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Declining summer snowfall in the Arctic: causes, impacts and feedbacks

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Abstract Recent changes in the Arctic hydrological cycle are explored using in situ observations and an improved atmospheric reanalysis data set, ERA-Interim. We document a pronounced decline in summer snowfall over the Arctic Ocean and Canadian Archipelago. The snowfall decline is diagnosed as being almost entirely caused by changes in precipitation form (snow turning to rain) with very little influence of changes in total precipitation. The proportion of precipitation falling as snow has decreased as a result of lower-atmospheric warming. Statistically, over 99% of the summer snowfall decline is linked to Arctic warming over the past two decades. Based on the reanalysis snowfall data over the ice-covered Arctic Ocean, we derive an estimate for the amount of snow-covered ice. It is es-

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timated that the area of snow-covered ice, and the proportion of sea ice covered by snow, have decreased significantly. We perform a series of sensitivity experiments in which inter-annual changes in snow-covered ice are either unaccounted for, or are parameterized. In the parameterized case, the loss of snow-on-ice results in a substantial decrease in the surface albedo over the Arctic Ocean, that is of comparable magnitude to the decrease in albedo due to the decline in sea ice cover. Accordingly, the solar input to the Arctic Ocean is increased, causing additional surface ice melt. We conclude that the decline in summer snowfall has likely contributed to the thinning of sea ice over recent decades. The results presented provide support for the existence of a positive feedback in association with warming-induced reductions in summer snowfall.

Keywords Arctic · Precipitation · Snow · Sea ice · Albedo feedback · Climate change

1 Introduction

Recent climate change has been especially pronounced in the Arctic region, with surface temperatures rising two to four times faster than the global average (Solomon et al, 2007; Bekryaev et al, 2010; Miller et al, 2010) and an accompanying rapid decline of sea ice (Serreze et al, 2007; Stroeve et al, 2007). Both the Arctic warming and sea ice loss in the past few decades are unprecedented over at least the last few thousand years (Kaufmann et al, 2009; Polyak et al, 2010). A multitude of climate feedbacks have been proposed that amplify the Arctic surface air temperature response to climate forcing (either natural or anthropogenic). Whilst some remain poorly understood and their existence unconfirmed (Francis et al, 2009), others, for example the ice-albedo feedback, are already believed to be active and contributing significantly to recent Arctic change (Serreze et al, 2009; Screen and Simmonds, 2010a).

In its simplest form the ice-albedo feedback can be understood as decreases in sea ice cover, that expose open water with a lower albedo than ice and increase the solar energy absorbed by the coupled ocean-ice-atmosphere system. As the system warms, the sea ice cover further declines reinforcing the warming. The decline in sea ice extent over recent decades and its associated positive feedback have been widely documented (e.g., Serreze and Francis, 2006; Perovich et al, 2007; Screen and Simmonds, 2010a). However, changes in sea ice cover are not the only driver of the ice-albedo feedback. Changes occurring within the ice pack, for example to its snow cover or melt pond fraction, also contribute (Curry et al, 1995). Such changes have received less attention than the pronounced decline of sea ice extent.

The characteristics of the Arctic sea ice cover and its albedo change through the annual cycle. The winter ice cover is overlaid by a relatively thick layer of snow with a high albedo (note that during the polar night the albedo is irrelevant to the energy budget as there is no incoming sunlight). As sunlight returns to the Arctic in spring, the snow cover begins to melt. The albedo decreases, first as dry snow turns to wet snow, and then further as bare ice is exposed (Perovich et al, 2002). By mid-summer, the seasonal snow cover has largely disappeared (Warren et al, 1999). Arctic storms, most prevalent in summer (Serreze and Barrett, 2008; Simmonds et al, 2008), produce snowfall and an ephemeral summer snow cover. The formation and development of melt ponds during the summer further lower the ice albedo (Perovich et al, 2002). In autumn, as surface temperatures decrease, the melt ponds refreeze and the snow cover returns, raising the albedo.

Rising air temperatures over recent decades have likely led to an earlier onset and lengthening of the melt season (Markus et al, 2009). Warming may also be influencing Arctic precipitation patterns with associated impacts on the snow cover. Evaluations of Arctic precipitation changes are largely based on gauge measurements at meteorological stations on land. They suggest an increase in Arctic precipitation over recent decades (White et al, 2007; Min et al, 2008). Supporting evidence comes from the monitoring of discharge from the major Eurasian rivers that drain into the Arctic Basin. These rivers have predominantly shown increases in discharge (e.g., Peterson et al, 2002; McClelland et al, 2006), likely in part due to increased continental precipitation (Wu et al, 2005), but also other factors such as earlier snow melt, thawing permafrost, land use change, shifting fire regimes and changing river management (e.g., dams) (McClelland et al, 2004). Changes in precipitation over the Arctic Ocean are more uncertain, in most part due to spatially and temporally sparse observations. Peterson et al (2006) report an increase in maritime precipitation over the last half-century based on the European Centre for Medium-range Weather Forecasts' (ECMWF) ERA-40 reanalysis. However, this data set has known weaknesses in its representation of the Arctic moisture budget (Cullather et al, 2000; Serreze and Hurst, 2000). Data from rawinsondes and satellites suggest no coherent large-scale change in net precipitation (precipitation minus evaporation) over the Arctic Ocean (see review of White et al, 2007).

Continental snow cover has decreased in March associated with warmer winters, greater snow melt and a decrease in the fraction of precipitation that occurs as snow (e.g., McCabe and Wolock, 2010). Again, changes over the ice-covered Arctic Ocean are harder to ascertain. Warren et al (1999) analyzed snow depths from the Russian North Pole drifting stations and reported a decline in snow depth in all months over the period 1954-1991, with the largest decline in May. These authors found no evidence of earlier onset of melt and concluded that the reduction in May snow depth was likely related to reduced snowfall. Here we provide the first Arctic-wide estimates of the recent evolution of summer snow cover over the ice-covered Arctic Ocean. We document a significant decline in summer snowfall and examine its causes. Then, in a series of sensitivity experiments, we go on to explore the importance of these changes for the ice-albedo feedback.

2 Data

We draw on data from two primary sources: atmospheric reanalyses from ERA-Interim (ERA-I) and station observations from northern Canada.

2.1 ERA-Interim

ERA-I is the latest global atmospheric reanalysis produced by the ECMWF, covering the data-rich period since 1989 (Simmons et al, 2006). The ECMWF, applying lessons learned from earlier reanalysis efforts and well-documented weaknesses in older reanalyses, have implemented a number of improvements in ERA-I. These include an assimilating model with higher spectral resolution, improved model physics, a more sophisticated hydrological cycle, and data assimilation based on a 12-hourly four-dimensional variational analysis (4D-var) (Dee and Uppala, 2009). ERA-I only became available to the scientific community in 2009 and consequently the validation and evaluation of the output is in its infancy. However, early indications suggest that there have been significant improvements in the representation of Arctic temperature trends (Screen and Simmonds, 2010a, in press), and in the global hydrological cycle (Simmons et al, 2006; Uppala et al, 2008; Simmons et al, 2010) in ERA-I versus older reanalyses. These improvements were in the forefront of our mind when choosing the most appropriate data set from a number of available products. That said, the accuracy of long-term precipitation changes in ERA-I remain unclear due to differences in the various reference data sets used for validation purposes (Simmons et al, 2010). This uncertainty in ERA-I precipitation trends, and its implications for our conclusions, is discussed further in later sections.

Snowfall, precipitation, surface albedo and net solar radiation data are archived as daily or monthly means for midday and midnight, and at time steps of 3-, 6-, 9- or 12-hours. For precipitation and snowfall, we summed the 12-hour accumulated totals for midday and midnight to give daily accumulated precipitation (mm d^{-1}) and snowfall (mm-we d^{-1}). For non-accumulated fields (albedo, solar radiation) we averaged the values at the same time-steps to give a daily-mean. Sea ice fraction and air temperature are archived as daily means. Where monthly or seasonal data are used these are monthly or seasonal means of the daily data.

