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Global warming response of snowpack at mountain range in northern Japan estimated using multiple dynamically downscaled data

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Abstract

We estimate the response of snowpack to global warming along with the uncertainty of the snowpack change by using a combination of multiple general-circulation models (GCMs), a single regional atmospheric model, and a one-dimensional multi-layered snowpack model. The target site is Mt. Annupuri in Kutchan, Hokkaido, Japan. The forcing of the snowpack model is taken from dynamically downscaled data from GCMs for the present climate and GCMs in a decade when the global-mean temperature has increased by 2 K from present conditions. The results show that global warming would decrease the monthly-mean snow depth throughout the winter season. Other salient features are the decrease of snow depth by 60 cm with maximum uncertainty of 20 cm at the beginning of the snow ablation period, the occurrence of the snow-depth peak a month earlier, and the dominance of melt forms in an earlier season. The ratio of melt forms for all snowpack layers increase with little uncertainty before the snow ablation period. The ratio of hoar does not change much, even though the air temperature increases. The uncertainty in snowpack evaluation is also discussed.

Keywords

Snowpack modeling; Snowpack change; Snow type change; Uncertainty

1 Introduction

In the present climate, snowpack occurs seasonally in the extratropics and even permanently in mountainous regions, and it is thought to be strongly affected by global warming especially at mid-latitudes. According to the latest report of the Inter-governmental Panel for Climate Change (IPCC), the emission of greenhouse gases leads not only to an increase in air temperature but also changes in wind, precipitation amount and intensity, and cloudiness (Stocker et al., 2014). The temperature is one of the most important variables controlling the amount of snow, the properties of snowpack, and the snow cover period, because snow and ice are retained below freezing but rapidly melt above this temperature (López-Moreno et al., 2008, 2013). Moreover, wintertime snowfall, which represents the major contribution to total snow accumulation, is frequently observed in storm-track areas, such as northern Europe, northeastern America, and northern Japan. Because this snowfall is often caused by extratropical cyclone passage, it might be changed by global warming. The physical properties of snowpack might be also changed as a result of efficient transformation from solid precipitation particles to melt forms in a warmer climate. This transformation of grain types may also change the occurrence of wet-avalanche because that of wet-avalanche is related to the wetting of snowpack (Mitterer et al., 2011). Although the wet-avalanche sometimes occurs in Japan (Akitaya et al., 2015) and changes of the wet-avalanche occurrence have been pointed out in the North America (Lazar and Williams, 2008) and in France (Castebrunet et al., 2014), the transformation of grain types responding to the global warming in Japan is not well addressed. Moreover, considering the utilization of water resources (Beniston, 2003), the mitigation of snow disaster (Nakai et al., 2012), and winter tourism (Beniston, 2003; Uhlmann et al., 2009), the impact of climate change on snowpack dynamics deserve to be examined (e.g. Niwano et al., 2012; Mellander et al., 2007; Inatsu et al., 2016), even though there is still inherent uncertainty in models of the impact of climate change.

