

Future Changes in Northern Hemisphere Snowfall

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ABSTRACT

Using simulations performed with 18 coupled atmosphere–ocean global climate models from phase 5 of the Coupled Model Intercomparison Project (CMIP5), projections of the Northern Hemisphere snowfall under the representative concentration pathway (RCP4.5) scenario are analyzed for the period 2006–2100. These models perform well in simulating twentieth-century snowfall, although there is a positive bias in many regions. Annual snowfall is projected to decrease across much of the Northern Hemisphere during the twenty-first century, with increases projected at higher latitudes. On a seasonal basis, the transition zone between negative and positive snowfall trends corresponds approximately to the -10°C isotherm of the late twentieth-century mean surface air temperature, such that positive trends prevail in winter over large regions of Eurasia and North America. Redistributions of snowfall throughout the entire snow season are projected to occur—even in locations where there is little change in annual snowfall. Changes in the fraction of precipitation falling as snow contribute to decreases in snowfall across most Northern Hemisphere regions, while changes in total precipitation typically contribute to increases in snowfall. A signal-to-noise analysis reveals that the projected changes in snowfall, based on the RCP4.5 scenario, are likely to become apparent during the twenty-first century for most locations in the Northern Hemisphere. The snowfall signal emerges more slowly than the temperature signal, suggesting that changes in snowfall are not likely to be early indicators of regional climate change.

1. Introduction

Snowfall is an important quantity that has meteorological, climatic, and societal dimensions. This study examines how total snowfall and the ratio of snowfall to total precipitation are projected to change during the twenty-first century. From a climate perspective, changes in snowfall potentially translate into changes in snow cover. Such changes are important to the radiative processes on Earth because of the reflective properties of snow-covered surfaces. Snow cover increases longwave emissivity and has an insulating effect

on soil temperatures (e.g., Zhang et al. 2008). Snowfall also plays an important role in the hydrologic cycle through spring melting (e.g., Lettenmaier and Gan 1990; Groisman et al. 2001). In terms of human impact, snowfall has been linked to increases in automobile accidents (e.g., Eisenberg and Warner 2005) and heart attacks (e.g., Rogot and Padgett 1976). There are also large economic disruptions associated with snow events (e.g., Rooney 1967). Snowfall is also important for local tourism and winter recreation (e.g., Scott et al. 2008).

In the Working Group 1 Fourth Assessment Report of the Intergovernmental Panel on Climate Change (WG1 IPCC AR4), discussions of future changes in snowfall are incorporated into the discussions of snow cover changes, and there is an implicit assumption that changes in snowfall are similar. Northern Hemisphere snow

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cover since the 1920s has decreased little during the peak of the winter seasons. Decreases in snow cover extent and increased variability, however, are noted in the fall and spring transition seasons. Upward trends in higher-latitude temperatures are implicated in the observed trends in snow cover, as both the type of precipitation that falls and the rate that fallen snow melts are governed by temperature (Lemke et al. 2007).

Several studies examined how observed North American snowfall patterns changed over time. Scott and Kaiser (2004) examined snowfall trends from 1948 to 2001 and observed that snowfall increased over a narrow band from Colorado to Wisconsin to the lee of Lakes Ontario and Erie. Karl et al. (1993) found that area-averaged snowfall across northern Canada (55° – 80° N) increased at a statistically significant rate of $8.8 \text{ cm decade}^{-1}$ in the late twentieth century and that this increase resulted from an increase in precipitation and not a redistribution of precipitation between snowfall and rainfall. In southern Canada ($<55^{\circ}$ N), the trend was $-0.65 \text{ cm decade}^{-1}$, with most of the decrease occurring during the 1980s. Precipitation in southern Canada has significantly increased by $9.7\% \pm 2.1\%$ century $^{-1}$, implying that the snow fraction (i.e., the fraction of total precipitation falling as snow) has decreased. Karl et al. (1993) also found an increase in total precipitation in the United States since 1950, with decreases in both snowfall and the snow fraction, although these changes were not statistically significant. In their analysis of snowfall and precipitation across Canada and the United States, Groisman and Easterling (1994) found that most of the observed increases in North American annual precipitation during the twentieth century south of 55° N have occurred along the eastern sections of the continent and are consistent with a negative trend in the snow fraction in this region. Over Greenland, Thomas et al. (2006) examined laser altimeter measurements of surface elevations and concluded that the ice is thickening at elevations greater than 1500 m, which is interpreted by the authors as indicative of increasing snowfall.

When considering the potential for future changes in snowfall in a warming world, it is important to recognize that snowfall requires the coexistence of sufficiently cold temperatures and the occurrence of precipitation. Changes in snowfall will be governed, therefore, by the interplay between changes in temperature and precipitation. In locations in which temperatures are often marginal for snowfall, warming may tend to reduce the amount of snowfall by reducing the fraction of the total precipitation that falls as snow. In places where air temperatures are cold enough that increases in temperature will not greatly impact the partitioning of

precipitation into snow and rain, the increased moisture content of the warmer climate (e.g., Held and Soden 2006) may lead to increased snowfall. For some locations and seasons, these two tendencies will oppose one another leading to an uncertain outcome.

Compared with other climate variables, such as surface air temperature and total precipitation, less attention has been given to analyzing climate model projections of snowfall. Kunkel et al. (2002) analyzed results from two atmosphere–ocean general circulation models (AOGCMs) in an attempt to better understand how the frequency of lake-effect snowfall events might change in a warmer climate. While there might be a period in the early part of the twenty-first century where the length of time that the Great Lakes remain unfrozen enhances snowfall, there is an indication that the warming of the surface air temperatures will likely reduce the frequency of heavy lake-effect snowfall events by the late twenty-first century. In IPCC AR4, Christensen et al. (2007) mention that projected increases in precipitation may lead to increased snow accumulations in some locations across North America, particularly near the Arctic Ocean and in some parts of the Rocky Mountains.

This study examines future changes in snowfall over the Northern Hemisphere in an ensemble of climate model simulations coordinated under phase 5 of the Coupled Model Intercomparison Project (CMIP5). Section 2 of the study includes a discussion of the CMIP5 ensemble to be examined. Section 3 consists of an evaluation of the simulated snowfall climatology from the ensemble, which is performed as a test of model credibility. Section 4 discusses simulated snowfall trends during the twenty-first century and the mechanisms responsible for their seasonal and spatial variability. Section 5 examines the emergence of these trends from the background of unforced climate variability. Section 6 summarizes the overall findings of the study and discusses their implications.

2. The CMIP5 ensemble

The current analysis utilizes results from two standardized experimental designs from the World Climate Research Programme (WCRP)'s CMIP5 (Taylor et al. 2012), and the model data were downloaded using the Program for Climate Model Diagnosis and Intercomparison (PCMDI) web portal (http://www.pcmdi.llnl.gov/ipcc/about_ipcc.php). The first is a simulation of twentieth-century climate (hereafter designated 20C), with time-varying anthropogenic and natural forcing agents specified as modeling groups deemed appropriate, ending in 2005. The second is a twenty-first-century

TABLE 1. List of CMIP5 models used in this study, with full expansions.