For oceanic grid-boxes, the albedo in ERA-I, a_{era}, is given by:

$$a_{era} = c(1 - a_i) + (1 - c)(1 - a_w), \tag{1}$$

where a_i is the albedo of sea ice, a_w is the albedo of water and c is the sea ice fraction. In ERA-I, a_i has a crude seasonal cycle that is held constant from year-to-year (Fig. 1). These values have been interpolated from seasonal mean albedo values in Ebert and Curry (1993), with the value for summer (0.51) representative of bare sea ice and the value in winter (0.77) representing dry snow. This partially mimics the observed seasonal evolution of a_i: a_i is high during winter as the ice is covered by dry snow, begins to decrease in early summer as the snow cover starts to melt, reaches its lowest values in mid-summer due to melt pond formation and rises rapidly in early fall as melt ponds refreeze and the snow cover returns (Curry et al, 2001; Perovich et al, 2002). In comparison to observations during the Surface Heat Budget of the Arctic (SHEBA) field campaign (Curry et al, 2001; Perovich et al, 2002), the ERA-I a_i is slightly too low in the cold season and too high in mid-to-late summer. The lower a_i observed in summer is due to the presence of melt ponds that are not accounted for in ERA-I.

In contrast to a_i, a_w does not vary with season and is fixed at 0.06. Thus, a_{era} over ocean varies only according to the climatological annual cycle of a_i and the time-varying c. In Section 6, we modify this albedo parameterization so that the albedo also varies according to time-varying changes in estimated snow cover.

2.2 Canadian observations

We obtained monthly total precipitation, rainfall, snowfall and mean surface temperature for twelve meteorological stations in northern Canada from Environment Canada (http://www.climate.weatheroffice.gc.ca/index.html). The stations used were (from north to south) Eureka, Resolute, Cambridge Bay, Kugluktuk, Baker Lake, Coral Harbour, Mayo, Rankin Inlet, Fort Simpson, Hay River, Watson Lake and Fort Smith. These stations were chosen based on their high-latitude locations (north of 60°N) and because they had data spanning the whole period 1989-2009. On occasion, these stations were separated into two categories: Arctic (north of

70°N; Eureka and Resolute) and sub-Arctic (60-70°N; other ten stations). In addition to the monthly data, we have examined daily data for the Arctic stations. These meteorological data have been independently processed and quality control checks undertaken by Environment Canada. Missing data are flagged and we make no effort to infill these. Data are also flagged when "incomplete" or "estimated". In both cases, we treat these data points as missing data. Days with small amounts (less than about 0.1mm or 0.1mm-we) of a precipitation type are flagged as "trace". At the Canadian Arctic stations, up to 50% of the observations have snowfall reported as trace amounts. Inclusion of the trace events is therefore important. Here we consider a trace amount to be equal to 0.05mm or 0.05mm-we. Note, the results were not sensitive to small changes in these values.

3 Snowfall-to-Precipitation ratio

The proportion of precipitation occurring as snow can be expressed by the snowfall-to-precipitation ratio (SPR):

$$SPR = \frac{S_{we}}{P}, \tag{2}$$

where S_{we} is the daily total snowfall water equivalent (mm-we d^{-1}) and P is the daily total precipitation (mm d^{-1}). Days with no precipitation are not considered. Therefore, an SPR of zero indicates a day when all precipitation fell in liquid form rather than a day with no precipitation.

Fig. 2 (top left) shows the daily SPR as a function of daily mean surface temperature, observed at Resolute in Arctic Canada. A clear, but highly non-linear, relationship exists between the proportion of precipitation falling as snow and surface temperature. Precipitation falls almost entirely as snow on days with mean temperatures below around 260K. Conversely, precipitation falls almost entirely as rain on days with mean temperatures above around 278K. On days with mean temperatures between 260-278K, precipitation can exist in both liquid and solid forms. However, as temperatures approach melting point there is a rapid transition from predominantly snowfall to rainfall. (The cluster of points at SPR of 0.5 is partly an artifact of the observing precision: a large number of days had "trace" amounts of rainfall and snowfall, hence a SPR of 0.5). A nearly identical relationship between SPR and temperature is found in ERA-I sub-sampled at the grid-box containing Resolute (Fig. 2, top right). Similar plots were obtained for other stations (not shown) and have been shown by other authors (e.g., Ledley, 1985).

SPR is influenced by changes in both snowfall and total precipitation, but neither of these variables display a simple relationship with daily mean temperature (Fig. 2 middle & bottom). Very little snowfall is observed, or depicted by ERA-I, when the daily mean temperature exceeds 275K. The upper limits of both daily snowfall and total precipitation decrease with decreasing temperature (below about 275K) in a quasi-exponential manner, because changes in precipitable water are tied to changes in temperature. However, changes in temperature do not lead to a consistent change in total snowfall or precipitation.

The dependence of SPR on temperature can be further seen in Fig. 3, which shows the mean annual cycles of SPR and surface temperature averaged over the Arctic from ERA-I. Here and in what follows, the Arctic-mean SPR was calculated as:

$$SPR_{arctic} = \frac{\overline{S_{we}}}{\overline{P}}, \tag{3}$$

where the overbars denote area-averages north of 70°N. Through a large portion of the year, the Arctic-mean temperature is well below freezing point and the vast majority of precipitation falls as snow. In winter (December-February), precipitation falls entirely as snow over most of the Arctic and rain normally only occurs in the Norwegian, Greenland and Barents Seas. The largest regional contribution to the Arctic-mean winter rainfall comes from the Norwegian Sea area where as much as 60% of precipitation falls as rain even in winter. With the exception of this region, atmospheric warming at this time of year is unlikely to result in a large change in SPR because it is too cold for rain. However, in summer (June-August) the Arctic-mean temperature is near to melting point and even small changes in temperature have the potential to cause changes in precipitation form, as SPR is highly sensitive to changes in temperature within this mean temperature range (Fig. 2, top). We hypothesize that in a warming Arctic (warming is observed in all months, see Screen and Simmonds (2010b)), the proportion of summer precipitation falling as snow will decrease as a direct result of atmospheric warming. In the next section, we test this hypothesis based on Canadian meteorological observations and ERA-I reanalyses.

An often reported problem with precipitation observations is the underestimation of the *real* precipitation due to gauge undercatch (e.g., Forland and Hanssen-Bauer, 2000). The wet-day mean precipitation at Resolute is 0.62 mm d^{-1} compared to 0.91 mm d^{-1} in ERA-I sub-sampled at the grid-box containing Resolute. Part of this difference may be related to precipitation undercatch. Forland and

Hanssen-Bauer (2000) estimated that the true annual precipitation may be up to 50% greater than the recorded precipitation at sites in the Norwegian Arctic. Undercatch of solid precipitation may be greater than liquid precipitation (Forland and Hanssen-Bauer, 2000). However, the SPR is highly consistent between observations and ERA-I, with both having a mean SPR at Resolute of 0.84 over all wet-days. Looking only at wet-days when the mean surface temperature was above 260K (i.e. the temperature range when solid and liquid precipitation both occur), the mean SPR is 0.61 and 0.69 for the observations and ERA-I, respectively. Thus, the observed SPR may be slightly underestimated in the warm season, although bias in ERA-I cannot be ruled out as a cause of the difference. Importantly for the trend analyses that follow, we found no obvious tendencies or discontinuities in the SPR difference between observations and ERA-I as a function of time.

4 Arctic hydrological changes

Fig. 4 shows the linear change in SPR, snowfall, rainfall and total precipitation over the period 1989-2009, in each season and at each of the Canadian meteorological stations (gray crosses). The stations show a wide range of observed change, both in the sign and magnitude of the trends. In order to simplify this spatially variability, we also plot multi-station means for the Arctic (north of 70°N) and the sub-Arctic (60-70°N). A large SPR decrease is found in summer for the Arctic stations, associated with a pronounced decrease in summer snowfall and an increase in summer rainfall. In the sub-Arctic, summer precipitation has increased, almost exclusively in liquid form. Outside of summer, noteworthy decreases in spring and autumn precipitation and snowfall have occurred at the Arctic stations.