We found several studies of snowpack response to climatic change, such as an estimation of snowpack including its physical property along with its uncertainty by the use of one-dimensional and multi-layered snowpack model (e.g. Rasmus et al., 2004; Lazar et al., 2006; Rousselot et al., 2012). They consist mostly of three steps: (i) climate change projection with atmosphere–ocean general-circulation models (GCMs), in which atmospheric concentrations of greenhouse gases are prescribed as a function of year (Solomon et al., 2007; Stocker et al., 2014); (ii) downscaling, which creates climatic variables at a particular site or in a limited area with higher spatial resolution in order to compensate for the insufficient resolution of GCMs (Wang et al., 2004; Wilby et al., 2004); and (iii) snowpack estimates, either with a physical model (e.g. Rasmus et al., 2004) or with a statistical relation that has been empirically determined in advance (Inoue and Yokoyama, 1998). For step (i), there are no other technical choices than using GCMs. The GCM projection introduces an uncertainty in evaluating global-mean temperature, however, because of differing climate sensitivities among GCMs, mainly due to their physical parameterizations, and because future greenhouse gas emissions depend on the socio-economic scenario (Stocker et al., 2014). The GCM projections of wintertime temperature and precipitation at a particular mid-latitude site are also uncertain, due to uncertain changes in storm tracks and jet streams at mid-latitudes (Chang et al., 2012), wintertime Asian monsoon (Ogata et al., 2014), and the Arctic Oscillation (Karpechko, 2010). These regional climate patterns certainly affect snow accumulation and melt dynamics, and also snowpack dynamics. It should be noted that GCMs are more or less biased, so one needs a bias correction for a particular site. One then proceeds to step (ii) based on coarse-resolution GCM projection with uncertainty and bias. The methods of step (ii) can be classified into dynamical downscaling (DDS) (Wang et al., 2004) and statistical downscaling (Wilby et al., 2004). The former provides higher-resolution climatic variables in a limited area by integrating a regional atmospheric model (RAM) with the GCM output imposed as its lateral boundary condition. The latter estimates a future state by simply applying a statistical relation between local-site information and weather patterns such as the Siberian-Japan pattern that brings much snowfall along the Japan Sea side of northern Japan (Takano et al., 2008). The Siberian-Japan pattern is established based on the singular-value decomposition analysis between synoptic weather pattern and local precipitation, for example (Kuno and Inatsu, 2014). Recently, DDS has been widely used in the community, in spite of the need for additional computation, because it has the ability to produce a physically consistent dataset (e.g. Wang et al., 2004; Kuno and Inatsu, 2014; Inatsu et al., 2015). Multiple RAMs, even with a single GCM imposed as the lateral boundary condition, also provide uncertainty, mainly due to the variability among the RAMs' physical parameterizations, but the uncertainty is not large for the extratropics in winter because the DDS results are strongly controlled by lateral boundary conditions (Inatsu et al., 2015; Kuno and Inatsu, 2014). In step (ii) when using DDS, a bias correction should be made just before step (iii) because the DDS results have the systematic biases in atmospheric variables such as temperature, precipitation, and so on, due to physical parameterizations and resolution (Ishizaki et al., 2012). It should be noted that an alternative choice in step (ii) is the pseudo-global warming (PGW) experiment, in which observed weather time-series are added to the climatological difference estimated from GCM integrations so as to form the lateral boundary condition of the RAM (Kimura and Kitoh, 2007). Finally step (iii) estimates the future snowpack change, which is still a challenging problem. Although Inoue and Yokoyama (1998) estimated maximum snow depth and major snow type over Japan by using a statistical relation between snowpack and meteorological characteristics, recent studies have tended to use one-dimensional multi-layered snowpack models, such as CROCUS (Brun et al., 1992), SNTHERM (Jordan, 1991), and SNOWPACK (Bartelt and Lehning, 2002) and its modification for wet-heavy snow (Hirashima, 2014). A one-dimensional multi-layered snowpack model enables us to calculate the temporal evolution of snowpack structure with multiple layers at a particular site, driven by atmospheric variables, such as air temperature, precipitation, humidity, wind, and shortwave radiation at the snow surface. Step (iii) is, therefore, undertaken on the basis of bias-corrected atmospheric variables obtained from step (ii).

Several previous studies have been devoted to an evaluation of future snowpack change for particular areas, such as Switzerland (Bavay et al., 2009, 2013), Finland (Rasmus et al., 2004), France (Rousselot et al., 2012), and North America (Lazar et al., 2006). Basically, uncertainty in future snowpack was estimated under multiple emission scenarios of greenhouse gases. The emission scenario strongly controls the global-mean temperature increase, a factor to which snowpack estimation is sensitive. For example, Rousselot et al. (2012) revealed that the snow water equivalent change in the A2 scenario of the Special Report on Emissions Scenarios (SRES) was double to that in the B1 scenario. Bavay et al. (2009) also pointed out a great discrepancy between results for the A2 and B2 scenarios. Moreover, under the same emission scenario, different GCMs provide different global-mean temperature increases due to climate sensitivity. The use of multiple GCMs, therefore, increases the range of estimates of future snowpack (Bavay et al., 2013; Lazar et al., 2006; Rasmus et al., 2004).