Model name	Expansion	No. of ensemble members	Resolution (°lat × °lon)
BCC_CSM1.1	Beijing Climate Center, Climate System Model, version 1.1	1	2.8 × 2.8
CanESM2	Second Generation Canadian Earth System Model	5	2.8 × 2.8
CCSM4	Community Climate System Model, version 4	6	0.9 × 1.25
CSIRO Mk3.6.0	Commonwealth Scientific and Industrial Research Organisation Mark, version 3.6.0	10	1.875 × 1.875
GFDL CM3	Geophysical Fluid Dynamics Laboratory Climate Model, version 3	1	2.0 × 2.5
GFDL-ESM2G	Geophysical Fluid Dynamics Laboratory Earth System Model with Generalized Ocean Layer Dynamics (GOLD) component (ESM2G)	1	2.0 × 2.5
GFDL-ESM2M	Geophysical Fluid Dynamics Laboratory Earth System Model with Modular Ocean Model 4 (MOM4) component (ESM2M)	1	2.0 × 2.5
GISS-E2-R	Goddard Institute for Space Studies Model E, coupled with the Russell ocean model	16	2.0 × 2.5
HadGEM2-ES	Hadley Centre Global Environment Model, version 2 - Earth System	4	1.24 × 1.875
INM-CM4.0	Institute of Numerical Mathematics Coupled Model, version 4.0	1	1.5 × 2.0
MIROC5	Model for Interdisciplinary Research on Climate, version 5	3	1.4 × 1.4
MIROC-ESM-CHEM	Model for Interdisciplinary Research on Climate, Earth System Model, Chemistry Coupled	1	2.8 × 2.8
MIROC-ESM	Model for Interdisciplinary Research on Climate, Earth System Model	1	2.8 × 2.8
MPI-ESM-LR	Max Planck Institute Earth System Model, low resolution	3	1.875 × 1.875
MPI-ESM-MR	Max Planck Institute Earth System Model, medium resolution	3	1.875 × 1.875
MRI-CGCM3	Meteorological Research Institute Coupled Atmosphere–Ocean General Circulation Model, version 3	1	1.125 × 1.125
NorESM1-ME	Norwegian Earth System Model, version 1 (intermediate resolution, interactive carbon cycle)	1	1.875 × 2.5
NorESM1-M	Norwegian Earth System Model, version 1 (intermediate resolution)	1	1.875 × 2.5

climate simulation using the representative concentration pathway (RCP4.5) forcing scenario (Moss et al. 2010; Thomson et al. 2011) extending to the year 2100. The atmospheric CO₂ concentration increases to approximately 525 ppm by 2100 in this scenario. The RCP4.5 scenario was chosen because it was run by a large number of modeling centers, and the results from these simulations were widely available at the start of this analysis.

Eighteen state-of-the-art coupled climate models were incorporated in this analysis (Table 1). A total of 60 RCP4.5 ensemble members were available for download from the PCMDI portal as of 1 September 2012. In some of the analyses performed in this study, each individual ensemble member is treated as an independent and plausible realization of the future climate. In cases where multimodel ensemble means are analyzed, averaging was performed across all ensemble members of a given model first so that the 18 models were weighted equally.

There is some variety in the manner in which models determine precipitation type. Some models use a diagnostic scheme based on temperature. If the temperature in the model layers closest to the surface is below freezing, any precipitation that falls is characterized as snow. Improvements in bulk microphysical schemes in climate models have resulted in changes in the way

precipitation type is determined, such that some of the models used in this study are capable of representing mixed-phase (ice, graupel, and water) clouds in their microphysical schemes. In these schemes, precipitation type is determined at the time of condensation in the cloud. Snow, ice, and rain are represented as individual mass fluxes in a vertical gridcell column. If falling snow and ice encounter a layer of above-freezing temperatures, then the mass is converted from frozen to liquid precipitation.

3. Evaluation of simulated snowfall climatology

The evaluation of the models' ability to simulate snowfall is complicated by the lack of gridded snowfall observations that are hemispheric or global in nature. In the literature, it is not uncommon to determine snowfall from the climatology of temperature and precipitation. For example, Roesch (2006) compared model-simulated snowfall with precipitation data from the Global Precipitation Climatology Centre, where the University of East Anglia Climatic Research Unit (CRU) temperature data were used to indicate snow if the temperature was below freezing. Here, we developed a derived snowfall product based on the University of Delaware Global Surface Air Temperature and Precipitation Climatology, version 2.01 (Willmott and Matsuura 2009a,b, hereafter WM).

Version 2.01 of the WM climatology is based on a blend of multiple temperature and precipitation station data sources, including the Global Historical Climate Network (version 2, Peterson and Vose 1997) and the station archive documented in Legates and Willmott (1990). Full documentation of the data sources and methods is available online (Willmott and Matsuura 2009a,b). The number of temperature stations used in a given month varied between 1600 and 12 200 globally while the number of precipitation stations varied between 4100 and 22 000 globally. WM interpolated the station data to a $0.5^\circ \times 0.5^\circ$ grid using a spherical distance-weighting method. They also accounted for the effects of elevation by using a digital elevation model (DEM) and adjusting the temperature using a fixed environmental lapse rate of 6°C km^{-1} . No attempt was made to adjust the precipitation for elevation. The data were obtained from the Climate Data Archive of the University of Washington's Joint Institute for the Study of the Atmosphere and Ocean (<http://jisao.washington.edu/data/>).

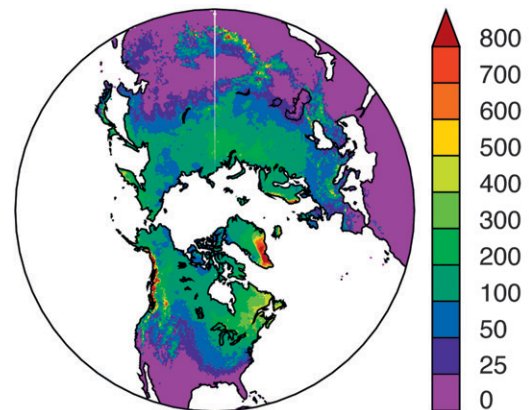
An empirical relationship between the fraction of precipitation falling as snow and monthly temperature (e.g., Cehak-Trock 1958; Legates 1987; Rawlins et al. 2006) shown in Eq. (1) was applied to the WM temperature and precipitation data to produce large-scale gridded estimates of snowfall for the Northern Hemisphere for the period from 1951 to 2005:

$$f = \frac{1}{1.0 + 1.61 \times (1.35)^T}. \quad (1)$$

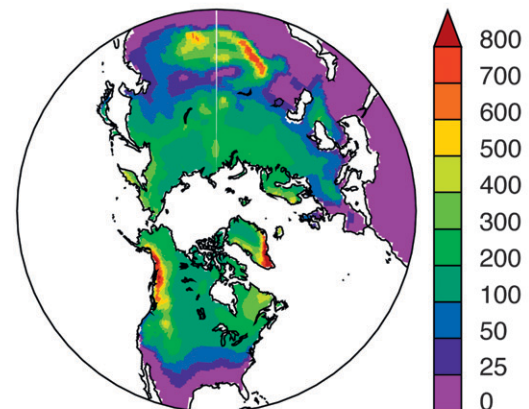
In Eq. (1), f represents the monthly mean fraction of total precipitation that falls as snow as determined by the monthly mean 2-m temperature T . For both the derived historical snowfall data and the model output, a constant 10:1 ratio was used to convert liquid precipitation to a snowfall depth equivalent for easier comparison to snowfall observations. There is considerable spatial and temporal variability of this ratio, but it is poorly constrained by observations. Changes in this ratio are not included in this study.