In ERA-I, the largest seasonal-mean Arctic-mean changes in SPR are in summer, when the proportion of precipitation falling as snow has significantly decreased (Fig. 5), consistent with the Canadian observations. A smaller, but still statistically significant (at the 90% level), decrease in SPR is depicted by ERA-I in autumn. However, these two seasons show contrasting changes in snowfall and total precipitation. In summer, there has been a decrease in ERA-I snowfall and total precipitation, again in agreement with observations at the Arctic stations. The decline of snowfall exceeds the total precipitation decrease resulting in the decrease in SPR. Given that the majority of precipitation in summer falls as rain (Fig. 3), one would expect a decrease in total precipitation to be associated with a decrease in rainfall. That rainfall has increased and not decreased (Fig. 5), when the total precipitation has decreased, reflects the changes in SPR. Had SPR remained constant, rainfall would have decreased in line with decreasing precipitation. In autumn, ERA-I shows a large increase in total precipitation and only a small increase in snowfall. This increase in snowfall cannot be related to changes in precipitation form as SPR has decreased and must be related to the increase in total precipitation. This additional precipitation has disproportionately fallen as rain rather than snow, as reflected by the decrease in SPR and the large increase in rainfall. In contrast, autumn precipitation decreased at the Arctic stations (Fig. 4). Spring total precipitation, snowfall and rainfall have all decreased in ERA-I, resulting in negligible change in SPR (Fig. 5). Lastly, in winter, there has been little change in any of these hydrological indicators in ERA-I or observations. It is worth noting that the Arctic-mean changes in ERA-I are broadly consistent with the observed changes at the Arctic stations in summer, spring and winter, but discrepancies exist during autumn.

Since the atmosphere has warmed in all seasons over the past two decades (Screen and Simmonds, 2010a,b) and that there is a general expectation of greater Arctic precipitation in a warming climate (Finnis et al, 2007; Holland et al, 2007; Kattsov et al, 2007), it is worth considering briefly why we don't find a consistent picture of precipitation increases in ERA-I (Fig. 5) or observations (Fig 4). In the observations, the only pronounced precipitation increases are found in summer at the sub-Arctic stations. Autumn is the only season that displays an Arctic-mean precipitation increase in ERA-I, although we reiterate this is not supported by single-point observations from Eureka or Resolute. Despite increases in air temperature and humidity in ERA-I (Screen and Simmonds, 2010a), it depicts substantial decreases in total precipitation in spring and summer. A possible explanation for this discrepancy may be that a decrease in storm activity has counter-acted the increase in precipitable water. To explore this possibility, we applied the University of Melbourne cyclone tracking algorithm (Simmonds et al, 2008; Simmonds and Keay, 2009) to ERA-I mean sea level pressure fields. Fig. 6 shows the seasonal-mean Arctic-mean changes in two important cyclone statistics: cyclone number and mean cyclone depth that provide information on the number of cyclones and their average intensity, respectively. (For more details of these statistics and their derivation, the reader is directed to Simmonds et al (2008) and references therein.) ERA-I depicts significant decreases in both cyclone variables in spring, giving credence to the explanation of reduced spring precipitation due to reduced cyclone activity. In summer, there is a shift toward weaker Arctic cyclones over the study period that may partly explain the decrease in summer precipitation. We note that these cyclone changes are a robust feature in three alternative reanalyses over the last two decades (not shown).

Whilst the cyclone changes identified in Fig 6 help reconcile the precipitation changes over the same time period, they must be viewed with caution in the context of longer-term Arctic trends. Over the longer period, 1979-2008, the cyclone statistics show few significant trends across the seasons and a number of reanalyses (updated from Simmonds et al, 2008). By extension, the precipitation changes in ERA-I over the last two decades may not be representative of multi-decadal Arctic precipitation trends. Indeed, land-based observations suggest increases in Arctic precipitation since 1950 (Min et al, 2008). Over this longer period, it may be that long-term warming and the associated increases in precipitable water have had greater influence on precipitation trends than have changes in Arctic cyclones. Whilst further work is required to confirm this, it would not be surprising if longer- and shorter-term precipitation changes were predominately driven by different processes. We do, however, expect the changes in SPR over the last two decades to be broadly consistent with longer-term changes, as they are very strongly related to atmospheric warming and Arctic air temperatures have risen in each decade since 1970 (Gillett et al, 2008). Further discussion on the accuracy of the precipitation, snowfall and SPR changes in ERA-I is provided in Section 5.

Fig. 7 shows the spatial extent of SPR changes in the summer months. We pay particular attention to SPR changes in northern Canada, where stations observations are used to validate ERA-I. Decreases in SPR are found over much of the Arctic Ocean and in all three summer months. There are some differences in the patterns, most notably, the largest SPR decreases are found in the Beaufort Sea region during June and at higher latitudes in July. This latitudinal shift is also apparent in the observations. During June, the Arctic stations of Resolute and Eureka display strong decreases in SPR. SPR decreases at these stations are less in July, when the largest changes have occurred at higher latitudes. In northern Canada, ERA-I and station observations show broadly the same SPR changes. In June, large SPR decreases have occurred at the most northerly stations (Resolute and Eureka) and are well-captured by ERA-I. The north-east coastal regions of mainland Canada exhibit modest increases in SPR, which are also seen in ERA-I, albeit with some minor differences in regional extent. The inland stations show little change in SPR and nor does ERA-I over inland northern Canada. In July and August, the most pronounced SPR changes have occurred at the far-northern stations with only small changes at the other stations. This north-south gradient is well-represented in ERA-I.

The spatial pattern of observed SPR change is related to temperature change. Stations and months that display SPR decreases (the far-northern stations) have generally warmed whilst stations that display SPR increases (north-east mainland) have generally cooled (Fig. 8). However, the temperature changes alone are insufficient to explain the lack of SPR change at many stations and months. This insensitivity of SPR to temperature change can be understood when the mean temperature is also taken into account. All the stations without a change in SPR have mean monthly surface temperatures above roughly 280K (Fig. 8) and as a consequence the vast majority of precipitation falls as rain (Fig. 3). Thus, atmospheric warming at these stations has had little effect on SPR. Cooling could have caused an increase in SPR, but none of these stations have cooled sufficiently for this to happen.

5 Causes of the snowfall decline

The summer snowfall decline has occurred in unison with decreases in SPR and total precipitation (Fig. 5). It is possible to quantitatively estimate the proportion of the snowfall decline that is associated with changes in precipitation form and that associated with changes in total precipitation. To separate the component of the snowfall decline due to changes in precipitation form, S_{form}, we held the total precipitation constant at the 1989 value (as given by the y-intercept of the linear trend) and estimated the snowfall from the time-varying SPR by:

$$S_{form} = P_{1989} * S/P \tag{4}$$

Conversely, to estimate the snowfall variability due to changes in total precipitation, S_{amount}, we held SPR constant at the 1989 value and estimated the snowfall from the time-varying total precipitation:

$$S_{amount} = S_{1989}/P_{1989} * P \tag{5}$$

In Fig. 9 (upper panel) the solid line denotes the summer-mean Arctic-mean snowfall in ERA-I, with S_{form} and S_{amount} shown by the dotted and dashed lines, respectively. The changes in total precipitation have a weak influence on the snowfall variability. Instead, the snowfall variability is almost entirely explained by changes in SPR, hence, changes in precipitation form. As discussed earlier, we expect this component of the snowfall change to be strongly dependent on temperature. The lower panel in Fig. 9 shows the summer-mean Arctic-mean 900hPa air temperature. Snowfall and air temperature are very highly correlated (r =

-0.92) and both display significant trends. Statistically, the trend in 900hPa air temperature explains over 99% of the snowfall trend.

Previous studies have demonstrated that the temperature trends in ERA-I are realistic (Screen and Simmonds, 2010a,b, in press) and Fig. 2 suggests that the temperature-snowfall relationships in ERA-I are also valid. Accordingly, we argue that snowfall changes related to changes in precipitation form (temperature changes) will be accurate in ERA-I. The accuracy of the component of snowfall change due to change in total precipitation, which is less dependent on temperature, is more uncertain. With this in mind, let us consider that possibility that ERA-I is incorrect and that Arctic precipitation has increased rather than decreased over the period 1989-2009. Given that snow represents a small fraction of the total precipitation in summer (approximately 15% on average, Fig. 3), it would take a relatively large increase in total precipitation to ameliorate the loss of snowfall due to changes in SPR. For example, ERA-I depicts a 0.1mm-we d^{-1} decrease in snowfall over the period 1989-2009 due to changes in precipitation form. To balance out this loss, the Arctic-mean summer precipitation would have had to have increased by around 6.6mm d^{-1}. To put this estimate into context, Arctic-mean summer precipitation would have had to have increased ten-fold in ERA-I between 1989 and 2009 to counter the loss of snowfall driven by changes in SPR. Observations suggest a large-scale precipitation increase of around 8% over the past century (Symon et al, 2004), although substantial uncertainties remain owing to lack of observations, particularly over the Arctic Ocean.

