Five decades of warming: impacts on snow cover in Norway

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ABSTRACT

Northern latitudes are experiencing faster warming than other regions in the world, which is partly explained by the snow albedo feedback. In Norway, mean temperatures have been increasing since the 1990s, with 2014 being the warmest year on record, 2.2 °C above normal (1961–1990). At the same time, a concurrent reduction in the land area covered by snow has been reported. In this study, we present a detailed spatial and temporal (monthly and seasonal) analysis of trends and changes in snow indices based on a high resolution (1 km) gridded hydro-meteorological dataset for Norway (seNorge). During the period 1961–2010, snow cover extent (SCE) was found to decrease, notably at the end of the snow season, with a corresponding decrease in snow water equivalent except at high elevations. SCE for all Norway decreased by more than 20,000 km² (6% of the land area) between the periods 1961–1990 and 1981–2010, mainly north of 63°N. Overall, air temperature increased in all seasons, with the highest increase in spring (particularly in April) and winter. Mean monthly air temperatures were significantly correlated with the monthly SCE, suggesting a positive land–atmosphere feedback enhancing warming in winter and spring.

Key words | Norway, snow cover, snow water equivalent, trend, warming

INTRODUCTION

Seasonal snow cover is a fundamental part of the cold climate water balance by storing water during winter until the snowmelt season (Wilson et al. 2010; Irannezhad et al. 2015) and of the Earth’s energy balance by controlling the amount of reflected solar (shortwave) radiation through the high albedo of snow (Betts 2000; Dery & Brown 2007). Snow has major societal and economical importance in regions with a seasonal snow cover, affecting sectors such as hydropower production, agriculture, forestry, recreation and tourism. In the Northern Hemisphere, snow covers up to 30% of the global land area in winter (Vaughan et al. 2013), and its largest influence on the climate is during spring when the day length and solar radiation increase and snow starts to melt (NSIDC 2016). Recent changes in temperature (general warming) and precipitation (both increases and decreases) have altered snow conditions over the northern hemisphere, reducing both the fraction of precipitation falling as snow and the duration of the snow season (Dery & Brown 2007; Callaghan et al. 2011); the exception being some high elevation regions, where larger snow volumes have been observed due to an increase in winter precipitation, as reported by Dyrrdal et al. (2013) and Skaugen & Randen (2013) for Norway. A reduction in the snow cover, and thus in surface albedo, may enhance the warming signal because more energy is absorbed at the surface. This positive feedback to global warming in regions with a seasonal snow cover is thought to partially explain the accelerated warming over the Northern Hemisphere.
(Serreze et al. 2009). Observed Arctic warming has been more than twice as large as the global average (Cohen et al. 2014). Chapin et al. (2005) described and quantified the terrestrial feedback and found a 3.3 Wm$^{-2}$ increase in absorbed energy at the surface because of advancing snow-melt, a much greater effect than that of vegetation changes.

Snow cover extent (SCE) in the northern hemisphere has decreased since the late 1960s, with an accelerated rate from 2003 (Derkelsen & Brown 2012). This is reflected in a large shift of the start and end of the snow season, with a stronger change in spring than in autumn (Peng et al. 2015) and with larger decreases in maritime regions north of 60° N (Callaghan et al. 2011).

The climate of Norway shows large regional as well as local variations spanning over 13° of latitude and climate zones ranging from maritime mild temperate to arctic. Norway has a rugged topography ranging from coasts to high mountains over small distances. This makes Norway an interesting cold climate laboratory that provides a unique opportunity to study changes and trends in snow cover and interactions with climate over a wide range of conditions. Such spatial and temporal differences may help identifying the main processes governing observed changes in Norway and provide relevant insights for similar trends in other regions with comparable conditions.

The climate of Norway is warming. Temperature trends for the period 1900–2014 have been found to be statistically significant for all seasons except winter (with a maximum trend of 0.13°C/decade in spring) (Forland et al. 2016), with 2014 the warmest year in the record (2.2°C above normal, 1961–1990; Gangstø et al. 2016). The strongest temperature trends have been observed in northern Norway (Hanssen-Bauer et al. 2017). Forland et al. (2016) reported a larger temperature trend for the period 1955–2014 than for the longer 1900–2014 period, at 12 lowland stations located across the country. For this shorter period, they found the greatest warming in winter and spring for most stations, with up to 0.63°C/decade in winter at Gardermoen in southeastern Norway. Mean annual precipitation shows an overall increase of 18% in the period 1900–2014 (linear trend), with the largest increase seen during autumn (eastern and southern Norway) and spring (west and north) and a less pronounced increase during summer (Hanssen-Bauer et al. 2017).