The 55-yr climatology of the WM-derived snowfall is shown in Fig. 1a. The WM-derived climatology exhibits a meridional gradient of annual snowfall along the eastern coast of North America, with a maximum of approximately 400 cm yr^{-1} in eastern Québec. Local maxima of snowfall are also found across the western United States and Canada. Other significant features of the Northern Hemisphere climatology of snowfall are also captured by the WM-derived data, as evident from comparisons with data published in the *World Survey of Climatology*. Across Europe and Asia, snowfall is well represented. Local maxima exist in the mountainous

A. WM-Derived



B. CMIP5 Ensemble



C. Difference (B minus A)

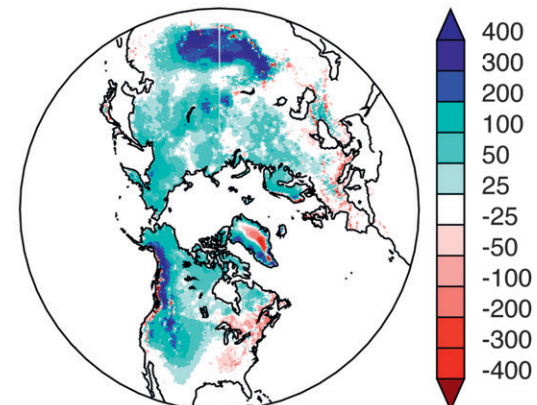


FIG. 1. (a) The 1951–2005 average annual snowfall (cm yr^{-1}) from the WM-derived data, (b) the CMIP5 multimodel ensemble, and (c) the difference.

regions of the Pyrenees, Central Alps, and Scandinavian Peninsula. The WM-derived data also exhibit a minimum of snowfall in northwestern China. This lack of snowfall is consistent with the findings of Watts (1969), who finds that low amounts of snowfall are the result of low amounts of precipitation in this region.

Figure 1b depicts the climatology of annual snowfall across the Northern Hemisphere for the multimodel ensemble. Snowfall generally increases with latitude in the Northern Hemisphere and is heaviest across western North America, northern Québec, parts of Russia, and Japan. Snowfall is enhanced by orography and is heaviest in locations that feature cold temperatures and a nearby source of moisture. Snowfall totals are found to be relatively small across the southern United States and western and southern Europe. A pronounced relative minimum of snowfall appears in the desert regions of northwestern China.

A difference map between the simulated and observed snowfall (Fig. 1c) indicates a positive snowfall bias in the multimodel ensemble relative to the observed data over much of North America. The tendency for phase 3 of the Coupled Model Intercomparison Project (CMIP3) models to exhibit a cold bias (John and Soden 2007) still persists in the CMIP5 models, particularly in the middle troposphere (Brands et al. 2013), and this general cold bias contributes to the positive snowfall bias seen in this analysis. There are indications that this bias might be reduced in the CMIP5 models—such as over China (Xu and Xu 2012). More analysis is needed to determine whether this is the result of improved model physics, from different historical representations of climate forcing agents (i.e., aerosols, well-mixed GHGs, solar, and volcanic), or both. The patterns of relative maxima and minima of snowfall, however, are captured reasonably well by the models. The magnitude of annual snowfall is in better agreement over the eastern half of North America, while the largest absolute errors are found in the western part of the continent. Note that lake-effect snow features are on a scale that is too small to be simulated by the models or captured in the WM-derived data. Over Eurasia, the snowfall maximum in the central plateau of Siberia is significantly oversimulated by the models by as much as 250–300 cm. In all of the central Eurasian mountain ranges (e.g., Alatai, Tien Shan, Sayan, and Aldan), snowfall is overestimated in the multimodel ensemble. Snowfall is also undersimulated in the models across central Europe, while it is overestimated across the Himalaya Mountains and Southeast Asia.

The ability of the models to adequately represent continental topography and the distribution of climate-observing stations in regions of rugged topography are important factors that must be considered when assessing model performance. These issues can be illustrated

by considering the snowfall simulation over North America. Better agreement between the multimodel ensemble and the observed data in the eastern half of North America coincides with the relatively flat terrain and a higher density of stations to capture the regional variations of snowfall. The larger errors in western North America are consistent with the coarse representation of the Rocky Mountains in the models as well as with biases that may be present in the observational data.

Snowfall is underestimated in observational data in regions of rugged terrain because climate-observing stations are biased toward lower elevations. Such valley locations are likely to be warmer than the surrounding higher elevations and thus experience a smaller fraction of precipitation falling as snow. The WM dataset attempts to correct for this bias by applying an elevation adjustment to temperature. There is no adjustment, however, for elevation effects on the precipitation amount, which would also lead to an underestimation of snowfall.

Lack of spatial resolution in the model topography also has a serious impact on the simulation of snowfall. The horizontal resolution of the climate models is on the order of 100 km, making the models incapable of resolving individual peaks and valleys. A comparison of the mean topography of the models with higher-resolution topographic data indicates that the high-elevation regions in the models are more broad and flat than in actuality, and the individual mountain ranges within larger regions of mountainous terrain are not well represented.

Taken in combination, potential biases in observed snowfall data and inadequate resolution of topography by global climate models greatly complicate the interpretation of model data comparisons in regions of complex terrain. The existence of these issues also requires that results regarding future snowfall trends from such regions should be regarded with considerable caution. It is encouraging to note, however, that improvements in the simulation of snowfall are likely to result from increases in model resolution (Kapnick and Delworth 2013).

4. Simulated future trends in snowfall

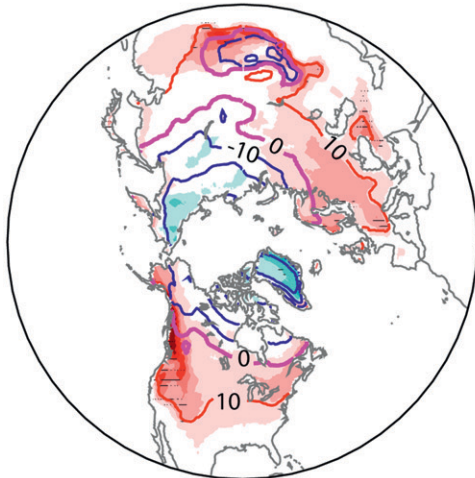
a. Trends in annual- and seasonal-mean snowfall

The changes in annual and seasonal snowfall between the periods 2006–25 and 2081–2100 are shown for the multimodel ensemble (Fig. 2). The change in snowfall over a given region, or ΔS , is defined as

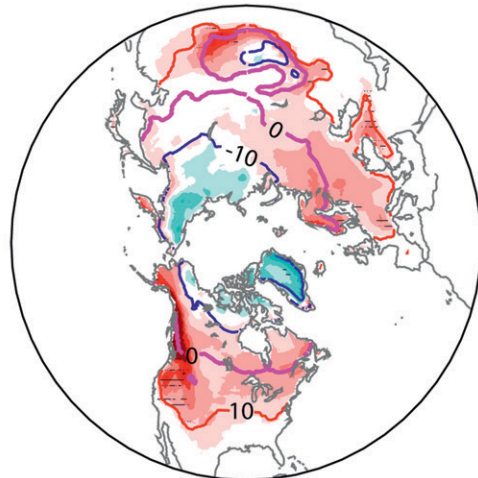
$$\Delta S = S_1 - S_0, \quad (2)$$

where S_0 and S_1 are the 20-yr average snowfall values from 2006 to 2025 and 2081 to 2100, respectively. The difference in the two 20-yr averages was divided by the