Therefore, we assume that the impact of changes in total precipitation on summer snowfall have been small and are negligible in comparison to the impacts of changes in SPR. In other words, we argue that the critical factor driving the summer snowfall decline is lower-tropospheric warming, which is reasonably well-represented in ERA-I, and not precipitation changes that may not be accurate in reanalyses. This assumption is unlikely to hold for other seasons, hence, our focus on summer. Summer is also of especial interest because the radiative impacts of changes in surface albedo are greatest in this season.

We now explore the implications of the summer snowfall decline and its role in recent Arctic climate change.

6 Impacts on surface albedo

It is well known that snow has a higher albedo than ice and that snow-covered ice has a higher albedo than bare ice. The surface albedo is also sensitive to characteristics of the snow cover, for instance whether the snow is wet or dry, and to the presence of surface melt ponds. Neither of these effects are directly considered here. ERA-I includes 6-hourly output of the surface albedo. The modeled albedo over ocean in ERA-I is dependent only on the sea ice concentration and the seasonally varying a_i. The model does not accumulate snowfall on ice. By mid-summer, all sea ice is considered to be bare ice with no snow cover or melt ponds. Thus, trends in the ERA-I albedo only relate to changes in sea ice cover. To assess the radiative impacts of changes in snowfall over the Arctic Ocean, we have performed a series of "nudging" experiments using ERA-I output. The premise of these experiments is to represent changes in snow-covered ice and their impacts on albedo and surface solar radiation using simple, but physically reasonable, parameterizations. The outputs from these "nudged" experiments are compared to the unmodified

ERA-I output to quantify the relative importance of changes in snowfall on the surface radiation budget, in comparison to changes in sea ice concentration.

Using daily output from ERA-I, grid-boxes were identified that had snowfall. In these grid-boxes the albedo was nudged according to the following parameterization:

$$a_{nud} = a_{era} + c(a_s - a_i), \tag{6}$$

where a_{nud} is the nudged albedo, a_{era} is the model albedo from ERA-I (from Eq. 1), and a_i is the seasonally-varying albedo of sea ice (Fig. 1). The albedo of snow-covered ice, a_s, was set as 0.75, in line with the observed albedo of wet (melting) snow (Curry et al, 2001; Perovich et al, 2002). Snowfall over open water, which has no direct impact on the surface albedo, was accounted for by scaling the nudge factor by the ice cover fraction, c. The nudging was only applied to grid-boxes with snowfall on that day and the albedo returns to the ERA-I value the day after snowfall ceases. This is consistent with observations in summer, which suggest that the snow cover rapidly melts following a snowfall event (Perovich et al, 2002). We assume that all ice within a grid-box is snow-covered when snowfall occurs in that grid-box and on that day. Note, the nudged albedo does not consider snow depth, with the grid-box being considered snow-covered or not depending on whether there was snowfall on that day or not.

Fig. 10 illustrates the effect of the parameterization during June and July 1989 at a grid-box in the Barents Sea. As the ice cover decreased during June and early July, the ERA-I albedo also decreased. By mid-July, the grid-box was ice free and the albedo was 0.06 (a_w). Superimposed on this variability due to ice cover, is a small decline in albedo during June due to prescribed seasonal cycle of a_i (see Fig. 1). The nudged albedo includes the effects of sea ice cover, but additionally it increases relative to the ERA-I albedo on days with snowfall (if there is sea ice present). The nudge factor (the difference between the ERA-I and nudged albedo) on snowy days is dependent on c, reflecting the increased likelihood of snow falling over open water rather than sea ice as c decreases, and on the difference between a_i and a_s. When c is zero, snowfall events result in no change in albedo (as seen on 30 July in Fig. 10).

The sea ice area has significantly declined in summer over the last 21 years (Fig. 11). The area of ice assumed to be snow-covered has also decreased significantly. The latter decline has occurred at a faster rate, resulting in a decrease in the fraction of ice covered by snow. Thus, not only has the ice cover declined exposing open water, the snow cover on top of the ice has declined exposing more bare ice. Both of these changes will have an effect on the surface albedo. The questions we address now are how large are these effects and how important is the decline in snow-covered ice relative to the more widely documented effects of reduced sea ice cover.

Fig. 12 (top) shows time-series of the Arctic-mean summer albedo from ERA-I and our nudged experiment. The former varies according to the sea ice cover (compare with Fig. 11, top) and shows a significant decline over the period 1989-2009 (Fig. 12, bottom). Relative to the ERA-I albedo, the nudged albedo is higher throughout the period reflecting the allowance for the presence of snow-covered ice. The difference between the ERA-I and nudged albedo decreases as a function of time because the fraction of ice covered by snow decreases. As a result, the nudged albedo shows a larger decline than the ERA-I albedo over the 21 years.

The difference between the linear change in the ERA-I and nudged albedo, reflects the albedo change due solely to changes in the fraction of ice covered by snow. The parameterised changes in snow-on-ice lead to an albedo decrease of 0.03. This change is comparable in magnitude to the albedo decrease in ERA-I, that is solely due to changes in sea ice cover.

We now assess the impact of these albedo changes on the surface radiation budget. In ERA-I, the net surface solar radiation, q_{era}, can be written as:

$$q_{era} = q_{in}\left[(c(1-a_i) + (1-c)(1-a_w)\right], \tag{7}$$

where q_{in} is the surface incoming solar radiation. For grid-boxes where the ice is assumed to be snow-covered, the net surface solar radiation response to the nudged albedo, q_{nud}, becomes:

$$q_{nud} = q_{era} + q_{in}\left[c(a_i - a_s)\right], \tag{8}$$

The nudging only occurs over the ice-covered portion of the grid-box, so (7) reduces to:

$$q_{era} = q_{in}(1 - a_i), \tag{9}$$

Rearranging (9) and substituting into (8) gives:

$$q_{nud} = q_{era} + \frac{q_{era}}{1-a_i}[c(a_i - a_s)] = q_{era}[1 + \frac{c(a_i - a_s)}{1-a_i}], \tag{10}$$

Fig. 13 shows time-series of the Arctic-mean summer net surface solar radiation under clear-sky conditions, in ERA-I and our nudged experiment. Here we use the clear-sky radiation rather than the all-sky radiation in order to remove the effects of cloud cover and changes in cloudiness. This isolates the solar radiation response to changes in albedo. Furthermore, it circumvents concerns about the reliability of cloud cover and its trends in reanalysis products. Of course, changes in albedo have a greater influence on the surface solar radiation under clear skies than under cloudy skies. Thus, by using the clear-sky radiation we are clearly overestimating the surface solar radiation response and our estimates must be viewed as an upper boundary of the radiation change due to the changing albedo over the Arctic Ocean. However, our primary concern here is not the absolute magnitude of the solar radiation change but the relative magnitudes of the response due to changes in sea ice cover and the response due to changes in snow-covered ice. Consistent with the decrease in albedo (Fig. 12), ERA-I depicts a significant increase in the net surface solar radiation over the period 1989-2009 (Fig. 13). Including the effects of changing snow-covered ice results in reduced solar energy input throughout the period, in line with the higher albedo versus the case with no representation of snow-covered ice. The difference between the ERA-I and nudged radiation decreases with time as the proportion of ice covered by snow decreases. The solar radiation change due to the decline in snow-covered ice is slightly larger than that associated with the change in sea ice cover, again pointing to the importance of changes in snow-covered ice for the ice-albedo feedback.

We estimate an upper-bound for the solar radiation change due to the decline in snow-covered ice of 21.4 W m^{-2} (Fig. 13). We have repeated the analyses using the ERA-I all-sky solar radiation (not shown). In this case, the changes in snow-covered ice result in a 11.8 W m^{-2} increase in solar heating over the period 1989-2009, approximately half of the increase under clear-sky condition. Finally, we estimate the amount of surface melt that could be sustained by these increases in solar energy input to the ice cover:

$$\Delta I = \frac{\Delta Q * t}{L_f * \rho_{ice}}, \tag{11}$$

where ΔI is the depth of surface ice melted (m), ΔQ is the change in net solar radiation (W m^{-2}), t is the elapsed time, L_f is the latent heat of fusion (334000 J kg^{-1}) and ρ_{ice} is the density of ice (917 kg m^{-3}). If sustained over the summer (t = 3 months) over the ice-covered ocean (recall the parameterized solar radiation change is explicitly over ice), ΔQ would result in surface melt of 0.56m or 0.31m, for the clear-sky and all-sky estimates, respectively. Whilst these rough estimates mask a large degree of spatial variability, they nonetheless represent a sizable fraction of the ice thickness in summer. Thus, the decline in summer snowfall may have significantly contributed to the thinning of sea ice over recent decades.