As climate is warming, impacts are felt on the ground. Previous studies have found decreasing trends in several snow indices such as snow depth, SCE and the length of the snow season across Norway (e.g., Dyrrdal 2009; Dyrrdal & Vikhamar-Schuler 2009; Dyrrdal et al. 2013). For the period 1981–2010, Dyrrdal et al. (2015) found a decline in snow depth at elevations lower than 1,000 m a.s.l. and in coastal regions (particularly at the southwestern coast). On the other hand, increasing snow depth has been observed for inland mountain stations in the period 1961–2010, due to increasing winter precipitation (Dyrrdal et al. 2013). A significant decrease in the number of days when precipitation falls as snow was found in half of the 585 meteorological stations studied by Dyrrdal (2009) over the period 1968–2007, particularly after 1990. Similarly, the highest percentage of significant negative trends was found in southern Norway (mainly along the coast). Furthermore, the findings of Dyrrdal et al. (2013) point towards an accelerated decline in snow depth during the last decades (i.e., 1961–2010). The above studies, overall, demonstrate a clear decline in snow cover. At the same time, the results also demonstrate a large sensitivity to the selected time period, the region under investigation and the method. Thus, care should be taken when comparing across studies, regions and periods.

The studies referred to above are based on observations from selected stations across Norway, and vary in the number of stations included, spatial coverage, time period chosen and methodology used to detect and map the changes. Thus, there is a need for a more complete view of all Norway to highlight regional differences and drivers of change. The availability of high-resolution observation-based gridded data makes this possible.

In this study, we perform a detailed analysis of trends and changes in snow climatology based on a gridded (1 × 1 km) hydro-meteorological dataset for Norway (seNorge; Tveito et al. 2005) covering the period 1961–2010. It is postulated that higher temperatures lead to less snow accumulation during autumn and winter, which when combined with higher winter and spring temperatures, produces an earlier snowmelt. This, in turn, may lead to accelerated warming (i.e., higher temperatures) in spring due to snow albedo feedbacks (see Groisman et al. 1994; Dery & Brown 2007).
The study is designed to assess the validity of this hypothesis by measuring process signals through:

1. a temporal analysis, estimating annual, seasonal and monthly changes in snow indices and related climate variables (i.e., precipitation and temperature) for all Norway (national scale) and two regional scales (i.e., 19 climatological regions and four aggregated macro regions; see next section);
2. a regional analysis, evaluating spatial patterns of changes and correlation between climate variables and snow indices for different spatial scales, ranging from the grid cell scale to the national level;
3. a comparative analysis of temperature trends and changes in snow indices.

The study adds to previous work by performing a detailed and comparative analysis of temporal and spatial change patterns in snow climatology based on daily time series from 1961 to 2010. The time series analysed include six snow indices as well as mean daily temperature and daily precipitation totals. Four grid cell (local) indices were derived from seNorge: snow cover area (SCA), snow water equivalent (SWE), duration of the snow season, and precipitation falling as either snow (snowfall) or rain (rainfall). In addition, two indices derived at the regional scale were used: SCE, and regional SWE derived over a region or all Norway (regSWE). Norway was separated into 19 prevailing climatological regions following Dyrrdal et al. (2011), successively grouped into four macro regions. We derived changes in 10-year (decadal) running mean (monthly and seasonal) and three overlapping 50-year periods (1961–1990, 1971–2000, 1981–2010) for the selected snow indices for all Norway as well as for the separate climatological regions. When comparing across periods, 1961–1990 was used as a reference period.

**STUDY AREA AND DATA**

**Climate regions and snow climatology**

The Norwegian mainland area covers 323,781 km² and features a long north–south coastline and a mountain chain stretching across the Scandinavian Peninsula (Figure 1). Western Norway is characterized by a complex terrain due to large elevation gradients over small distances, whereas inland regions in eastern Norway have a more undulating landscape.

Despite Norway’s high latitudes, the coastal climate is mild and wet (maritime) due to the influence of the North Atlantic Current and the prevailing southwesterlies, bringing mild, maritime air on shore (Aune et al. 1993). Precipitation is dominated by frontal storm events most of the year, with additional convective precipitation in inland regions during warm months (Hanssen-Bauer et al. 2017). The highest air temperatures are found close to the coast in southern Norway, whereas the lowest are found in the mountains and the far north (Hanssen-Bauer et al. 2017) (Figure 1).

Approximately 30% of the annual precipitation in Norway falls as snow (Dyrrdal et al. 2013). The geographic distribution of snow is highly irregular in regions of complex topography because of large variations in precipitation and redistribution by wind. The largest amount of snow during the period 1961–1990 was recorded in southwestern Norway, where the mean annual maximum SWE may exceed 1,000 mm. The lowest SWE values are seen along the coast (<100 mm). The duration of the snow season varies accordingly, from less than a month along the coast in southern Norway to more than 9 months at high elevations.