When considering the effect of climate change on snowpack, one may desire to separate the changes due to differences in global-mean temperature from those due to changes of synoptic-scale climate around a particular site. However, since the temperature increase affects snowpack estimation quite strongly, it is difficult to determine the uncertainty arising from changes of meso-scale convection, storm tracks, and quasi-stationary pressure patterns by using a set of arbitrarily chosen GCMs.

From the point of view of numerical snowpack experiments, biased input data can cause problems. For example, a warm bias would shorten the snow season and a dry bias would effectively decrease the snow depth. Hence, we need a rational treatment for bias in GCMs and downscaled data. One way to ameliorate this problem is to offset climatological differences between present and future conditions assuming that the model biases are stationary. This approach has been used in several studies. Rasmus et al. (2004) calculated the snowpack difference for present and future climates using DDS without any bias correction of input data. Bavay et al. (2009; 2013) and Nakamura et al. (2011) estimated a change in snowpack using the PGW strategy. It is also expected that statistical downscaling will correct for any bias without requiring additional procedures (Rousselot et al., 2012; Lazar et al., 2006).

The purpose of this study is to estimate future snowpack evolution along with its uncertainty by a combination of multiple GCMs, a single RAM, and a one-dimensional snowpack model (Fig. 1). The analysis is based on the idea proposed in Inatsu et al. (2015), in which the synoptic-scale response was successfully separated from global temperature increase by performing DDS for a decade during which the global-mean surface air temperatures increase by 2 K. Here, we use the dataset archived by Kuno and Inatsu (2014) and skip steps (i) and (ii) of the procedure in Fig. 1. In pre-processing before the snowpack calculation, we make bias corrections for temperature and precipitation and height correction for temperature in order to discuss differences of the snowpack response with altitude. After pre-processing, the numerical snowpack calculation is performed for a particular mountain range at Mt. Annupuri in Kutchan, Hokkaido, Japan (Fig. 2).

This paper is organized as follows. Section 2 describes the study area including its climate. Section 3 briefly describes the observation, the downscaled data, and the model for numerical snowpack calculation together with the bias correction method. Section 4 shows the snowpack results for downscaled data under present and future climates. We also present the uncertainty of the estimates by using multiple GCMs. Section 5 discusses how the results can be interpreted. Finally, section 6 gives the conclusion.

2 Study area

We chose Mt. Annupuri as a particular mountain range for three reasons. First, the climate at the site is categorized as Dfb in the Köppen-Geiger climate classification characterized by cold, no dry season, and warm summer (Peel et al., 2007). The climatological air temperature is −4.7 °C and total precipitation attains 500 mm in December–February at the observation site of the Japan Meteorological Agency (JMA) in Kutchan. The snowfall is heavier than other areas around Kutchan because moisture-rich air produced above the sea is advected by winter monsoonal wind (e.g.

Takano et al., 2008). The climatological feature holds the snow-cover period exceeding 4 months and a maximum snow depth of 190 cm even at the mountain base. Second, this site encompasses the mountain top, at 1,308 m above sea level, down to a wide steep hill with its base around 200 m above sea level (Fig. 2b). The large difference of height at a single mountain enables us to facilitate the discussion on snowpack change with different temperature baselines. For the estimation, we considered three locations: the top (1,300 m), the hillside (800 m), and the base (173 m), which is the level of the JMA's meteorological station (Fig. 2b). The site is the downwind side of winter monsoon that brings heavy snowfall. The slope of Mt. Annupuri directs from southwest to northeast and the part of mountain area is leeward, but we do not consider the effect of such small-scale topography on the mountain slope. Finally, there is a social demand for the estimation because a famous ski resort with high-quality snow is located at the site.

3 Data and Methods

3.1 Data

3.1.1 Downscaled data

We used a dataset of DDS results provided by Kuno and Inatsu (2014). For this dataset, the 1990s in the 20th century experiment (20C3M) was chosen as a period of present climate. Periods of future climate were the decades in which each GCM estimated the global-mean surface air temperature increase by 2 K under the SRES A1b condition compared with the present climate. This selection of the different decades may distinguish the uncertainty due to changes in synoptic phenomena from the uncertainty due to the climate sensitivity and emission scenario (Inatsu et al., 2015).

