., Groisman, P.Y., Labba, N., Radionov, V., Barry, R.G., Bulygina, O.N., Essery, R.H.L., Frolov, D.M., Golubev, V.N., Grenfell, T.C., Petrushina, M.N., Razuvaev, V.N., Robinson, D.A., Romanov, P., Shindell, D., Shmakin, A.B., Sokratov, S.A., Warren, S., Yang, D., 2011a. The changing face of Arctic snow cover: a synthesis of observed and projected changes. Ambio 40, 17–31.

Callaghan, T.V., Johansson, M., Brown, R.D., Groisman, P.Y., Labba, N., Radionov, V., Bradley, R.S., Blangy, S., Bulygina, O.N., Christensen, T.R., Colman, J.E., Essery, R.L.H., Forbes, B.C., Forchhammer, M.C., Golubev, V.N., Honrath, R.E., Juday, G.P., Meshcherskaya, A.V., Phoenix, G.K., Pomeroy, J., Rautio, A., Robinson, D.A., Schmidt, N.M., Serreze, M.C., Shevchenko, V.P., Shiklomanov, A.I., Shmakin, A.B., Sköld, P., Sturm, M., Woo, M.-K., Wood, E.F., 2011b. Ambio 40, 32–45.

Casola, J.H., Cuo, L., Livneh, B., Lettenmaier, D.P., Soelinga, M.T., Mote, P.W., Wallace, J., 2009. Assessing the impacts of global warming on snowpack in the Washington Cascades. J. Clim. 22, 2758–2772.

DeBeer, C.M., Pomeroy, J.W., 2010. Simulation of the snowmelt runoff contributing area in a small alpine basin. Hydrol. Earth Syst. Sci. 14, 1205–1219.

Derksen, C., Brown, R.D., 2012. Spring snow cover extent reductions in the 2008–2012 period exceeding climate model projections. Geophys. Res. Lett. 39, L19504. http://dx.doi.org/10.1029/2012GL053387.

Dornes, P.F., Pomeroy, J.W., Pietroniro, A., Carey, S.K., Quinton, W.L., 2008. Influence of landscape aggregation in modelling snow-cover ablation and snowmelt runoff in a sub-arctic mountainous environment. Hydrol. Sci. J. 53 (4), 725–740.

Dyer, J.L., Mote, T.L., 2006. Spatial variability and trends in observed snow depth over North America. Geophys. Res. Lett. 33, L16503. http://dx.doi.org/10.1029/2006GL027258.

Eckerstorfer, M., Christiansen, H.H., 2011. Topographical and meteorological control on snow avalanching in Longyearbyen area, central Svalbard 2006–2009. Geomorphology 134, 186–196.

Ellis, C.R., Pomeroy, J.W., Brown, T., MacDonald, J., 2010. Simulations of snow accumulation and melt in needleleaf forest environments. Hydrol. Earth Syst. Sci. 14, 925–940.

Essery, R.L.H., Pomeroy, J.W., 2004. Vegetation and topographic control of wind-blown snow distributions in distributed and aggregated simulations for an Arctic tundra basin. J. Hydrometeorol. 5, 734–744.

Essery, R., Rutter, N., Pomeroy, J.W., Baxter, R., Stahli, M., Gustafsson, D., Barr, A., Bartlett, P., Elder, K., 2009. SNOWMIP2: an evaluation of forest snow process simulations. Bull. Am. Meteorol. Soc. 90 (8), 1120–1135.

Fang, X., Pomeroy, J.W., Westbrook, C.J., Guo, X., Minke, A.G., Brown, T., 2010. Prediction of snowmelt derived streamflow in a wetland dominated prairie basin. Hydrol. Earth Syst. Sci. 14, 991–1006. http://dx.doi.org/10.5194/hess-14-1-2010.

Førland, E.J., Hanssen-Bauer, I., 2000. Increased precipitation in the Norwegian Arctic: true or false? Clim. Chang. 46, 485–509.

Førland, E.J., Hanssen-Bauer, I., Nordli, P.Ø., 1997. Climate statistics and long-term series of temperature and precipitation at Svalbard and Jan Mayen. Nor. Meteorol. Inst. Rep. 72 21/97 KLIMA.

Førland, E.J., Benestad, R., Hanssen-Bauer, I., Haugen, J.E., Skaugen, T.E., 2011. Temperature and precipitation development at Svalbard 1900–2100. Adv. Meteorol. 1–14, ID893790 2011.

Granger, R.J., Pomeroy, J.W., 1997. Sustainability of the western Canadian boreal forest under changing hydrological conditions—2-summer energy and water use. In: Rosjberg, D., Boutayeb, N., Gustard, A., Kundzewicz, Z., Rasmussen, P. (Eds.), Sustainability of Water Resources Under Increasing Uncertainty. IAHS Press, Wallingford, pp. 243–250 IAHS Publ No. 240.