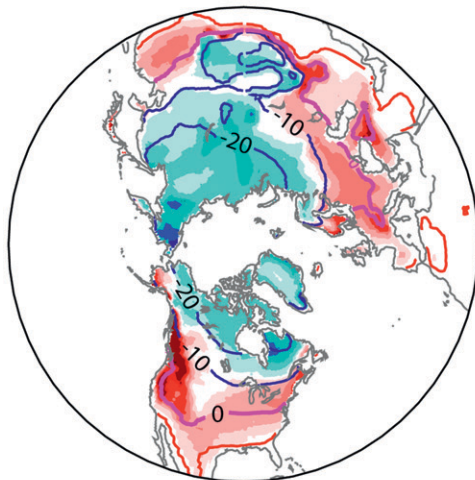
A.) Annual



B.) SON



C.) DJF



D.) MAM

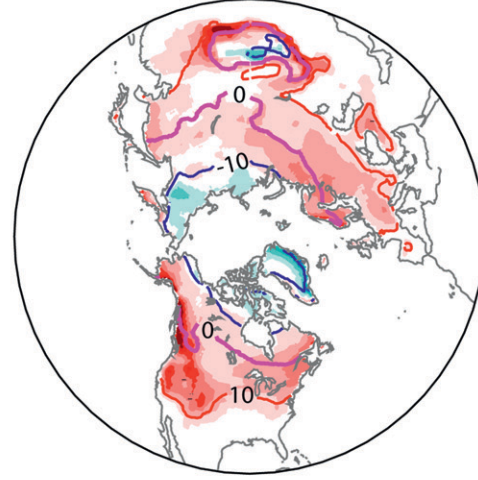


FIG. 2. Multimodel ensemble trends in snowfall (2006–2100; $\text{cm yr}^{-1} \text{ decade}^{-1}$): (a) annual, (b) September–November (SON), (c) December–February (DJF), and (d) March–May (MAM). Contours of the 2-m temperature at intervals of 10°C are shown from the multimodel ensemble for the period 1986–2005. Hatching denotes regions of statistically significant trends ($p \leq 0.01$).

time interval between the midpoints (75 yr) to give an average time rate of change over the twenty-first century. The 60 individual simulations were regridded to a common $0.5^\circ \times 0.5^\circ$ grid for analysis, and the data were masked if fewer than 45 simulations had valid data over land. The hatching in the figure denotes locations where

the two 20-yr periods differ significantly ($p \leq 0.01$) as determined by an independent t test where the degrees of freedom were adjusted for temporal autocorrelation (Santer et al. 2000).

The analysis of the multimodel ensemble indicates that annual snowfall is projected to decrease in the

twenty-first century across much of North America and Europe. Snowfall is projected to increase, however, at higher latitudes and across the interior parts of Greenland and Asia. The transition zone between positive and negative trends in annual snowfall is present for two reasons. First, precipitation type (i.e., the fraction of precipitation falling as snow) is insensitive to temperature changes at higher latitudes. In these locations, the air is sufficiently cold so that even modest warming is unable to raise the mean temperature above freezing. Second, there is consensus among the climate model projections that precipitation is likely to increase during the twenty-first century (e.g., Emori and Brown 2005). Both of these factors combine to produce an overall increase in snowfall in high latitudes.

The results of this analysis correspond well to regional studies that found increases in snowfall in locations where the snowfall-to-precipitation ratio was relatively constant (e.g., Knowles et al. 2006). The results also agree well with Hayhoe et al. (2006), who identify reductions in snowfall across the northeastern United States during the twenty-first century. The results also support the work of Thomas et al. (2006) and Box et al. (2006), who demonstrate that increasing snowfall is a necessary component in increasing the mass of the Greenland Ice Sheet.

While ice sheet growth and snow cover on land are distinctly different from snowfall, similar transition zones between positive and negative trends in the snow water equivalent (SWE) have been identified (e.g., Roesch 2006; Räisänen 2008). A key result from Räisänen (2008) is that a transition zone between positive and negative changes in SWE in midwinter broadly coincides with the November through March average -20°C isotherm of the late twentieth century (1986–2005). The snowfall trends from the CMIP5 multimodel ensemble, however, indicate that the transition between positive and negative trends in snowfall generally occurs near the average -10°C isotherm over the period 1986–2005 in the annual mean and in all three seasons (Fig. 2). The difference between values obtained in this analysis of snowfall and those in the Räisänen (2008) analysis of SWE may be explained by the fact that much colder temperatures are necessary in order to counteract the melting processes of snow and allow persistent snow cover to increase. In order for a location to experience an increase in snowfall without an increase in SWE, temperatures need not be as cold as those values identified by Räisänen. Since the temperatures corresponding to the boundaries of the transition zone show little variation with season, the relationship between seasonal mean temperature and snowfall trends is likely to be robust. This result provides the motivation for section 4c of this paper, where the individual effects of

temperature and precipitation changes on the overall snowfall trends are discussed.

Although the relationship between snowfall trends and the temperature isotherms is constant from season to season, substantial changes in seasonal snowfall are projected to occur (Fig. 2). Autumn trends are predominantly negative across most of the Northern Hemisphere, with the exceptions of Greenland and extreme northern Russia. In the winter season, however, positive trends are found in more locations—especially at higher latitudes. The most notable trend increases are simulated across northern Québec, Greenland, and central Russia. In the spring, snowfall decreases across much of the Northern Hemisphere.

In most locations throughout the Northern Hemisphere, temperatures during the transition seasons are higher than the temperatures during the winter. It is much more likely, therefore, that increases in temperature will influence the fraction of precipitation falling as snow in the transition seasons to a greater degree than in the winter (e.g., Knowles et al. 2006). These results also indicate that total annual snowfall in some locations is not projected to change, but the seasonal distribution of snowfall will be different.

b. Trends in monthly snowfall

The analysis in the previous section provides a first-order examination of the future trends in snowfall. In this section, the CMIP5 model simulations are examined individually on a monthly time scale. The objective is to provide a measure of uncertainty of projected snowfall trends across climate models and to highlight significant intraseasonal variability in those trends. Examining trends on a gridpoint scale can yield spatially noisy results, so a regional averaging approach was used for the analysis in this section to improve the signal-to-noise ratio. The 20 regions (Fig. 3) were chosen based on locations identified in the previous analysis. For these regions, only the land points within the regional boundaries are considered in this study.

As in the previous section, the difference between the mean snowfall values in the two periods can be converted to a rate of change by dividing the number of years that separate the midpoints of the two averaging periods — 75 years in this case. This method of examining future changes in snowfall was chosen to facilitate the decomposition of the snowfall change into the parts that are attributable to changes in temperature and precipitation (section 4c). The statistical significance of the change in snowfall between the two 20-yr periods ($p \leq 0.01$) was assessed using an independent t test that reduced the degrees of freedom according to the temporal autocorrelation of the data (Santer et al. 2000).



FIG. 3. Regions used in this study.

The regions can be broadly grouped into four classes. For space considerations, figures for only a few examples of the classifications are shown. The classes include decreases throughout all months (Figs. 4a,b), decreases during the transition seasons but increases in midwinter (Figs. 4c,d), decreases in snowfall that have little seasonal structure and high intermodel variability (Fig. 4e), or increases in snowfall that have little seasonal structure and high intermodel variability (Fig. 4f).

In the midlatitude regions (regions 1–3, 8, 12, and 15–16), this analysis indicates that snowfall will likely decrease throughout the entire snowfall season. The strongest changes in midwinter snowfall are projected to occur along the western coast of North America (region 9) and throughout much of central and southern Europe (region 16). There is also asymmetry between the reductions in snowfall during the transition seasons as the fall decreases are greater than the spring decreases.

In some regions, however, increases in midwinter snowfall are projected to occur. These regions are located at higher latitudes or higher elevations, such as northern Québec (region 6), Siberia (region 13), and Tibet (region 14). In these locations, midwinter increases in snowfall are accompanied by decreases in transition season snowfall. In Québec and Siberia, however, annual trends are close to zero. These results provide an example of how midwinter increases in snowfall are likely to be offset by the decreases in snowfall that will occur during the transition seasons.