In the DDS, three GCMs of the Coupled Model Inter-comparison Project phase 3 (CMIP3) were chosen as initial and boundary conditions for a RAM of the JMA/Meteorological Research Institute (JMA/MRI) nonhydrostatic model (Saito et al., 2006). The chosen GCMs were the high-resolution version of the Model for Interdisciplinary Research on Climate 3.2 (MIROC; Hasumi and Emori, 2004), the fifth-generation atmospheric GCM of the Max-Planck-Institut für Meteorologie (ECHAM5/MPI; Roeckner et al., 2003), and version 3 of the Community Climate System Model of the National Center for Atmospheric Research (CCSM3/NCAR; Collins et al., 2006). These three GCMs were able to reproduce the present climate around Hokkaido (Kuno and Inatsu, 2014). As for the RAM, the spatial resolution was 10 km and the domain size was ~2.1×10^6 km^2, ranging from 135°E to 150°E and 39°N to 49°N. Mt. Annupuri is not resolved in the topography of the RAM (Fig. 2a). The

DDS was performed for the present climate of 1990s for all three GCMs. The DDS was also performed for the 2050s for MIROC, the 2060s for the MPI model, and the 2080s for the NCAR model, these being the decades in which each GCM estimated that the global-mean surface temperature would have increased by 2 K. For the snowpack calculation, the DDS data corresponding to the nearest grid point to the study area approximately including the three locations of the top, hillside, and base in the same single grid, are used as atmospheric forcing (Fig. 2b). Although the selected grid do not include the base point (Fig. 2b), it is used as the forcing because the same observation data should be used for the bias correction of the following section 3.2. Forced variables are temperature, precipitation, relative humidity, incoming shortwave radiation, and wind.

3.1.2 Observed data

The temperature, precipitation, relative humidity, incoming shortwave radiation, wind, and snow depth observed with the Automated Meteorological Data Acquisition System (AMeDAS) operated by the JMA are basically used for validation of snowpack modeling. The validation was done at Sapporo, because all the meteorological data necessary for the snowpack model run have been operationally observed there, and because a snow pit observation twice a week at Sapporo (Niwano et al., 2012) enables us to validate the model. This snow pit observation measured the grain type of snowpack in depth; the type is classified into precipitation particles, graupel, decomposed precipitation particles, rounded grains, faceted crystals, depth hoar, ice formations, crust, and melt forms. Note that faceted crystals and depth hoar are regarded as a single type of hoar in this study. The temperature and precipitation of AMeDAS data at Kutchan were used for the bias correction for downscaled data.

3.2 Pre-processing of the forcing data

Bias corrections for DDS precipitation and temperature of data are made by comparing present-climate simulations with the JMA's observations at the base point. A temperature bias is defined monthly as the DDS-data climatology minus the observed climatology, and the bias is simply subtracted from the hourly DDS data. The temperatures at the hillside point and at the mountain top point are estimated by the temperature difference from the base decreased by means of the standard lapse rate of 6.5 K/km. As for precipitation, the scaling factor to correct the DDS data is determined month by month from the observed climate at the base point. This scaling factor is loaded for all the downscaled data (Prudhomme et al., 2002). We assumed no difference of precipitation among top, hillside, and base points, because no reference data are available for the hillside and top points. We did not make any pre-processing for other climatic variables.