In this study, we considered 19 regions defined by Dyrrdal et al. (2012), as shown in Figure 1. This regionalization is based on 15 precipitation (Hanssen-Bauer & Fröland 1998) and six temperature regions (Hanssen-Bauer & Nordli 1998), later modified by Dyrrdal et al. (2012) to separate highlands (higher than 1,000 m a.s.l.) from lower-lying regions (e.g., region 8 was split into regions 8.1 and 8.2). The 19 regions were further aggregated into four larger groups, referred to as macro regions. The four macro regions are labelled according to their location/characteristics, and in the paper we refer to them using their names with a capital letter. The Mountain macro region (composed of the high-elevation parts of five regions) corresponds to areas above 1,000 m a.s.l. in southern Norway; the South West macro region aggregates four regions along the western coast (from Trondheim to Kristiansand; Figure 1); the South East macro region combines five regions located on the east side of the Mountain macro region; and the North macro region aggregates the five northernmost regions (north of Trondheim).
The seNorge dataset

The seNorge dataset is a gridded hydro-meteorological dataset for Norway at 1 km spatial resolution (Tveito et al. 2005), tracking SWE, SCA, precipitation and temperature. Regular meteorological observations (from the Norwegian Meteorological Institute) are interpolated to the grid cell scale by Optimal Kriging interpolation (Tveito et al. 2005; Mohr 2008; Saloranta 2014a). Spatially interpolated values of temperature and precipitation are then used as input to NVE’s operational hydrological model (HBV), which includes a snow model (Saloranta 2014b). Within the snow
model, two modules are run in sequence: (i) the SWE model for snow pack water balance, based on the original snow routine in the HBV model (Sælthun 1996) and (ii) the snow compaction and density module converting SWE into snow depth (Saloranta 2012). A further refinement was later added to the snow compaction module to simulate the spatial variability of SWE and SCA within each grid cell (Saloranta 2014b). Studies evaluating the snow model against observations have shown that the SWE and snow depth agree well (Saloranta 2014a, 2014b). The seNorge dataset covers a total area of about 323,000 grid cells over land. Data are available from 1st September 1957 until the present (updated daily).

**Snow indices**

The snow indices consist of four local grid cell indices and two derived regional indices (Table 1). Local indices for SCA and SWE are taken directly from the seNorge database. The duration of the snow season (D), is derived according to the definition by Dyrrdal & Vikhamar-Schuler (2009), who classified the SCA of each grid cell into five classes (with five different snow cover codes) based on a qualitative description of the amount of snow cover on the ground (Table 2). Following this scheme, we introduced quantitative SCA thresholds (Table 2) for each code based on the SCE in the grid cell following the definition from Dyrrdal & Vikhamar-Schuler (2009). Subsequently, the first and last day of the snow season are identified as follows:

- **Start of snow season:** first day in autumn with 10 consecutive days classified with code 4 (i.e., SCA ≥ 95%);
- **End of snow season:** the first day in spring after the last period of 5 consecutive days classified with a code equal to or less than 2 (i.e., ≤ 50%).

Daily rainfall and snowfall rate were derived by apportioning precipitation into snowfall and rainfall using a temperature threshold of 0.5°C (Saloranta 2012).

The regional indices include SCE and regional average snow water equivalent (regSWE). SCE is the total area covered by snow (in km²) for a region or all Norway, whereas regSWE is the average SWE (in mm) over all grid cells with snow. SCE is calculated by multiplying the fraction (%) of the SCA with the total area of each grid cell (1 km²) and then summing the snow covered area over all grid cells in the region or country.

### METHODS AND ANALYSIS

Daily time series of snow indices and climate variables were aggregated to the monthly and seasonal time scales before analyses for five decades between 1961 and 2010 and three partially overlapping 30-year periods (1961–1990, 1971–2000, 1981–2010), with 1961–1990 representing the reference period. Four seasons were defined: winter (DJF), spring (MAM), summer (JJA) and autumn (SON). Changes in these time series of snow indices were assessed for each grid cell, regional values, and all Norway by analysing:

- **Decadal (10-year) running means for seasonal and monthly snow indices (SCE and SWE) covering the full period (1961–2010; Figure 2);**

### Table 1 | Snow indices used in the study, separated into local (grid cell) indices and regional indices (derived from grid cell values)

<table>
<thead>
<tr>
<th>Grid cell index</th>
<th>Symbol</th>
<th>Unit</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>Snow cover area</td>
<td>SCA</td>
<td>% or km²</td>
<td>Percentage (%) or area (km²) of grid cell (1 km²) covered by snow</td>
</tr>
<tr>
<td>Snow water equivalent</td>
<td>SWE</td>
<td>mm</td>
<td>Amount of water contained in the snowpack</td>
</tr>
<tr>
<td>Snowfall</td>
<td>Snowfall</td>
<td>mm/month</td>
<td>Monthly precipitation falling as snow</td>
</tr>
<tr>
<td>Snow season duration</td>
<td>D</td>
<td>days</td>
<td>Length in days from the start to end of snow season</td>
</tr>
<tr>
<td>Regional index</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Snow cover extent</td>
<td>SCE</td>
<td>km²</td>
<td>Total surface covered by snow derived from the grid cell snow covered area (SCA)</td>
</tr>
<tr>
<td>Regional SWE</td>
<td>regSWE</td>
<td>mm/month</td>
<td>Spatial average of SWE within each region</td>
</tr>
</tbody>
</table>
• differences in mean monthly SCE and SWE (grid cell scale) between the last 30-year period (1981–2010) and the first 30-year (1961–1990) reference period, and in the duration of the snow season for each region and macro region (Figures 3 and 4);