The experiments presented highlight the sensitivity of Arctic climate to the estimated recent changes in summer snowfall. Whilst they are not expected to fully represent reality due to the omission of a number of other factors that influence the ice albedo, such as melt ponds and ice thickness, our aim was to estimate and isolate the effects of changing snow-on-ice. It is important to note, that although ERA-I has no allowance for time-varying changes in snow-on-ice, this need not imply that its output is erroneous because of this. ERA-I output is a combination of observations and model-estimated fields. If the observational constraint is high, the biases or errors in the model physics will have little effect. A case in point may be the ERA-I surface temperature trends. Based on the discussion above, one could reasonably hypothesize that because ERA-I does not account for changes in snow-cover, it may underestimate the warming over recent decades. However, this does not appear to be the case (Screen and Simmonds, 2010a,b, in press). This may reflect the fact that surface temperatures (and sea ice concentrations) are relatively highly constrained by satellite observations over the period considered. Problems are more likely to arise in variables with weak (or no) observational constraint, for example the radiative heat fluxes, but this is hard to confirm. Finally, we note that the errors in surface albedo are likely to be one of many sources of error in the ERA-I model physics, and the net effect of all these errors will likely be wholly different to those we have isolated here.

7 Feedbacks

The results suggest that increasing Arctic temperatures have led to decreased snow-on-ice (Fig. 11), that has decreased the surface albedo (Fig. 12) and increased energy gain by the ocean-ice-atmosphere system (Fig. 13). A logical next question to ask is: do these changes constitute a positive feedback on Arctic warming? This question is harder to answer than it may first appear. Increased energy gain by the Arctic Ocean during summer, as a result a lower albedo, is associated with increased energy transfer from the ocean to the atmosphere in autumn and winter (Screen and Simmonds, 2010b). Some of this energy will be lost to space as longwave radiation, but a proportion will warm the atmosphere. So, in isolation the aforementioned changes represent a positive feedback on Arctic warming. However, it has been previously noted that the interpretation of a feedback depends on the temporal scale of the changes under consideration, and it is essential to consider how the feedback mechanism operates when integrated through at least a full annual cycle (Curry et al, 1995). This view recognizes the fact that linkages that constitute a positive feedback in one season may not constitute a feedback, or represent a negative feedback loop, at another time of the year. Two of the linkages in the temperature-snowfall-ice feedback are characterized by competing effects (Francis et al, 2009) and their relative importance vary by season.

Snowfall, or more precisely snow cover, has competing effects on the sea ice. On one hand, the snow cover increases the surface albedo, reducing energy absorption and decreasing ice melt. Since some snow gets converted to ice, more snow also increases ice formation rates. On the other hand, the snow cover insulates the ocean from the atmosphere. In the colder seasons, this reduces oceanic heat loss and ice growth (Ledley, 1991, 1993). The albedo effect is greatest is summer when insolation is greatest, whereas the insulation effect is largest in the ice-growth season.

The linkage between changes in air temperature and snowfall can also operate in both senses. Warmer air temperatures may be associated with increases or decreases in snowfall. Here we have shown that in summer, warming leads to reduced snowfall due to changes in SPR. Considering the cold season, increasing temperatures are expected to led to more precipitation because higher air temperatures can hold more water vapor. Since most precipitation in cold season falls as snow, this would translate to an increase in snowfall. However, total precipitation is influenced by the complex interplay of numerous factors in addition to atmospheric warming, for example, changes in cyclone activity (Figs. 5 and 6).

In summer, the albedo effect dominates over the insulation effect (Ledley, 1991, 1993) and warming is likely associated with decreased snowfall (Fig. 9) - a positive feedback. If snowfall was to increase due to autumn warming, the thicker snow cover may reduce ice growth and led to a thinner sea ice cover - again, a positive feedback. If autumn snowfall was to decrease, due to warming-induced changes in SPR, the feedback could be negative in this season. Another consideration is that increased winter or spring snowfall could result in the delayed melt of the snow cover, a higher albedo in spring and reduced ice melt - a negative feedback. This effect could be countered by spring warming that would promote earlier snow melt. In short, the net climatic effect of the temperature-snowfall-ice feedback is dependent on the sign, magnitude and timing of warming-induced snowfall changes, and their interactions with other aspects of Arctic change. Whilst the results presented here provide support for the existence of a positive feedback in association with the summer snowfall decline, the net climatic effect of past (and projected) snowfall changes in all seasons remains uncertain.

8 Conclusions

Our main conclusions can be summarized as:

1. The fraction of Arctic summer precipitation occurring as snow has declined over the last two decades.
2. As a result of (1), summer Arctic snowfall has declined by 40% over the period 1989-2009.
3. (1) and (2) are primarily due to lower-atmospheric warming.
4. (1) and (2) have significantly reduced the area of snow-covered ice during summer.
5. (4) has led to a substantial decrease in the Arctic-mean surface albedo.
6. (5) has likely contributed to the recent thinning of Arctic sea ice.

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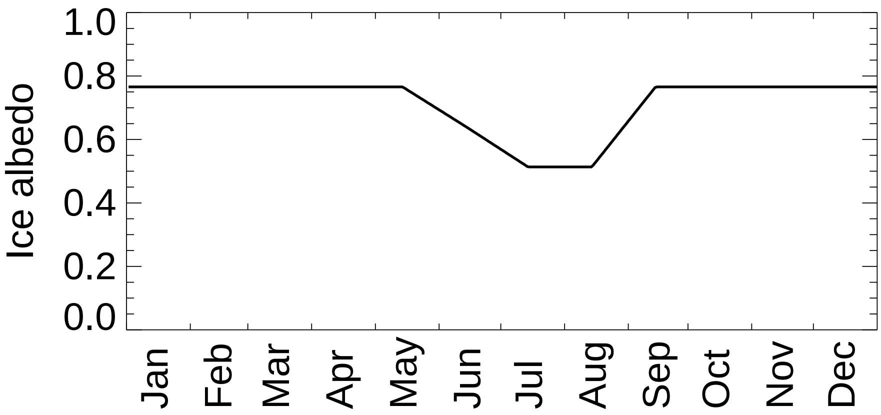

Fig. 1 Seasonally varying ice albedo in ERA-I.

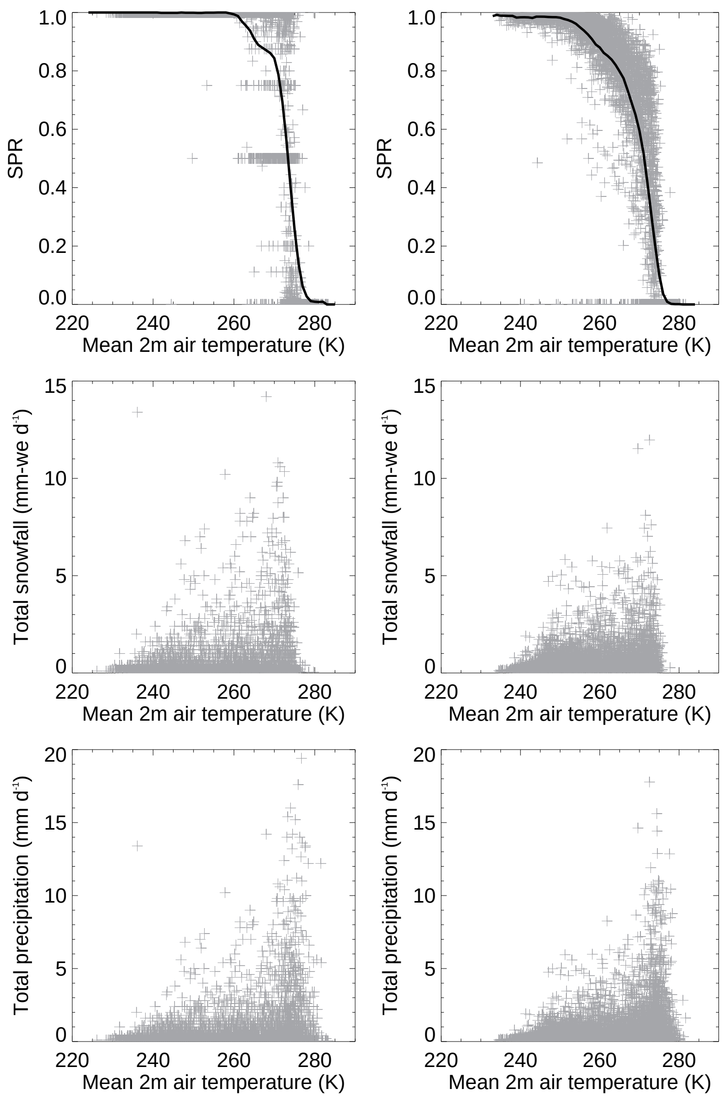

Fig. 2 (top left) Daily snowfall-to-preciptation ratio (SPR) as a function of mean surface air temperature at Resolute and (top right) in ERA-I sub-sampled at the grid-box containing Resolute. Each cross denotes a day in the period 1989-2009. Days with no precipitation are not plotted; 81% of the days considered were "wet" at Resolute and 74% in the ERA-I grid-box. The black line shows the mean SPR calculated for 1K bins and smoothed with a boxcar average of 5-bin-width. (middle) As top, but for daily snowfall and (bottom) for daily total precipitation