3.3 SNOWPACK model setup

We used version 3.2.1 of SNOWPACK for step (iii) in the procedure (Fig. 1). SNOWPACK is based on a one-dimensional multi-layered snowpack model and solves the mass balance for water vapor, liquid water, and snow, and the energy balance for snowpack. See Bartelt and Lehning (2002) for more details. This model has some achievement to be applied to Japan and have been suitable for cold regions including Hokkaido (Hirashima et al., 2004; Nakamura et al., 2011; Nishimura et al., 2005). This study applied the NIED scheme (Hirashima et al., 2010) for a better representation of the wet, heavy snow typically observed in Japan. We forced this model with hourly meteorological data of air temperature, relative humidity, wind speed, incoming shortwave radiation, and precipitation at the snow surface. Snowfall is discriminated from rainfall according to a threshold of 1.2 °C in surface air temperature and a threshold of 50% in relative humidity in the model. The volume of precipitation particles is estimated from precipitation with a snow density parameterization slightly modified from that in Lehning et al. (2002a), but this modification is unpublished. Net longwave radiation is estimated from externally given air temperature and snow surface temperature as calculated in the model, because incoming longwave radiation is not prescribed (Lehning et al., 2002a). The soil temperature is fixed at 0 °C. Latent and sensible heat fluxes from snow surface to air are calculated under the Monin–Obukhov bulk formulation (Monin and Obukhov, 1954). The integration period for a single season is from 1 October of a year to 25 June in the following year, and the 10-season integration is done for bias-corrected downscaled data with a particular GCM under present or future climate.

In SNOWPACK, snow grains are classified into eight types with majority and minority forms: precipitation particles, decomposed precipitation particles, rounded grains, faceted crystals, depth hoar, surface hoar, ice formations, and melt forms (Lehning et al., 2002b). The snow grain type is determined by four parameters in the model: dendricity, sphericity, grain size, and grain type history. Precipitation particles have higher dendricity; rounded grains are characterized by higher sphericity, while faceted crystals are characterized by low sphericity. This study regards faceted crystals, depth hoar, surface hoar, and their mixed forms, which were originally differentiated in the model, as a single type of hoar. We also ignore the minority forms. This study focuses on only the ratios of melt forms and hoar as traditional indices of snowpack property, which can be readily validated by a comparison with the snow pit observation because number of layers observed is different with that of layers calculated. The ratio of melt form is useful to diagnose wet avalanches (Techel and Pielmeier, 2010), while the hoar is recognized as one of the factors for dry avalanches at Mt. Annupuri (Nishimura et al., 2005). In the following sections, we define the date of maximum snow depth as the boundary between the “accumulation period” and the “ablation period”.

3.4 Sensitivity experiment

This paper has basically excluded the effect of small-scale topography on the mountain slope and assumed the uniform precipitation field among top, hillside and base points in the pre-processing (section 3.2), mainly because we have no reference data of precipitation at top and hillside points. However, even the small-scale topography more or less contributes to the total amount of precipitation (Houze, 2012), so that the amount depends on the points where we address. Therefore we conduct an additional sensitivity experiment to precipitation at the top in order to discuss an influence of the possible orographic effects on the snowpack estimation. In this sensitivity experiment, the snowpack model ran with the same downscaled data except for hourly precipitation data increasing or decreasing by 20%.

4 Results

4.1 Atmospheric changes

Figure 3 shows the global warming response of the monthly-mean temperature and monthly precipitation at Kutchan. Although the future climate is defined as the decades when global-mean temperature has increased by 2 K compared with 1990s, the DDS results showed a temperature increase of about 2.5 K, probably because of land-sea contrast in the Northern Hemisphere. Remember that DDS with MIROC, MPI, and NCAR GCMs was performed for the future periods of 2050s, 2060s, and 2080s, respectively, and they are compared with the reference of present climate. The DDS results from MIROC show the least month-to-month variation while the NCAR GCM shows the most; the amount of increase is slightly smaller in the MPI case. The precipitation change also has a large seasonal variation, but the total amount of wintertime precipitation does not increase. There is a small tendency toward increased precipitation in January and April, however.

4.2 Validation of the model

We validated the SNOWPACK model by comparing the calculations enforced by the atmospheric variables observed at Sapporo with the snow data observed at Sapporo. Figure 4 shows some snowpack properties at Sapporo for three winters of 2009/2010, 2010/2011, and 2011/2012 (The winter season is between December and the following May). Snow depth is well simulated through the season, except for a slight overestimation in March and April 2011 (Figs. 4a–c). The ratio of melt forms for all snowpack layers is also well simulated, but the simulated ratio in December and January 2011/12 is twice larger than the observed ratio (Figs. 4d–f). The ratio of hoar is overestimated as well, especially in 2010/11 (Figs. 4g–i). SNOWPACK cannot reproduce the hoar realistically as in our experiments, presumably because snowpack surface temperature tends to decrease in the model. It is remarked, however, that we will show the results of hoar by regarding it as the diagnosed quantity that is a function of temperature gradient inside of the snowpack (Lehning et al., 2002b).