• differences in mean monthly SCE, regSWE, temperature and precipitation between the last 30-year period (1981–2010) and the first 30-year (1961–1990) reference period in each macro region (Figure 5);

• correlations between time series of snow indices and climate variables (snowfall, rainfall and temperature) for the complete 50-year period (Figure 6);

• trend magnitude for running mean 30-year periods of temperature (Figure 7);

• variability in monthly temperature trends across Norway (Figure 8).

The trends are calculated using the Theil–Sen estimator (Theil 1950; Sen 1968), which is a nonparametric method for trend detection that is widely applied to hydro-meteorological time series (e.g., Martinez et al. 2012). The trend magnitude is defined as the median of the slopes connecting all possible pairs of values of the time series:

$$TS = \text{median} \left( \frac{y_j - y_i}{x_j - x_i} \right)$$

where $x_i$ is the time of observation $i$, $y_i$ is the value of observation $i$, and $(y_j - y_i)/(x_j - x_i)$ are slopes calculated for all pairs of values.

### Table 2 | Snow cover codes following Dyrrdal & Vikhamar-Schuler (2009)

<table>
<thead>
<tr>
<th>Snow cover code</th>
<th>Description</th>
<th>Percentage used in this paper</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>No snow</td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>Minor parts of the ground covered with snow</td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>Equal areas with and without snow</td>
<td>SCA = 50%</td>
</tr>
<tr>
<td>3</td>
<td>Major parts of the ground covered with snow</td>
<td></td>
</tr>
<tr>
<td>4</td>
<td>Ground fully covered</td>
<td>SCA &gt; 95%</td>
</tr>
</tbody>
</table>

Figure 2 | Seasonal decadal running mean for SCE (a), regSWE (b), temperature (c) and precipitation (d) for all Norway. Black thin continuous lines show linear trends for each season for ease of comparison.
Figure 3 | Difference in mean monthly SCE (km²) between 1981–2010 and 1961–1990 (first and third row) and mean monthly SCE (km²) for the reference period 1961–1990 (second and fourth row). Please refer to the online version of this paper to see this figure in colour: http://dx.doi.org/10.2166/nh.2017.051.
Figure 4 | Differences in mean monthly SWE between 1981–2010 and 1961–1990 (first row) and mean monthly SWE in the reference period 1961–1990 (second row). Please refer to the online version of this paper to see this figure in colour: http://dx.doi.org/10.2166/nh.2017.051.

Figure 5 | Difference in mean monthly values between 1981–2010 and the reference period 1961–1990 in each macro region for: SCE (a), rSWE (b), temperature (c) and precipitation (d).
Among the advantages of the Theil–Sen estimator are its lower sensitivity to outliers and its higher accuracy, compared to less-robust simple linear regression (El-Shaarawi & Piegorsch 2002). Regional trends were derived by aggregating the grid cell time series across each of the 19 regions and four macro regions prior to deriving the Theil–Sen trend magnitude.

Correlation between climate (temperature, snowfall and rainfall) and snow variables (SCE and SWE) was estimated using the Pearson correlation coefficient (Pearson 1933) at the macro region scale. The significance of the correlation coefficient is calculated as follows (assuming normal distribution):

$$t = \frac{r}{\sqrt{(1 - r^2)/(N - 2)}}$$  \hfill (2)

where $t$ is the critical t-value (we chose a significance level of 0.05), $N$ is the number of observations used in the correlation (here: 50 years, from 1961 to 2010), and $r^2$ is the explained variance in a linear regression.
Figure 7 | Spatial distribution of trend in monthly temperature, shown as significance (Mann–Kendall test; first and third rows) and trend magnitude (Theil–Sen trend; second and fourth rows) for the period 1981–2010. Please refer to the online version of this paper to see this figure in colour: http://dx.doi.org/10.2166/nh.2017.051.
RESULTS

We begin by presenting seasonal and monthly changes and trends in local and regional snow indices (SCE and regSWE), including an analysis of their variability in time and space. Further, a comparative analysis of changes and trends in snow indices and climate variables (temperature and precipitation), including their mutual correlation, is presented.

Temporal patterns

Decadal (10-year) running means of snow cover extent (SCE in km^2) and regional average snow water equivalent (regSWE in mm) for all Norway are shown in Figure 2 along with corresponding values for temperature and precipitation. The whole period (1961–2010) is covered, starting with the decadal mean for 1961–1970 and then shifting the 10-year period forward 1 year at a time (i.e., 1961–1970, followed by 1962–1971, and so on, until 2001–2010). Each season is plotted separately, together with corresponding linear trends to assist interpretation.