Graversen, R.G., Mauritsen, T., Tjernström, M., Källén, E., Svensson, G., 2008. Vertical structure of recent Arctic warming. Nature 451, 53–56. http://dx.doi.org/10.1038/nature06502.

Gray, D.M., Granger, R.J., Landine, P.G., 1986. Modelling snowmelt infiltration and runoff in a prairie environment. Cold Regions Hydrology. AWRA, Bethesda, Maryland, USA, pp. 427–438 (Symp. Proo., ed. Kane, D.L.).

Hanssen-Bauer, I., 2002. Temperature and precipitation in Svalbard 1912–2050: measurements and scenarios. Polar Record 38 (206), 225–232.

Hanssen-Bauer, I., Førland, E.J., 1998. Long-term trends in precipitation and temperature in the Norwegian Arctic: can they be explained by changes in atmospheric circulation patterns? Clim. Res. 10, 143–153.

Harder, P., Pomeroy, J.W., 2013. Estimating precipitation phase using a psychrometric energy balance method. Hydrol. Process. 27 (13), 1901–1914. http://dx.doi.org/10.1002/hyp.9799.

Haugen, J.E., Bjørge, D., Nordeng, T.E., 2000. Dynamic downscaling: further results. RegClim General Technical Report, No.4 (Available from the Norwegian Institute for Air Research).

Hirsch, R.M., Slack, J.R., Smith, R.A., 1982. Techniques of trend analysis for monthly water quality data. Water Resour. Res. 18 (1), 107–121.

Hopkinson, R.F., McKenney, D.W., Milewska, E.J., Hutchinson, M.F., Papadopol, P., Vincent, L.A., 2011. Impact of aligning climatological day on gridding daily maximum–minimum temperature and precipitation over Canada. J. Appl. Meteorol. Climatol. 50, 1654–1665. http://dx.doi.org/10.1175/2011JAMC2684.1.

Iacozza, J., Barber, D.G., 2010. An examination of snow redistribution over smooth land-fast sea ice. Hydrological Processes 24, 850–865. http://dx.doi.org/10.1002/hyp.7526.

Isaksen, K., Vonder Mühll, D., Gubler, H., Kohl, T., Sollid, J.L., 2000. Ground surface temperature reconstruction based on data from a deep borehole in permafrost at Janssonhaugen, Svalbard. Ann. Glaciol. 31, 287–294.

Isaksen, K., Sollid, J.L., Holmlund, P., Harris, C., 2007. Recent warming of mountain permafrost in Svalbard and Scandinavia. J. Geophys. Res. 112, F02S04. http://dx.doi.org/10.1029/2006JF000522.

Jones, H.G., Pomeroy, J.W., Walker, D.A., Hoham, R.W., 2001. Snow Ecology: An Interdisciplinary Examination of Snow-Covered Ecosystems. Cambridge University Press, UK (394 pp).

King, J.C., Pomeroy, J.W., Gray, D.M., Fierz, C., Föhn, F.M.B., Harding, R.J., Jordan, R.E., Martin, E., Plüss, C., 2008. Snow-atmosphere energy and mass balance. In: Armstrong, R.L., Brun, E. (Eds.), Snow and Climate. Cambridge University Press, Cambridge, pp. 70–124.

Knox, S.H., Carey, J.K., Humphreys, E.R., 2012. Snow surface energy exchanges and snowmelt in a shrub-covered bog in Eastern Ontario, Canada. Hydrol. Process. 26 (12), 1876–1890.

Langen, P.L., Alexeev, V.A., 2007. Polar amplification as a preferred response in an idealized aquaplanet GCM. Clim. Dyn. 29 (2–3), 305–317.

Leavesley, G.H., Markstrom, S.L., Restrepo, P.J., Viger, R.J., 2002. A modular approach to addressing model design, scale, and parameter estimation issues in distributed hydrological modelling. Hydrol Processes 16 (2), 173–187.

Li, L., Pomeroy, J.W., 1997. Probability of blowing snow occurrence by wind. J. Geophys. Res. 102, D18.

Lopez-Moreno, J.I., Goyette, S., Beniston, M., 2009. Impact of climate change on snowpack in the Pyrenees: horizontal spatial variability and vertical gradients. J. Hydrol. 374 (3–4), 384–396.

López-Moreno, J.I., Pomeroy, J., Revuelto, J., Vicente-Serrano, S.M., 2013. Response of snow processes to climate change: spatial variability in a small basin in the Spanish Pyrenees. Hydrol. Process. 27 (18), 2637–2650.

López-Moreno, J.I., Revuelto, J., Gilaberte, M., Morán-Tejeda, E., Pons, M., Jover, E., Esteban, P., Garcia, C., Pomeroy, J., 2014. The effect of aspect on the response of snowpack to climate warming in the Pyrenees. Theor. Appl. Climatol. 117 (1–2), 207–219.