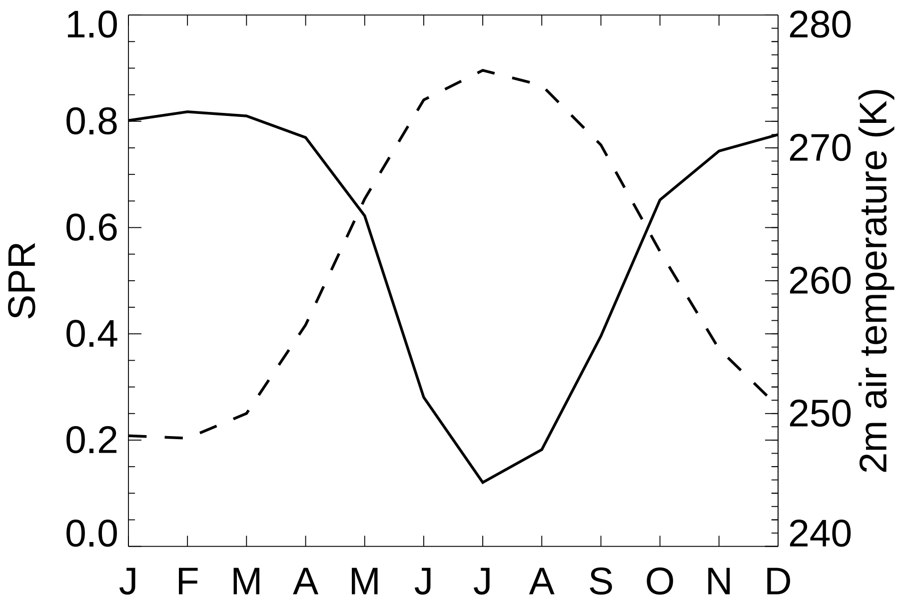

Fig. 3 Monthly climatologies of the Arctic-mean (north of 70°N) SPR (solid) and surface air temperature (dashed) in ERA-I, 1989-2009.

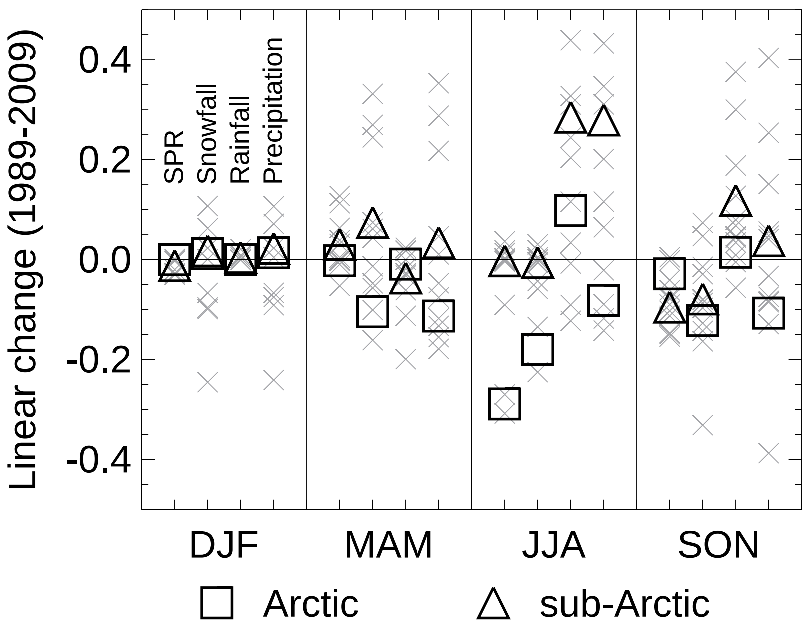

Fig. 4 Linear changes from 1989 to 2009 in seasonal-mean SPR, daily snowfall (mm-we d^{-1}), daily rainfall (mm d^{-1}) and daily total precipitation (mm d^{-1}) at the Canadian meteorological stations. Each gray cross represents the 21-year change at one station. Multi-station means for the Arctic (north of 70°N) and sub-Arctic (60-70°N) are shown by the squares and triangles, respectively. Note that in summer, two sub-Arctic stations have large rainfall and precipitation increases that fall outside the upper bound of the vertical scale and are not plotted, but are included in the sub-Arctic mean.

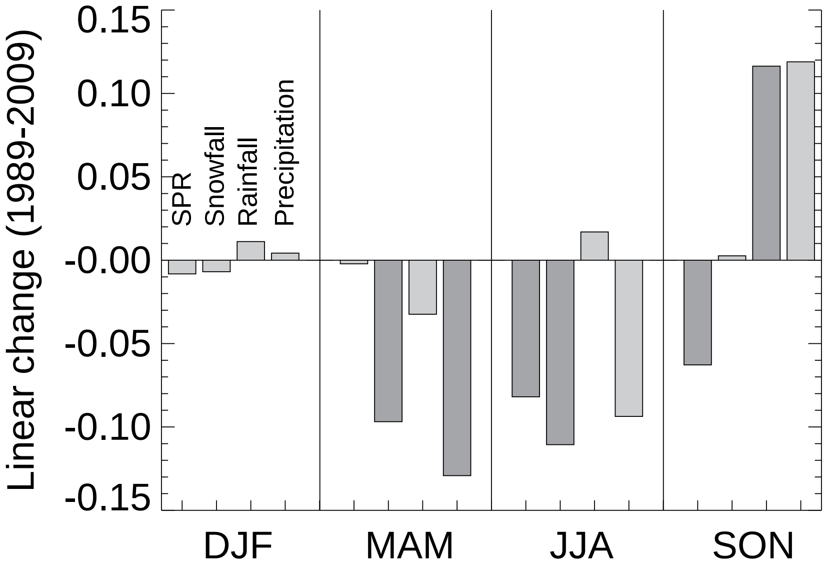

Fig. 5 Linear changes from 1989 to 2009 in seasonal-mean Arctic-mean SPR, daily snowfall (mm-we d^{-1}), daily rainfall (mm d^{-1}) and daily total precipitation (mm d^{-1}) in ERA-I. Darker bars denote linear changes that are statistically significant at the 90% level or better.

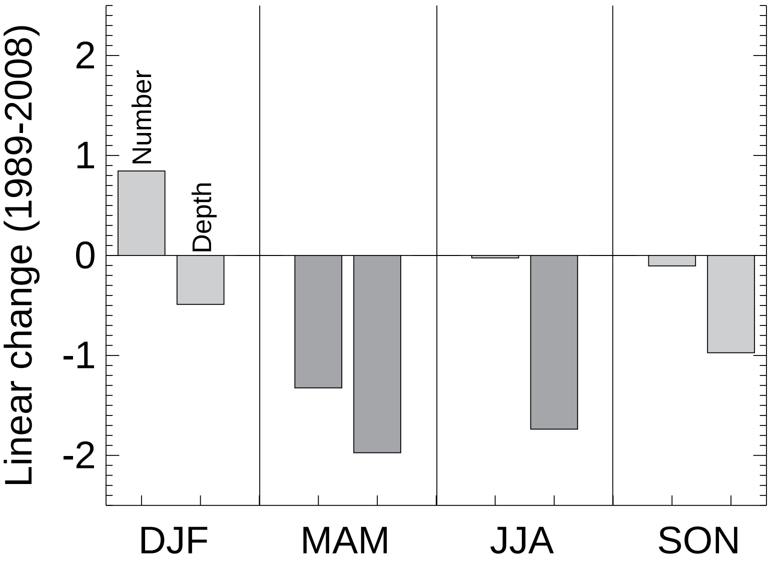

Fig. 6 Linear changes from 1989 to 2009 in seasonal-mean Arctic-mean cyclone number and mean cyclone depth in ERA-I. Darker bars denote linear changes that are statistically significant at the 90% level or better. For ease of plotting on the same vertical scale, both cyclone variables have been normalized by their standard deviation.