The highest SCEs (Figure 2(a)) are found in winter, with approximately 90% of the country covered by snow, followed by spring. For context, the total Norwegian land area (323,781 km^2) is indicated in the figure by a horizontal line. An overall decrease is seen in SCE, particularly in spring. The macro region average SWE (regSWE; Figure 2(b)) displays a similar decrease in winter and spring. Here the highest values of regSWE are found in spring, despite a lower area coverage (ref. highest SCE in winter). Overall, we see a marked decline in regSWE in more recent decades. In summer and autumn, when there is little snow and glaciers make up much of the SWE area, SWE shows only minor changes.

To help understand these patterns, decadal mean temperature and precipitation for all Norway are also plotted in Figure 2. The decadal mean temperature (Figure 2(c)) increased during all seasons, with the largest increase and highest temporal variability seen in winter (~2–3°C/decade). The data show warmer winters in the early 1970s and early 1990s, and colder winters in the early 1980s. Decadal mean precipitation (Figure 2(d)) shows a much more diverse pattern, particularly in winter and spring. Winter precipitation increased notably from the early 1980s and during the mid 1990s, and a similar, but weaker signal is also found in spring. Summer, on the other hand, shows moderate changes (slightly increasing precipitation) as does autumn (slightly decreasing precipitation). From 1961–1970 to 2001–2010, the annual precipitation (over all Norway) increased by 5%.

To better understand the processes controlling the changes observed in SCA and regSWE for all Norway, we...
disaggregated the results in space (grid cell scale) as well as in time (monthly averages). The first and third rows of Figure 3 show the difference in mean monthly SCE (in km²) between 1981–2010 and the reference period 1961–1990, i.e., SCE_{1981–2010} minus SCE_{1961–1990}, by month. The mean SCE for the reference period (1961–1990) is shown in the second and fourth row, scaled from no snow to fully snow-covered grid cell. Overall, a decrease in SCE is seen for most months, with the largest reduction in winter and spring. In January and February, the reduction is notable along the coast in southern Norway, extending further inland (mainly in low-lying areas below approximately 300 m a.s.l.), and northwards in spring (April, May) and early winter (November, December). Thus, the largest changes in SCE occurred at the beginning and end of the snow season. Minor, but consistent, changes were seen in summer and early autumn, when the presence of snow is limited to mountain and northern regions. For instance, June shows a reduction in the northern part of Norway where snow is still present, whereas an increase is seen in mountain regions in the south. Increasing SCEs are mainly limited to mountain regions in southwestern Norway in summer (June, July). Minor changes occurred in July and August when snow is mainly present on glaciers.

Using a similar format as Figure 3 (for SCE), Figure 4 shows monthly maps of the differences in mean monthly SWE (in mm/month) between the most recent 30-year period (1981–2010) and the reference period (1961–1990) (first row), along with reference maps of the mean monthly SWE (in mm) over the reference period 1961–1990 (second row). Here, only the winter and spring months are shown because only minor changes are found in summer and autumn. The main features of the SCE maps (Figure 3) are also reflected in SWE (Figure 4), such as a decrease in SWE for winter, becoming more pronounced in spring, particularly along the coast and in the north. However, there are also clear differences to SCE, notably the distinct increase in SWE in mountainous regions in southern Norway, particularly from February to May. These notable increases occur in regions with a high SWE. The exceptions are region 10.2 and the southern part of region 11, which both have high SWE values, but also experience a large decrease in SWE of up to 150 mm between the two periods.

Regional analysis

Differences in mean values of snow indices and climate variables between 1981–2010 and the reference period 1961–1990 are given in Figure 5 for each month and macro region. Changes in the mean seasonal SCE for Norway separated by macro region (Figure 5(a)) show a decrease in SCE for all macro regions in all months (except in Mountain during June and July). This trend is especially strong at the start and end of the snow season. In April and May, SCE decreased by more than 20,000 km² over all Norway (around 8% of the total country area), while in November and December, the total reduction over all Norway was around 15,000 km². The North had the largest monthly decrease of all macro regions (ca. 15,000 km² in May). The South-East experienced a nearly constant decrease of ca. 5,000 km² in each month from November to May, except in April, when SCE reduced by ca. 10,000 km² in the South-East. In fact, calculating the relative change in percent, the South-West and South-East had the greatest relative SCE reduction (−4.5% and −5.2%, respectively), followed by North (−2.8%) and Mountain (−1.6%).

During winter and spring, regSWE decreased for all macro regions except for Mountain (Figure 5(b)), where a consistent increase in regSWE was seen from January to July, peaking in March/April. During late spring there was a consistent decrease in regSWE for South-East and North regions, whereas regSWE increased for the South-West in early summer, when snow covers a limited surface area.