Lüers, J., Westermann, S., Piel, K., Boike, J., 2014. Annual CO2 budget and seasonal CO_2 exchange signals at a high Arctic permafrost site on Spitsbergen, Svalbard archipelago. Biogeosciences 11 (22), 6307–6322.

MacDonald, J., Pomeroy, J.W., 2008. Gauge undercatch of two common snowfall gauges in a prairie environment. Proceedings of the Eastern Snow Conference. 64, pp. 119–126.

Majchrowska, E., Ignatiuk, D., Jania, J., Marszalek, H., Wasik, M., 2015. Seasonal and interannual variability in runoff from the Werenskioldbreen catchment, Spitzbergen. Pol. Polar Res. 36 (3), 197–224.

Marks, D., Dozier, J., 1992. Climate and energy exchange at the snow surface in the alpine region of the Sierra Nevada: 2 snow cover energy balance. Water Resour. Res. 28 (11), 3043–3054.

Martín-Moreno, R., Serrano-Cañadas, E., 2013. Evolución del manto nival en el alto ártico, nordenskiöld land (Spitsbergen, svalbard). Boletín de la Asociación de Geógrafos Españoles 61, 215–234.

Maturilli, M., Herber, A., König-Langlo, G., 2013. Climatology and time series of surface meteorology in Ny-Ålesund, Svalbard, Earth Syst. Sci. Data 5 (1), 155–163.

Maturilli, M., Herber, A., König-Langlo, G., 2014. Surface radiation climatology for ny-ålesund, svalbard (78.9° n), basic observations for trend detection. Theor. Appl. Climatol. 1–9 http://dx.doi.org/10.1007/s00704-014-1173-4 (2014).

Meinshausen, M., Smith, S.J., Calvin, K.V., Daniel, J.S., Kainuma, M.L.T., Lamarque, J.-F., Matsumoto, K., Montzka, S.A., Raper, S.C.B., Riahi, K., Thomson, A.M., Velders, G.J.M., van Vuuren, D., 2011. The RCP greenhouse gas concentrations and their extension from 1765 to 2300. Clim. Chang. 109 (1–2), 213–241.

Mestre, O., Domonkos, P., Picard, F., Auer, I., Robin, S., Lebarbier, E., Böhm, R., Aguilar, E., Guijarro, J., Vertachnik, G., Klancar, M., Dubuisson, B., Stepanek, P., 2013. HOMER: a homogenization software – methods and applications. IDÖJÁRÁS 117 (1), 47–67.

Miller, G.H., Alley, R.B., Brigham-Grette, J., 2010. Arctic amplification: can the past constrain the future? Quat. Sci. Rev. 29 (15–16), 1779–1790.

Mott, R., Schirmer, M., Bavay, M., Grünewald, T., Lehning, M., 2010. Understanding snow-transport processes shaping the mountain snow-cover. The Cryosphere 4, 545–559. http://dx.doi.org/10.5194/tc-4-545-2010.

Nordli, Ø., Przybylak, R., Ogilvie, A.E.J., Isaksen, K., 2014. Long-term temperature trends and variability on Spitsbergen: the extended Svalbard Airport temperature series, 1898–2012. Polar Res. 33, 21349.

Nowak, A., Hodson, A., 2013. Hydrological response of a High-Arctic catchment to changing climate over the past 35 years: a case study of Bayelva watershed, Svalbard. Polar Res. 32, 19691.

Pomeroy, J.W., Brun, E., 2001. Physical properties of snow. In: Jones, H.G., Pomeroy, J.W., Walker, D.A., Hoham, R.W. (Eds.), Snow Ecology: An Interdisciplinary Examination of Snow-covered Ecosystems 45–118p. Cambridge University Press, Cambridge, UK.

Pomeroy, J.W., Gray, D.M., 1995. Snow Accumulation, Relocation and Management. National Hydrology Research Institute Science Report No. 7. Saskatoon, Environment Canada (144 pp).

Pomeroy, J.W., Li, L., 2000. Prairie and arctic areal snow cover mass balance using a blowing snow model. J. Geophys. Res. 105 (D21), 26619–26634. http://dx.doi.org/10.1029/2000JD900149.

Pomeroy, J.W., Toth, B., Granger, R.J., Hedstrom, N.R., Essery, R.L.H., 2003. Variation in surface energetics during snowmelt in complex terrain. J. Hydrometeorol. 4 (4), 702–716.

Pomeroy, J.W., Gray, D.M., Hedstrom, N.R., Quinton, W.L., Granger, R.J., Carey, S.K., 2007. The cold regions hydrological model: a platform for basing process representation and model structure on physical evidence. Hydrol. Process. 21, 2650–2667.