In the final two classes of regions, there is little seasonal structure to the monthly changes in snowfall and a large degree of intermodel variability. In these locations, confidence in model projections of future snowfall is somewhat reduced. One of the regions that exhibit a large degree of variability is northeastern Greenland (region 20; Fig. 4f). Although the multimodel mean change in snowfall is positive for every month for this region, the increases are very small in the transition seasons. It is plausible that increases in snowfall over Greenland throughout the entire snowfall season are a result of sufficiently cold temperatures, increased precipitation, and close proximity to the extratropical storm track. Differences in simulating these features, however, may be responsible for the large degree of spread among the models.

Aside from Greenland, however, there are no other regions where snowfall increases in every month of the snowfall season. In the remaining regions across the Northern Hemisphere, the effects of reduced snowfall from warmer temperatures are felt at some point during the snow season. The results lend support to previous studies that argue for a shortening of the overall snowfall season (e.g., Hayhoe et al. 2006). Furthermore, there is evidence of a concentration of snowfall into a shorter period of time in locations where annual trends are very weak or close to zero (particularly along the coast of eastern North America; regions 1–6).

c. Individual effects of temperature and precipitation changes

Both the robust relationships between snowfall trends and temperature (section 4a) and the seasonal structure of future changes in snowfall (section 4b) motivate a more detailed analysis of the competing effects of temperature and precipitation changes within the snowfall season. Broadly speaking, warmer temperatures will result in less snowfall. Given the absence of temperature changes, however, more precipitation would result in more snowfall. In reality, these relationships are more complex as temperature and precipitation are correlated. In this section, we examine these three effects.

Snowfall may be expressed as the product of precipitation and the snow fraction:

$$S = P \times f. \quad (3)$$

Similar to the change in snowfall expressed in Eq. (2) the changes in precipitation and snow fraction are given as

$$\Delta P = P_1 - P_0 \quad \text{and} \quad (4)$$

$$\Delta f = f_1 - f_0. \quad (5)$$

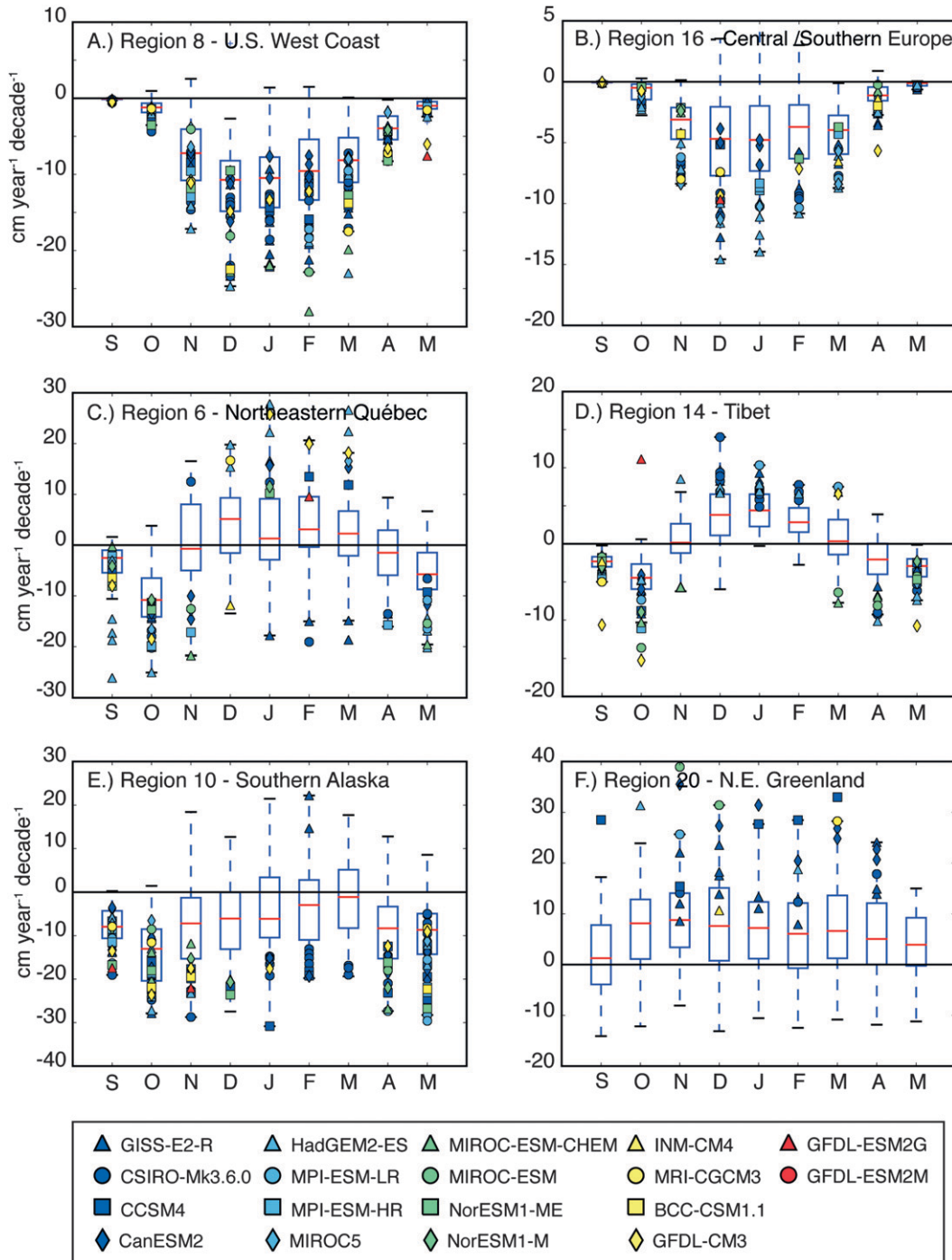


FIG. 4. Trends in monthly snowfall (2006–2100; $\text{cm yr}^{-1} \text{decade}^{-1}$). Box plots show the 25th, 50th, and 75th percentiles for all 60 ensemble members. Filled markers represent statistically significant trends ($p \leq 0.01$). Symbols were suppressed for model trends that are not statistically significant.

Through some algebraic manipulation, Eq. (2) may be rewritten as

$$\Delta S = S_1 - S_0 = P_0 \Delta f + f_0 \Delta P + \Delta P \Delta f, \quad (6)$$

where S_1 , P_1 , and f_1 represent the monthly snowfall, precipitation, and fraction of the precipitation falling as snow averaged over the period 2081–2100; and S_0 , P_0 , and f_0 are the same quantities averaged over the period

2006–25. If it is assumed that the snow fraction depends primarily on temperature, the three components on the right-hand side of Eq. (6) can be regarded as the contributions of temperature changes, precipitation changes, and their interaction to the overall change in monthly snowfall.

The analysis confirms that the projected changes in snow fraction have a contribution toward less snowfall during the twenty-first century in almost all regions of this study. The higher temperatures associated with greenhouse gas warming reduce the probability that precipitation will fall in the form of snow. Over northeastern Québec (Fig. 5a), higher temperatures and a change in the snow fraction dominate the transition season decreases in snowfall. This effect is most common in regions during the transition seasons and in some of the more southern regions considered in this study.

Across southern Alaska (Fig. 5b), this interplay is present throughout the entire snowfall season. The opposing effects of temperature and precipitation in this region are large and comparable to each other. Additionally, the interaction term between temperature and precipitation ($\Delta P \Delta f$) is more pronounced. This term, however, is small and does not have a first-order effect in determining the overall sign of ΔS .