Temperature increased for all macro regions and months except June, with the greatest changes occurring in late winter and early spring (Figure 5(c)). South-West registered the highest monthly temperature increase of more than 2.5°C (in February); however, the monthly variability in time overall follows a similar pattern across all macro regions.

Precipitation showed a notable increase in Mountain and South-West in January and February, with an increase of more than 120 mm/month compared to the reference (Figure 5(d)). These two regions overall follow a very similar seasonal pattern in monthly precipitation changes as do South-East and North although with a dampened amplitude. Here, a maximum precipitation increase is seen in
February (more than 50 mm/month) and small changes are observed during the rest of the year. The notable increase in precipitation in winter (January and February) coincides with increasing snow accumulation (SWE; Figure 5(b)) in Mountain, this being the snowfall season.

The reduction in SCE is clearly visible from September onwards, i.e., at the start of the snow season, peaking in November/December and then again towards the end of the snowmelt season in April/May.

To further analyse changes in SCE (Figure 5(a)) in April/May and November/December, we calculated changes in the start and end of the snow season along with mean duration. These variables were calculated based on SCE for each region and macro region over the first and last 30-year periods (1961–1990 and 1981–2010, respectively) and are summarized in Table 3. The snow season duration decreased in all regions except for region 7.1 and four out of the five regions included in Mountain. The largest decrease in snow season duration occurred in the South-West and North macro regions, where the snow season was, on average, 15 and 17 days shorter in 1981–2010 compared to the reference 1961–1990. Conversely, the snow season duration increased in four (of five) regions constituting the Mountain macro region, producing an average snow season increase of 10 days.

The shorter snow season in the other three macro regions was caused by a delay in the start of the snow season of approximately 1 week, and an earlier end to the snow season in two of the three macro regions (nearly 1 week earlier in the South-West and almost 2 weeks in North; Table 3). All, but two regions, displayed a later start to the snow season (both part of Mountain macro region), whereas the end of the snow season varied more among regions, with a majority showing earlier snow-free dates, but with both positive and negative shifts observed, even within the same macro region. The increase in snow cover duration in Mountain was mainly due to a delayed end of the season, reflecting the larger snow volumes, particularly in the two eastern regions (2.2 and 7.2) with more than 2 weeks.

### Concurrent changes in snow and climate variables

Monthly correlation coefficients between snow indices (i.e., SCA and SWE) and climate variables (i.e., temperature,
Temperature changes and trends

We analysed temperature trends to assess concurrent spatial and temporal patterns indicating potential feedbacks between snow cover and temperature over the period 1981–2010. The maps in Figure 7 show regions of the trend significance for temperature (first and third row) and the trend magnitude (second and fourth row). The maps reveal a general temperature increase across Norway with notable regional and seasonal patterns. The warming trend is strongest (trend magnitude) in the far north in December and January. In April and again from June to September, the warming trend affects most of Norway. These are also the regions and time of year depicting large-scale significant trends. More localized (significant) warming was found for smaller regions in northern Norway in December, January and May and in southern Norway in June (Figure 7). The strongest trend magnitudes, seen on the Finmarksvidda plateau (region 12) in December and January, were not significant in all grid cells because of large interannual variability in temperature in this region. Trends towards cooler conditions are found in December, February, May and October; however, these are not significant and thus not commented on further.

Temperature trends were subsequently calculated for the three 30-year periods (1961–1990, 1971–2000 and 1981–2010) to evaluate the rate of change over time, shown as boxplots of all grid cells for each month in Figure 8. The trend is usually stronger in winter and spring (when also the interannual variability is highest), with peak values in January and April for the most recent 30-year period (1981–2010). April, July, August and September also show the strongest trend magnitude in the most recent period), indicating an acceleration of the warming. This is not the case in February, March, May and October; however, all months still show a warming trend. The largest interannual variability in trend magnitude was found in the coldest months of the year, from November to February (Figure 8).

DISCUSSION

The performed analysis allowed us to detect a clear reduction in SCE for all macro regions and an uneven reduction in SWE (with an increase mainly in Mountain). Clear regional and seasonal differences are found – providing insight into what is driving these rapid and diverse spatial and temporal patterns of change. The highest temperature increases are occurring in periods when we also find the largest reduction in SCE and SWE, at a time when the snow cover is still extensive, but melts rapidly (i.e., towards the end of melting period in April and May). The exception is Mountain, where the increase in SWE can be explained by notably higher precipitation during winter. The positive correlation between SWE and rainfall found for Mountain in December may be caused by refreezing rainfall or precipitation that actually falls as snowfall, but is reported as rain by the model. SCE is less affected by the increase in winter precipitation. The pronounced warming
in late summer and early autumn has a minor influence on SWE because the snow cover then is limited.