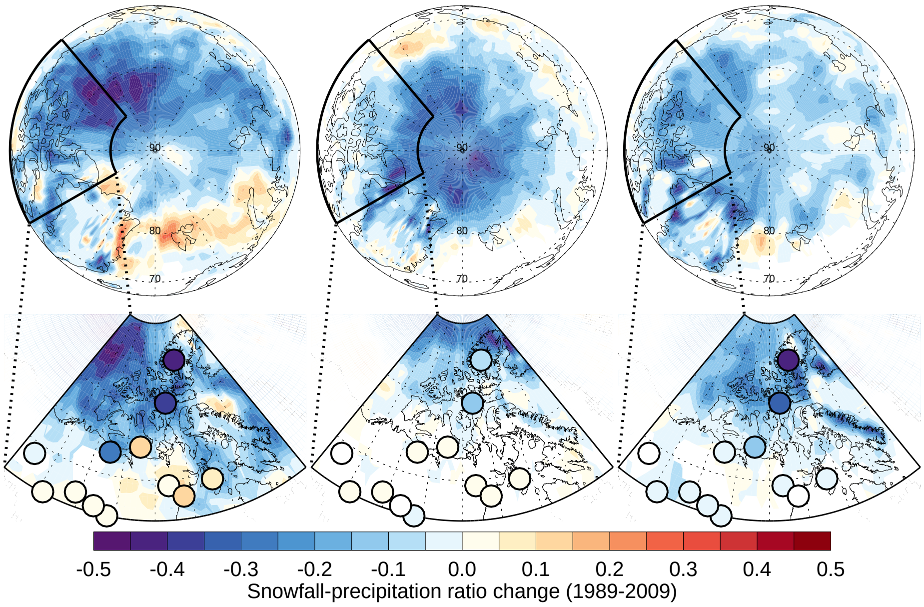

Fig. 7 Linear changes from 1989 to 2009 in (left) June-, (middle) July- and (right) August-mean SPR in ERA-I.. The colored dots denote SPR changes from Canadian meteorological stations. There is a change in map projection from satellite-view in the upper maps to polar stereographic in the lower maps.

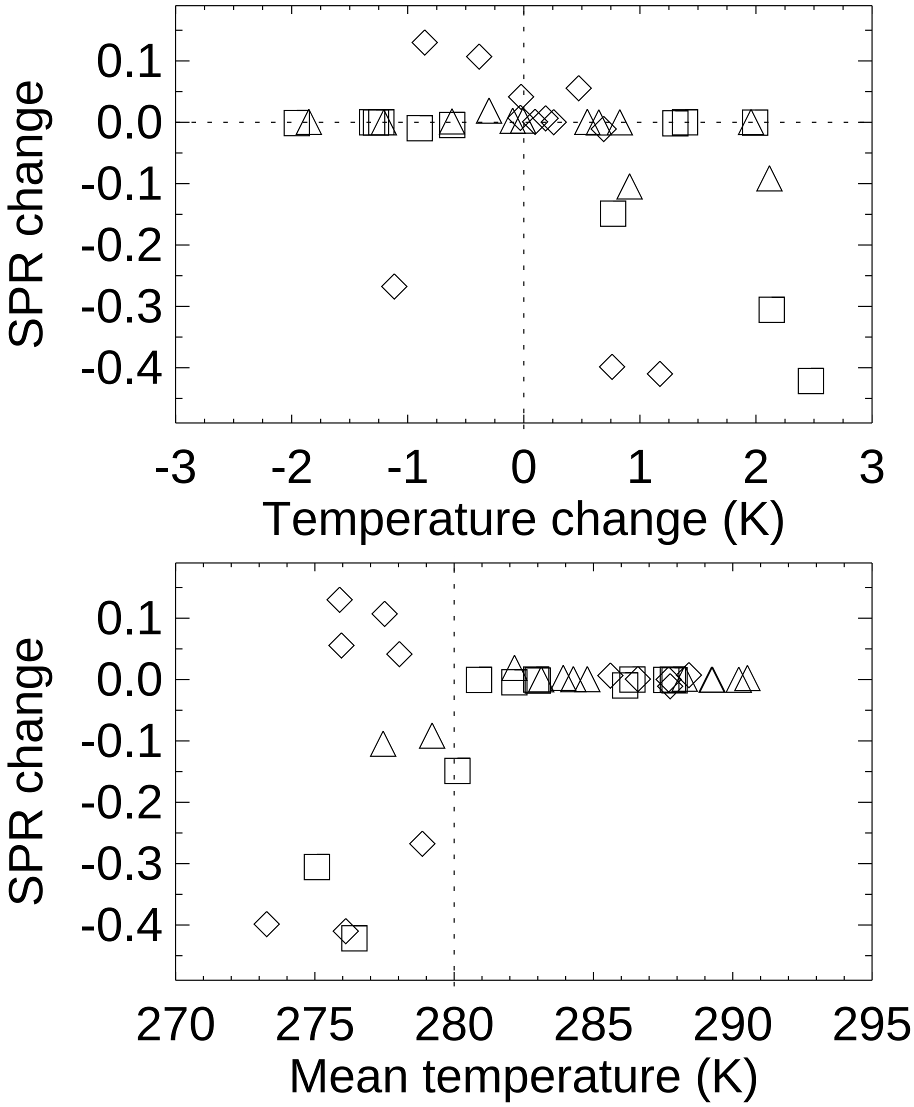

Fig. 8 Monthly-mean SPR change from 1989 to 2009 as a function of (top) mean surface temperature change and (bottom) mean surface temperature at the Canadian meteorological stations during the summer months (diamonds, triangles and squares denote June, July and August, respectively).

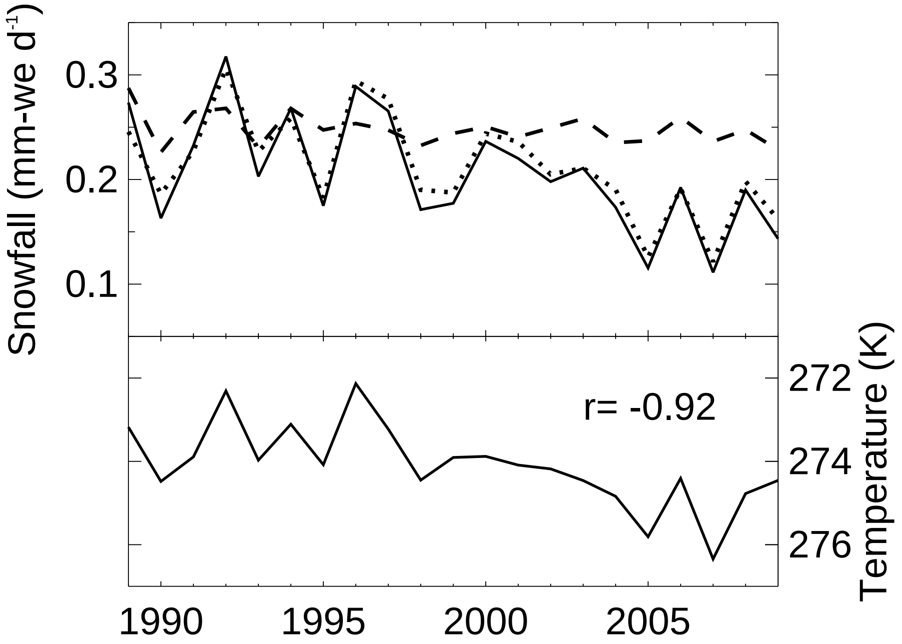

Fig. 9 (top) Time series of Arctic-mean summer-mean daily snowfall (solid). The dashed and dotted lines denote the estimated snowfall due to changes in precipitation or changes in SPR, respectively. (bottom) Time series of Arctic-mean summer 900hPa air temperature (note the inverted scale).