Increases in snowfall, particularly at high latitudes (Fig. 5c) and during the midwinter season (Fig. 5a), are exclusively related to both an increase in precipitation (ΔP) and an absence of an effect from changes in the snow fraction. Held and Soden (2006) found that while the general pattern of precipitation minus evaporation ($P - E$) changes little in a warmer climate, the magnitude of $P - E$ is enhanced. They explain this robust response among climate models as resulting from warming-induced increases in lower-tropospheric water vapor, which simply enhance the existing pattern of meridional water vapor transport. Since $P - E$ increases with latitude over the region of interest and evaporation changes do not vary as much in space as precipitation changes, twenty-first-century trends in winter precipitation are greater at higher latitudes. Thus, the effects of temperature changes on snowfall trends are more likely to be offset by the effects of precipitation changes at higher latitudes.

The changes in snowfall related to the changes in ΔP are less spatially uniform than Δf . The reason is that despite a large-scale amplification of $P - E$, the underlying patterns of precipitation changes are more complex. Across much of North America and Asia, projections of twenty-first-century climate show a poleward shift of the midlatitude jet stream (e.g., Yin 2005). The increased poleward tracks of the midlatitude storm system works to further increase high-latitude precipitation.

The relative importance of ΔP and Δf varies significantly from region to region. In western North America

and central Europe, decreases in monthly snowfall are primarily the result of warmer temperatures and are only marginally affected by increases in precipitation. In the mid-Atlantic regions (regions 3 and 4), the competing effects are more balanced. This result adds support to the argument presented in Hayhoe et al. (2006) that the probability of precipitation occurring on cold days in these regions may stabilize a decrease in the number of snow days by the end of the twenty-first century. In the higher-latitude regions, such as Greenland, any decreases in snowfall associated with warming are largely overwhelmed by increases in snowfall related to precipitation increases.

5. Signal emergence analysis

The time series of model-projected snowfall during the twenty-first century includes both unforced natural variability and an anthropogenic warming signal. To determine when the simulated changes in snowfall resulting from anthropogenic warming emerge above the background noise of unforced variability, a signal emergence analysis was performed. This analysis was generated using time series of annual snowfall for each of the 20 regions used in this study.

In practical terms, the point of emergence represents a time when the simulated changes in snowfall might become noticeable. The signal–noise analysis assumed the vantage point of a hypothetical observer who has available only the time series of snowfall up to a particular point. In this analysis, the historical simulations (1901–2005) were included for each of the models. Using pairwise t tests that account for autocorrelation (Santer et al. 2000), the analysis first determined if the final 21 years (2080–2100) of the RCP4.5 transient climate simulation were statistically significant from the start of its corresponding historical simulation (1901–21). The 21-yr window was shifted backward in time by 1 year (e.g., 2079–99) until there was no longer a statistical difference between the analysis period and the start of the historical simulations.

The emergence year was reported as the concluding year in the earliest 21-yr period in the transient climate simulation that is significantly different from the years 1901–21. Thus, the emergence year represents the point in the simulated snowfall time series in which the mean snowfall over the previous 21 years significantly differs from the first 21 years of the twentieth century and does not allow the difference to be reversed at a later date. If there is no statistical significance between the final 21 years of the transient climate simulation and the reference period, then no emergence year is given.

A key result of this analysis is that for all 20 regions none of the median emergence years for snowfall occurs

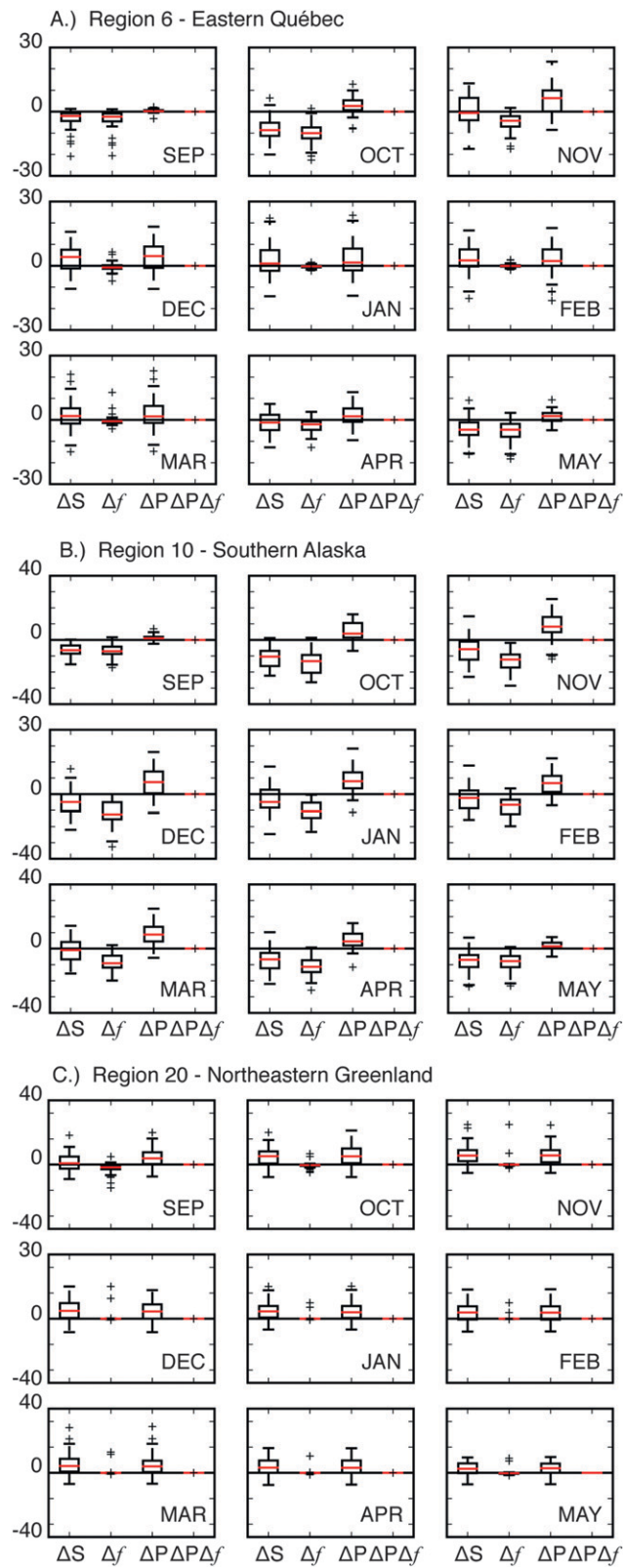


FIG. 5. Decomposition of the trends in monthly snowfall (ΔS) in the parts resulting from the change in snow fraction (Δf) and the change in precipitation (ΔP). The interplay of the two effects is shown as ($\Delta P \Delta f$). Trends are for the period 2006–2100 ($\text{cm yr}^{-1} \text{ decade}^{-1}$).