Comparison to other studies

Our findings are consistent with studies investigating the decline of the Arctic SCE (Tedesco et al. 2009; Callaghan et al. 2011; Derksen & Brown 2012). Early spring SCE in the Arctic has reduced by 11% in April during the period 1970–2010 (Brown & Robinson 2011), and by 18% in May to June in the period 2008–2012 compared to pre-1970 values (Derksen & Brown 2012), with larger decreases in maritime regions north of 60° N (Callaghan et al. 2011). The Arctic spring SCE decline observed over the past decades was found to be largely driven by increasing temperatures (Brown & Robinson 2011; Derksen & Brown 2012), enhanced by snow albedo feedbacks (Groisman et al. 1994; Dery & Brown 2007). Warming was, to a lesser extent, influenced by anomalous SST influencing the atmospheric circulation (Bao et al. 2011) or energy convergence (Mioduszewski et al. 2014).

A decline in snow cover duration is also observed in the Arctic, similar to what we observe for Norway, with up to 5 days/decade in autumn and up to 10 days/decade in the melt season for the periods 1979–2008 (Tedesco et al. 2009), 2008–2012 (Derksen & Brown 2012) and 1979–2007 (Callaghan et al. 2011). Previous studies of Norway have confirmed a shortening of the snow duration, particularly in southern Norway, with linear trends up to 25 days/decade (Dyrrdal & Vikhamar-Schuler 2009; Dyrrdal 2009, 2010). A significant decrease in the number of snow days was documented at almost half of the 585 meteorological stations studied by Dyrrdal (2009) across all Norway for the period 1960–2007, particularly after 1990 (which is when our first 30-year period ends).

We found the largest reduction in SCE during the snowmelt season. Dyrrdal & Vikhamar-Schuler (2009) similarly found the number of snow days in Norway to decline the most at the end of the melting period. These results are consistent with the explanations of Groisman et al. (1994) and Callaghan et al. (2011), who highlight that snow albedo feedbacks act more strongly in late spring and summer than at the onset of the snow season. Given the same reduction in SCE, more energy will be available at the surface in spring (end of the snow season) than autumn (start of the snow season), particularly at higher latitudes, when the incoming solar radiation is high. This is especially important north of the Arctic Circle where there is midnight sun in June, whereas the snowfall season usually starts after the autumn equinox. Accordingly, the snow albedo feedbacks are expected to be less important in autumn than in late spring and summer. Our results showed changes in SCE in both spring and autumn. The more mixed results for the snow duration are due to a combination of factors, including regionally varying precipitation trends, which is relevant for regions covering inland and coastal climates (e.g., regions 10.1, 10.2 and 11). Another factor is the high sensitivity to changes in temperature when the temperature is close to zero, which is relevant for regions with mild winters (e.g., regions 3, 4 and 5.2).

SWE changes as reported here for Norway are supported by previous studies, which similarly found increases in SWE at high elevations, particularly in mountain areas in southern Norway. Skaugen et al. (2012) found (significantly) increasing SWE for high elevation stations (elevation > 850 m a.s.l.) for the period 1961–1990. Dyrrdal et al. (2013) analysed trends in snow depth at 926 stations across Norway covering an elevation range of 1–1,700 m a.s.l. and found a significant decline in snow depth across lowland regions for the period 1961–2010. For 1981–2010, the decline was particularly visible at the western coast of Norway, but to a smaller degree in the eastern part – than was the case for the full period 1961–2010. Similarly to Dyrrdal et al. (2013), we find an accelerated decline in snow depth across all Norway for lowland regions during the past decades.

Accelerated warming in spring

The trend analyses presented here confirm a general warming in all seasons for Norway; most pronounced in South-West in winter, but with a similar seasonal (monthly) pattern within each macro region. Førland et al. (2016) similarly reported warming in all seasons for the period 1955–2014, with the largest temperature increase in winter and spring (12 stations).

As emphasized earlier, the pronounced warming in April coincides with the period of the strongest SCE changes. This
suggests that snow albedo feedbacks may play a role, as elaborated above. Further, with less snow on the ground, not only does the ground reflect less incoming radiation, but also, more energy is absorbed and available to heat the ground and lower atmosphere, leading to significant increases of the near surface temperature (Aas et al. 2017). The longer snow-free season allows more evapotranspiration and thus, potentially more intense drying of the soil in summer, which again may lead to soil moisture—temperature feedbacks (Seneviratne et al. 2010). Chapin et al. (2005) concluded that earlier snowmelt explained the increased summer warming in Alaska by snow and vegetation feedbacks during the last decades of the 20th century.