Pomeroy, J.W., Fang, X., Ellis, C., 2012. Sensitivity of snowmelt hydrology in Marmot Creek, Alberta to forest cover disturbance. Hydrol. Process. 26, 1891–1904.

Przybylak, R., 2000. Temporal and spatial variation of surface air temperature over the period of instrumental observations in the Arctic. Int. J. Climatol. 20, 587–614. http://dx.doi.org/10.1002/(SICI)1097-0088.

Putkonen, J., Roe, G., 2003. Rain-on-snow events impact soil temperatures and affect ungulate survival. Geophys. Res. Lett. 30 (4), 1188.

Räisänen, J., 2007. How reliable are climate models? Tellus A 59, 2–29. http://dx.doi.org/10.1111/j.1600-0870.2006.00211.x.

Rasouli, K., Pomeroy, J.W., Janowicz, J.R., Carey, S.K., Williams, T.J., 2014. Hydrological sensitivity of a northern mountain basin to climate change. Hydrol. Process. 28, 4191–4208. http://dx.doi.org/10.1002/hyp.10244.

Roth, K., Boike, J., 2001. Quantifying the thermal dynamics of a permafrost site near Ny-Ålesund, Svalbard. Water Resour. Res. 37 (12), 2901–2914.

Screen, J.A., Simmonds, I., 2010. The central role of diminishing sea ice in recent Arctic temperature amplification. Nature 464, 1334–1337. http://dx.doi.org/10.1038/nature09051.

Semenchuk, P.R., Christiansen, C.T., Grogan, P., Elberling, B., Cooper, E.J., 2016. Long-term experimentally deepened snow decreases growing-season respiration in a low- and high-arctic tundra ecosystem. J. Geophys. Res. Biogeosci. 121, 1236–1248. http://dx.doi.org/10.1002/2015JG003251.

Sepp, M., Jaagus, J., 2011. Changes in the activity and tracks of Arctic cyclones. Clim. Chang. 105, 577–595. http://dx.doi.org/10.1007/s10584-010-9893-7.

Serreze, M.C., Walsh, J.E., Chapin, F.S., Osterkamp, T., Dyurgerov, V., Romanovsky, V., Oechel, W.C., Morison, J., Zhang, T., Barry, R.G., 2000. Observational evidence of recent change in the northern high-latitude environment. Clim. Chang. 46, 159–207.

Taylor, K.E., Stouffer, R.J., Meehl, G.A., 2012. A Summary of the CMIP5 Experiment Design. Bull. Amer. Meteor. Soc 93, 485–498.

Verseghy, D., 1991. CLASS-A Canadian land surface scheme for GCMs, I. Soil model. Int. J. Climateol. 11, 111–133.

Vihma, T., Screen, J., Tjernström, M., Newton, B., Zhang, X., Popova, V., Deser, V., Holland, M., Prowse, T., 2016. The atmospheric role in the Arctic water cycle: a review on processes, past and future changes, and their impacts. J. Geophys. Res. Biogeosci. 121, 586–620. http://dx.doi.org/10.1002/2015JG003132.

Westermann, S., Lüers, J., Langer, M., Piel, K., Boike, J., 2009. The annual surface energy budget of a high-arctic permafrost site on Svalbard, Norway. Cryosphere 3, 245–263.

Westermann, S., Wollschläger, U., Boike, J., 2010. Monitoring of active layer dynamics at a permafrost site on Svalbard using multi-channel ground-penetrating radar. Cryosphere 4, 475–487.

Westermann, S., Boike, J., Langer, M., Schuler, T.V., Etzelmüller, B., 2011. Modeling the impact of wintertime rain events on the thermal regime of permafrost. Cryosphere 5 (4), 945–959.

Wheeler, J.A., Cortes, A.J., Sedlacek, J., Karrenberg, S., van Kleunen, M., Wipf, S., et al., 2016. The snow and the willows: earlier spring snowmelt reduces performance in the lowlying alpine shrub Salix herbacea. J. Ecol. http://dx.doi.org/10.1111/1365-2745.12579.

Winther, J.-G., Bruland, O., Sand, K., Gerland, S., Marechal, D., Ivanov, B., Glowacki, P., König, M., 2003. Snow research in Svalvard: an overview. Polar Res. 22 (2), 125–144.

Yue, S., Wang, C.Y., 2002. Applicability of prewhitening to eliminate the influence of serial correlation on the Mann-Kendall test. Water Resour. Res. 38 (6), 1068. http://dx.doi.org/10.1029/2001WR000861.

Zhang, X., Sorteberg, A., Zhang, J., Gerdes, R., Comiso, J.C., 2008. Recent radical shifts of atmospheric circulations and rapid changes in Arctic climate system. Geophys Res Lett 35, L22701. http://dx.doi.org/10.1029/2008GL035607.