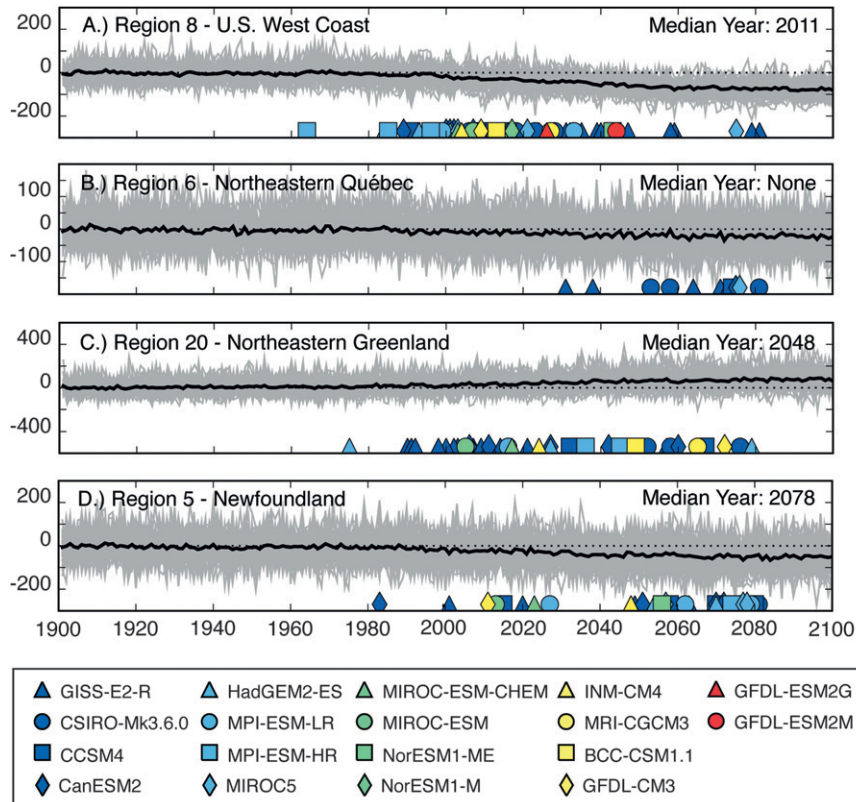


FIG. 6. Time series of regionally averaged annual snowfall from the 18 CMIP5 models (60 ensemble members total) used in this study (gray lines) and the multimodel ensemble (black line; cm yr^{-1}). The time series represent model projections based on the RCP4.5 climate change scenario. The symbols for each of the models denote years when the climate change signal emerges and remains above the background variability for the remainder of the experiment.

before 2002. There are, however, several regions where the median signal emergence year for the snowfall changes occurs in the early part of the twenty-first century. As an example, along the U.S. West Coast (region 8; Fig. 6a), the median emergence year is 2011. Other regions with early emergence years include western North America (regions 8–9), central Europe (regions 16–17), Japan (region 12), and the area near Turkey (region 15).

The values shown in Fig. 6 are median values, and in the example of Fig. 6a the projected emergence years for some ensemble members have already occurred. In these locations, the projected changes in snowfall might become apparent in the near future. As a result, noticeable changes in some of the factors affected by snowfall (e.g., changes in runoff and lower stream flows in spring) might also occur. The analysis shows that in some locations, strong trends in snowfall are likely to occur, and the effects of a changing climate should soon become apparent to snowfall observers.

In northeastern Canada, however, emergence years are much later in the twenty-first century. The absence of an emergence year is common in some regions at higher latitudes. Northeastern Québec (region 6; Fig. 6b), for example, does not have a signal emergence year in this analysis. The absence of emergent trends in these regions arises because of the large variability of snowfall and the offsetting effects of decreasing snowfall during the transition seasons and increasing snowfall during midwinter. The relatively late emergence years in southern regions and the absence of statistically significant trends in northern regions suggest that annual snowfall may not be a particularly sensitive indicator of climate change during the next few decades. Similarly, other Arctic subpolar and high-latitude regions show emergence years that occur during the middle and late in the twenty-first century. Examples include northeastern Greenland (region 20; Fig. 6c) and Newfoundland (region 5; Fig. 6d).

The signal emergence analysis was repeated for the 2-m temperature for comparison with snowfall (Fig. 7).

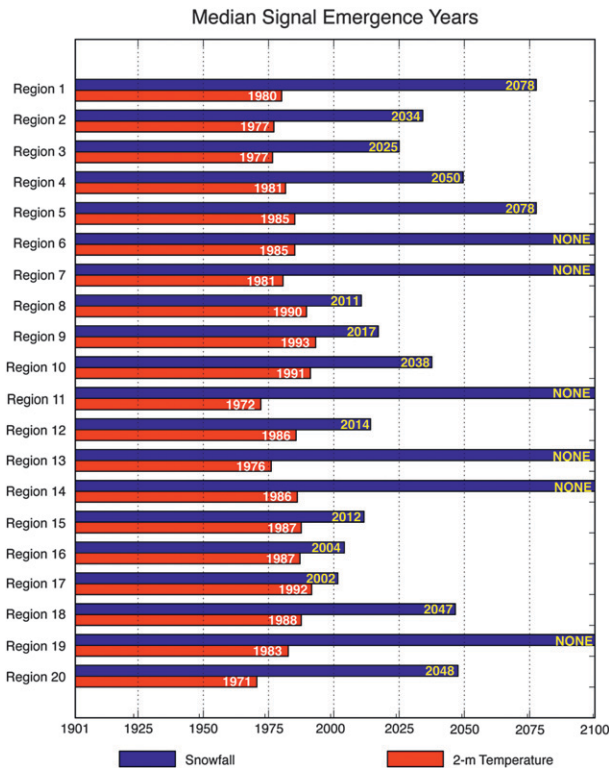


FIG. 7. Median of the signal emergence years ($n = 60$ CMIP5 simulations) for both temperature and precipitation based on the historical + RCP4.5 scenarios.

In general, the positive trends in temperature become apparent at an earlier point in time when compared to the trends in snowfall. For all of the regions examined, the median emergence year for the 2-m temperature occurs between 1971 and 1993—well in advance of the emergence of the snowfall signal. These results support Pierce and Cayan (2013) who also demonstrated that trends in temperature are detected earlier than trends in snowfall in the western United States.

The ability to detect the emergence of a signal above background noise depends on the variance, signal strength, and autocorrelation of the time series. Compared to temperature, the regional averages of snowfall have a weaker signal-to-noise ratio and less temporal autocorrelation. To demonstrate sensitivity to these parameters, the signal emergence methodology was applied to an idealized scenario. In the base scenario (Fig. 8a), 100 000 synthetic time series were generated from normal distribution where $\bar{x} = 0$, $\sigma^2 = 1$, $m = 1/100$, and $\rho_{t+1} = 0.4$, where m is the signal strength expressed as the slope of a linear trend and ρ_{t+1} is the lag + 1 autocorrelation. The sensitivity tests included a doubling of the signal strength (Fig. 8b), a doubling of variance (Fig. 8c), and a reduction of the lag + 1 autocorrelation by half (Fig. 8d). In the idealized cases,

increasing the signal strength and decreasing the autocorrelation both sharpen the distribution of emergence years around an earlier emergence year than in the base case. When the variance is doubled, however, the likelihood of the signal emergence test failing to detect an emergence year greatly increases.

The signal emergence years for snowfall occur later than temperature in every region. In the absence of any other differences, the reduced autocorrelation in the snowfall time series would result in earlier emergence years than the temperature time series. Therefore, the later emergence years are likely the result of a reduced signal-to-noise ratio in the snowfall time series. It is important to note that despite more variance in the time series of snowfall than in temperature, the signal emergence test is still able to identify an emergence year for snowfall for most regions and for most models. Although interannual variance is the primary control of snowfall signal emergence, the results based on the RCP4.5 scenario should not be interpreted as definitive as the signal-to-noise ratio would be different among the different RCP scenarios.

6. Discussion and conclusions

Simulations of monthly and annual snowfall were obtained for 18 AOGCMs (60 individual ensemble members in total) that participated in the WCRP CMIP5 project. The climate of the twentieth-century simulations from the models were compared against a snowfall climatology derived from the Willmott–Matsuura monthly temperature and precipitation climatology. The climatological mean snowfall amounts from the models are consistent with observations for most locations in the Northern Hemisphere, with a notable exception of locations with higher elevations. The bias toward increased snowfall at higher elevations may be the result of coarse terrain representation in the models while the observations of temperature and precipitation are biased toward lower elevations. In addition to the elevation concerns, the CMIP5 ensemble mean exhibits a cold bias that further contributes to a model overestimation of snowfall.