**Hydrological impacts**

Changes in SWE and snow cover duration significantly influence the flood regime, also affecting sectors like hydropower, water supply and tourism. The hydrological regimes in most inland and northern catchments in Norway are characterized by a snowmelt-generated spring flood, which continues into the summer months for glacier and high-alpine catchments (Gottschalk et al. 1973; Wilson et al. 2010; Vormoor et al. 2015). The observed decrease in snow volume during the winter has subsequently led to a shift in the timing and a decrease in snowmelt-generated floods (Wilson et al. 2010), which are observed earlier in spring (Dyrdal et al. 2012). This is in agreement with findings from other high-latitude regions, such as Finland (Irannezhad et al. 2015), Alaska (Chapin et al. 2005) and Canada/USA (Brown & Robinson 2011; Mioduszewski et al. 2014). This decreasing trend in snow volume is expected to continue into the future as warming trends continue; however, higher precipitation in high elevation regions may counteract this effect to some extent.

**Uncertainty**

The results of this study, which is based on interpolated climate data and modelled snow data, are affected by uncertainties in the data due to: (i) errors in the temperature and precipitation measurements, including reading errors, inhomogeneity, missing data and precipitation undercatch, and (ii) errors generated by the models/interpolation methods. Undercatch is particularly a large source of error when measuring snowfall and can reach 80% in areas exposed to high winds (Orskaug et al. 2011).

The spatial coverage of temperature and precipitation stations in Norway is denser at low elevations and populated areas, leading to under-representation of mountainous and remote regions. The quality of the seNorge data thus varies in space. seNorge performs well in gauged grid cells, but the performance in grid cells without observations is less reliable. While a grid resolution of 1 km produces an extremely fine grid for national-scale analysis, it does not perfectly reproduce the complex terrain in some extreme areas of Norway. The quality of seNorge also varies in time, because stations have been added and removed from the dataset, depending on data availability.

Trends are highly sensitive to the region and time period considered and care must therefore be taken when comparing across studies, as demonstrated also by our results. In particular, trend studies with annual or seasonally aggregated station values (e.g., Hanssen-Bauer et al. 2017) will have more smoothing of extreme values than our monthly data. For example, we found the largest temperature trend of almost 3°C/decade in December in Finnmark (1981–2010) as a grid cell value. In comparison, Hanssen-Bauer et al. (2017) found that, for all Norway, the linear temperature trend was highest for spring (0.13°C/decade for 1900–2014). Thus, the seasonal trend magnitudes reported in Hanssen-Bauer et al. (2017) are not directly comparable to our monthly trends, which implicitly will result in higher trends. Further, we report temperature trends at the grid cell scale, whereas Hanssen-Bauer et al. (2017) reported aggregated seasonal trends for six temperature regions based on station data. However, at individual stations, higher trends were reported (e.g., a linear trend of 0.63°C/decade for one station – Gardermoen).

Not only the trend magnitude, but also the trend significance depends on the time period under study. Natural climate variability adds to a long-term climatic trend, and in a short time series the influence of natural variability will be larger. For instance, we found the strongest trend magnitudes in December and January (for the period 1981–2010), similar to Nilsen et al. (2017). However, these trends were not significant because of the high temporal variability in temperatures in winter (for all sub-periods).
Atmospheric circulation is more variable in winter than in summer (Hurrell & Deser 2009), and winter temperatures are more closely related to the atmospheric circulation in winter than in summer (Vautard & Yiou 2009). The North Atlantic Oscillation (NAO), for example, has a strong influence on winter temperatures in northern Europe (Hurrell & Deser 2009) at an annual to decadal time scale. Between the 1960s and 1990s, NAO was in its positive phase, advecting warm, moist air from the Atlantic towards Norway, but returned to neutral or negative levels after the 1990s (Hurrell & Deser 2009). Our full period, from 1961 to 2010, and sub-periods therein, thus include the increase in NAO in whole or in part.

CONCLUSIONS

In this study, seasonal and temporal changes in snow indices and climate variables have been analysed to answer whether high latitude warming and the associated reduction in snow cover and volume, in turn, has led to accelerated warming in spring due to the snow albedo feedback. We have quantified changes in snow indices and climate variables as differences between two partly overlapping 30-year periods and as linear trends over the most recent of these 30-year periods.

First, we looked at the temporal development (decadal running means) for the most recent 50-year period, from 1961 to 2010, and detected a decrease in the SCE, especially at the start and end of the snow season. This implied a shortening of the snow season of 1–3 weeks. The SCE decreased by more than 20,000 km² for all Norway (8% of the country area) in April and May between the periods 1981–2010 and 1961–1990. A clear temperature increase is seen for most regions and seasons in Norway, with significant warming in spring, and late summer for most of the country.

Our findings suggest that warming, being more pronounced in spring (notably April), caused earlier snowmelt, whereas higher temperatures suppressed snow accumulation in autumn with the exception of high elevation (mountain) regions. Here, higher winter precipitation has led to more snow accumulating and a later end of the snow season. The fact that the largest changes occurred in spring when the days are longer and the solar radiation is stronger than in autumn, suggests that snow albedo feedbacks play a role.

This study has identified trends among climate and snow observations that are consistent with snow albedo feedback processes. Further study is needed using coupled land–atmosphere models to explore the coupling mechanism in greater detail.

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