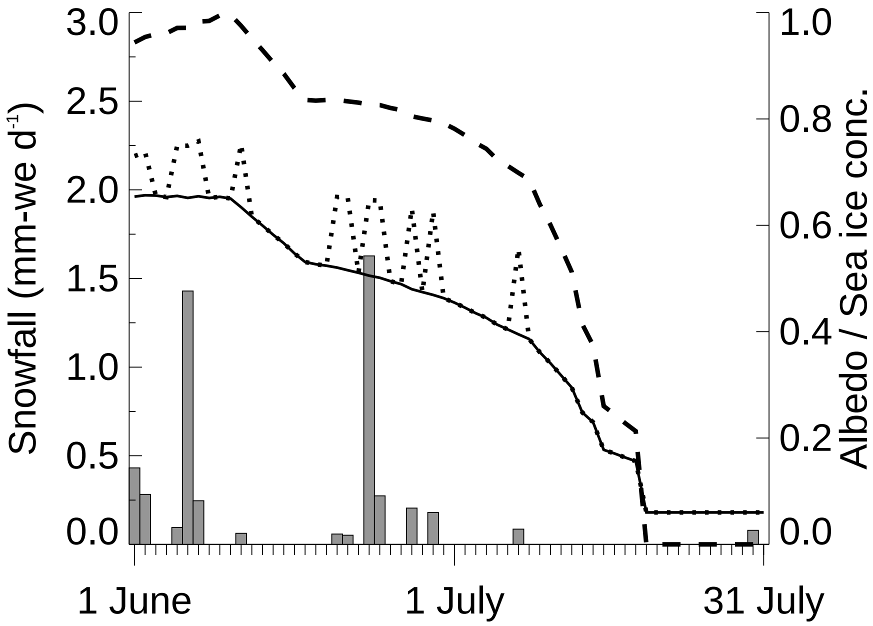

Fig. 10 Daily mean snowfall during June-July 1989 from an ERA-I grid-box in the Barents Sea. The dashed line denotes the daily sea ice fraction. The solid and dotted lines denote the albedo from ERA-I and the nudged experiment, respectively.

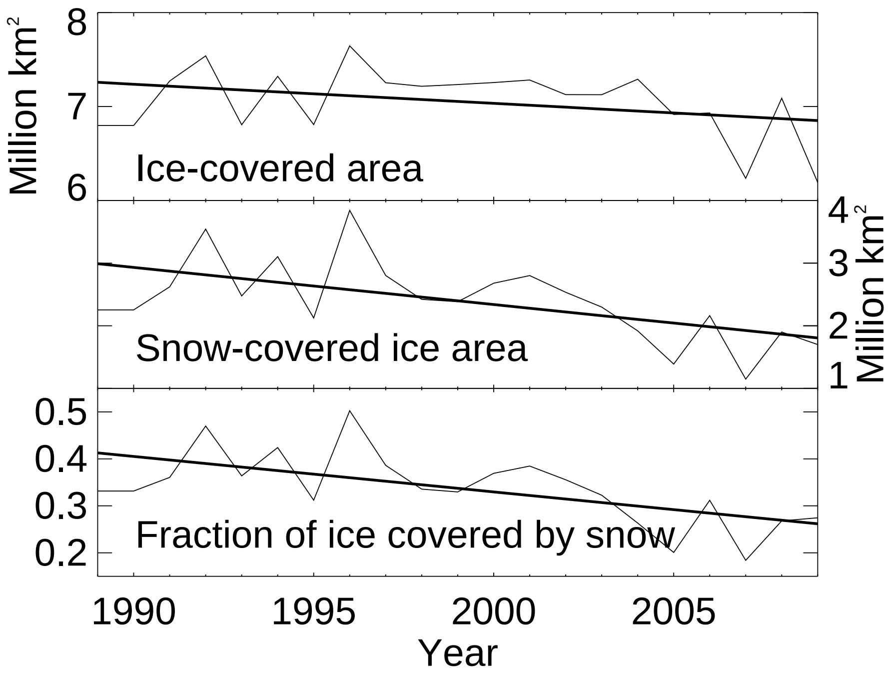

Fig. 11 Time series of summer (top) sea ice area, (middle) snow-covered ice area and (bottom) the fraction of ice covered by snow.

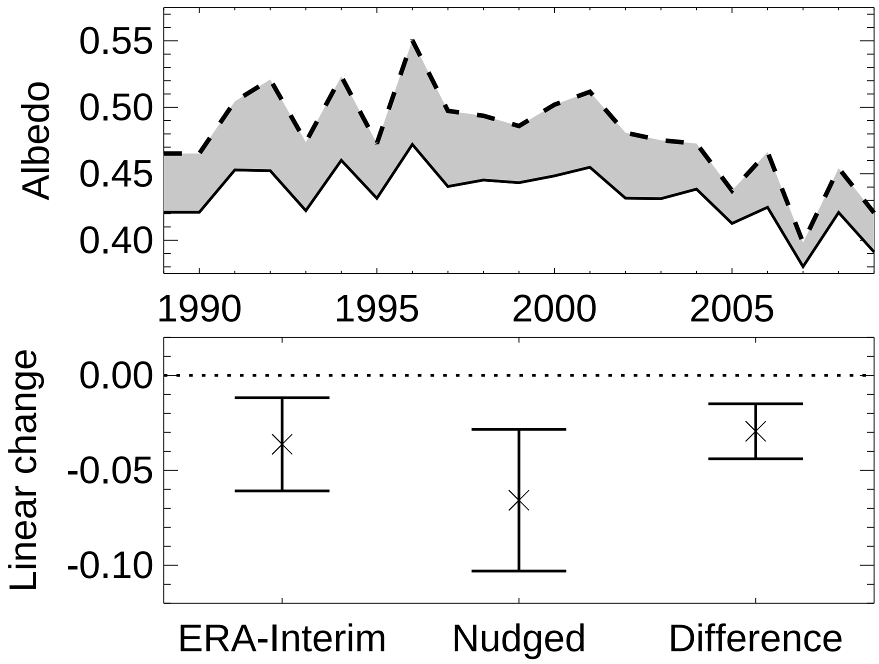

Fig. 12 (top) Time series of Arctic-mean summer albedo from (solid) ERA-I and (dashed) the nudged experiment. (bottom) Linear change from 1989 to 2009 in the Arctic-mean summer albedo in ERA-I and the nudged experiment, and their difference.

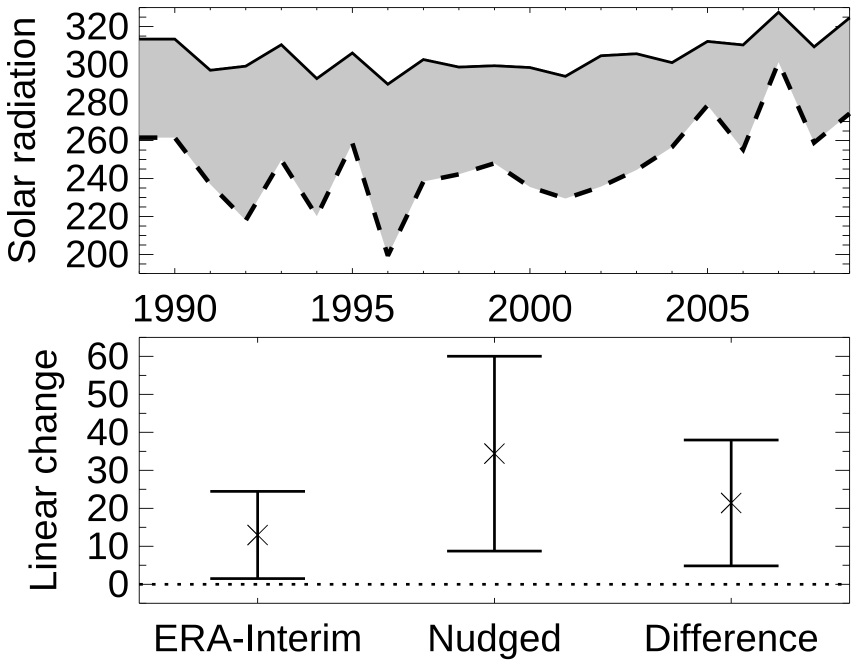

Fig. 13 top) Time series of Arctic-mean summer net surface clear-sky solar radiation (W m^{-2}) from (solid) ERA-I and (dashed) the nudged experiment. (bottom) Linear change from 1989 to 2009 in the Arctic-mean summer albedo in ERA-I and the nudged experiment, and their difference.