The analysis of the projected trends in annual and seasonal snowfall indicates the presence of a transition zone between positive trends at higher latitudes and negative trends in the midlatitudes. This transition zone is the result of high-latitude increases in precipitation in locations where warming during the late twentieth century does not alter the fraction of precipitation falling as snow. The gridpoint trends in annual and seasonal snowfall were compared to the late twentieth-century multimodel ensemble climatology of temperature. The transition zone between positive and negative trends of snowfall in the multimodel ensemble corresponds to

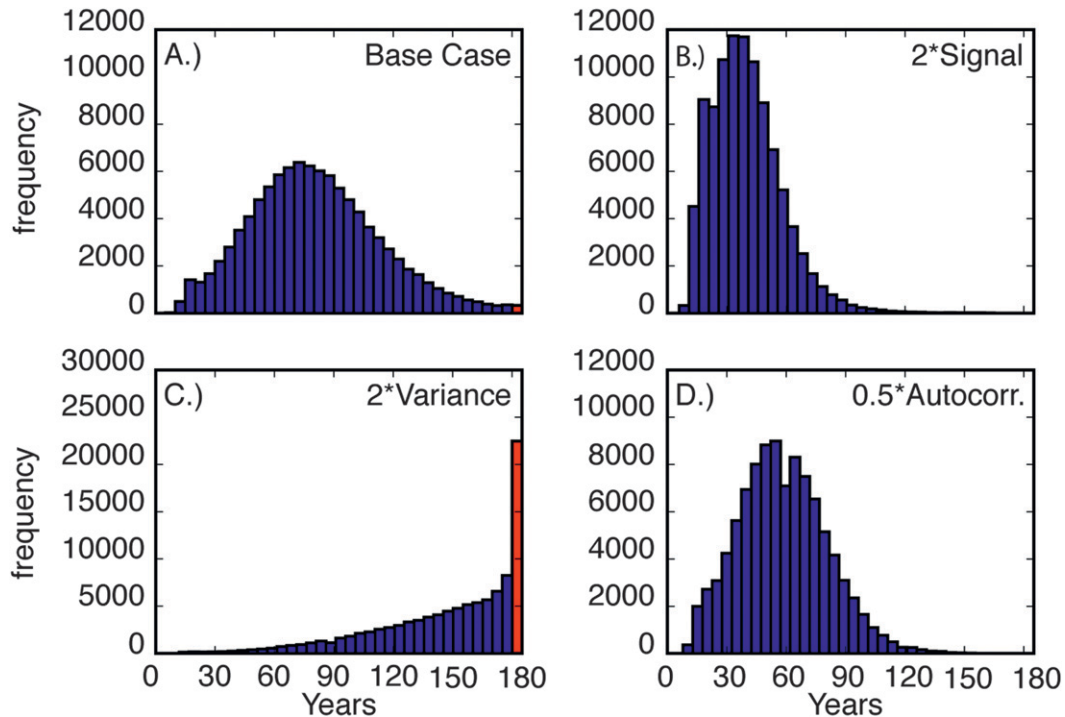


FIG. 8. Signal emergence years from an idealized scenario of 100 000 synthetic trends for (a) a base case, (b) a doubling of signal strength, (c) a doubling of variance, and (d) halving of the lag + 1 autocorrelation. Undetected signal emergence cases, if present, are shown in red along the right-hand side of each panel.

the -10°C isotherm from the period (1986–2005) in each of the fall, winter, and spring seasons.

Using a regional averaging technique, future trends in snowfall were examined on a monthly basis. The results of this analysis demonstrate that most regions experience decreases in snowfall during the fall and spring transition seasons. Some regions, however, show increases in midwinter snowfall. In some of these locations where midwinter increases in snowfall are found (i.e., region 6, northern Québec), gridpoint trends in annual snowfall are weak. This key result implies that decreases in snowfall that occur during the fall and spring seasons may be offset by increases in snowfall during the middle of the snowfall season. The projected trends in monthly snowfall were also decomposed into parts related to changes in the fraction of precipitation falling as snow and to changes in precipitation. The decomposition analysis indicates that changes in the snow fraction uniformly contribute to decreases in snowfall, as temperature is projected to increase in most locations around the world. Precipitation changes are generally positive, although their contributions are less uniform and greatest at higher latitudes. More impact analyses are needed in order to understand how the redistribution of snowfall throughout the snowfall season will influence other features of the climate system, such as runoff.

One of the major assumptions made in this study is the uniform 10:1 snow-to-liquid ratio in converting snowfall mass flux from the models to a snowfall depth. Research has shown that this ratio has substantial regional variations (e.g., Baxter 2005). As the temperature warms in the lower troposphere, the density of snowfall increases. In locations where temperature increases do not influence precipitation, the assumption of a 10:1 snowfall-to-precipitation ratio may overestimate the magnitude of the snowfall changes in these regions.

It is also possible that the snowfall-to-precipitation ratio varies as a function of the season. Since temperatures in the fall and spring seasons are warmer than in the midwinter, it is therefore likely that snowfall occurring in the middle of the snowfall season is less dense than snowfall that takes place in the transition seasons. It is also common for snow events to occur within the same season that have significantly different snowfall-to-precipitation ratios. Such variations are often the result of meteorological factors (i.e., temperatures in snow growth regions) and are difficult to simulate with climate models. Understanding of the regional variability of the snow-to-liquid ratio would further clarify the findings of this study.

The limitations of the CMIP5 AOGCMs must be considered when interpreting the results of this study.

Many of the AOGCMs used in this study diagnose precipitation type by using simple algorithms based on the temperature in the lower troposphere. Future work should investigate if explicit mixed-phase microphysical schemes that keep independent water and ice budgets significantly improve the simulation of snowfall in the models. Since snow is important in determining the vertical temperature structure of the atmosphere, evaluation of the feedbacks between snow and temperature in the climate models also requires further examination. The resolution of these models also fails to capture the mesoscale processes important to snowfall in certain regions and is partially responsible for the noted biases compared to observed data. Mesoscale modeling—either through increasing the resolution of these models or through dynamical downscaling—would improve the simulation of locally important features, such as lake-effect snowfall.

The results of this study are also important in the context of high-latitude climate change. The potential exists for an increase in snowfall to provide a positive contribution to the mass of the Greenland Ice Sheet. The increased ice and snow may partially offset the substantial melt that is expected to occur during the twenty-first century. These results support previous findings related to the nature of the ice balance over Greenland and underscore the necessity of including snowfall changes in future studies of the Greenland Ice Sheet.

The final analysis performed examined when future changes in snowfall might emerge from background climate variability. The results of this signal-to-noise analysis indicate that the forced changes in model-simulated annual snowfall are projected to rise above background climate variability at some point during the twenty-first century. In some locations, however, where the interannual variability is large or the trend signal is weak, or both, these conditions occur; no signal emergence year was found. The signal emergence years of the changes in snowfall were also found to be later than the emergence years of the changes in temperature.

The implications of this signal-to-noise analysis are important in broader discussions of anthropogenic climate change. Based on the sensitivity tests of the signal emergence method used in this study, large interannual variability hinders the ability to detect changes in snowfall. This suggests that changes in snowfall are not likely to be an early indicator of regional climate change. Despite signal emergence years that are projected to occur during the twenty-first century for most regions in the Northern Hemisphere, interannual variability will remain large. Therefore, it will still be possible for regions to experience high snowfall years in locations where there is a clear downward trend in annual snowfall.

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