The simulated snow depth at the base point with present-climate downscaled data at the nearest grid point to Mt. Annupuri is also compared with the snow depth observed at the JMA observatory at Kutchan. Recall that DDS temperature and precipitation data are bias-corrected but other climatic variables are not. The monthly-mean snow depth is highly correlated with the observations, though it is slightly underestimated in February to March and overestimated in April (Fig. 5a). This overestimation is consistent with an earlier report on the snow ablation period in the Japanese snowy area (Yamaguchi et al., 2004). It is also remarked that, though the validation of the snow depth at the hillside and top points was basically difficult for the paucity of observation, the special observation at the hillside by Nishimura et al. (2005) did not give much difference from our simulated result described below.

4.3 Snowpack estimation

Figure 5 shows the snowpack estimations based on DDS under the present climate. The monthly-mean snow depth attains seasonal maximums of 130 cm at the base point, 190 cm at the hillside point, and 220 cm at the top point. The maximum monthly-mean snow depth increases by about 8 cm per 100 m in altitude. In addition, the greatest depth occurs later at higher altitudes because the freezing environment extends the snowfall period.

Global warming significantly decreases monthly-mean snow depth at all points throughout the season (Fig. 6). From December to February, the snow depth decreases by 30 cm at the base and hillside points and by 20 cm at the top point, because the beginning of snow season is retarded by a warmer climate. The decrease is then larger for the snow ablation period. The snow depth decrease likely exceeds 60 cm at the base in March, at the hillside in April, and at the top in May because the snow ablation period starts much earlier: the time of greatest snow depth is shifted by about a month. These are consistent with obtained a set of PGW experiments for another site near Sapporo (Nakamura et al., 2011).

Figure 7 shows the monthly-mean ratio of melt forms for all snowpack layers in the simulation. Under the present climate, at the base point, melt forms occupy about 20% of all snowpack in November, and the ratio gradually increases throughout the season. This pattern does not change substantially under the future climate, but the increase is slightly faster and the melt forms become dominant about a month earlier. Under the present climate, the hillside and top points have a ratio of melt forms that is about 10–20% during the snow accumulation season. Since snow ablation starts earlier in the future, melt forms are dominant at the hillside point in March, and at the top point between March and April. This is also about a month earlier than in the present-climate case.

Figure 8 shows the ensemble mean of the monthly-mean ratio of hoar. Under the present climate, at the base point, the ratio of hoar is approximately constant, at around 10%, from December to March, and the ratio rapidly declines in the snow-melting months. This pattern is also found in the future climate, but the ratio in mid-winter decreases down to 7%. At the hillside point, the ratio of hoar gradually decreases from March to May under the present climate. The percentage is slightly less in the future-climate case. The ratio of hoar is about 15% at the top point from January to April under the present climate. Interestingly, the future-climate case shows no decrease in the ratio of hoar throughout the season at the top point, though the atmospheric warming weakens the temperature gradient in the snowpack. This may be partly because a temperature increase would have little effect on the physical properties of the snowpack in a sufficiently low temperature environment. This is in line with Inoue and Yokoyama (1998), suggesting that global warming would not reduce hoar in eastern Hokkaido.

4.4 Uncertainty

There is fundamentally little uncertainty in the effect on snow depth of global warming because we ruled out the uncertainty associated with climate sensitivity and the emission scenario (Fig. 6). The snow ablation period is uncertain to some extent, however. For example, the difference in snow depth ranged from 30 to 40 cm in January at the base, primarily because precipitation change is insignificant in the MIROC case and +25% in the MPI and NCAR cases (Fig. 3). Although the given climatic variables in February have less variation among GCM cases, the snow accumulation process may increase the variation of the snow-depth difference; it becomes largest at the snow ablation period (Fig. 6a). The uncertainty in the snow-depth difference at the hillside and top points is also noticeable after March (Figs. 6b, c).

The ratio of melt forms from base to top points before the snow ablation period has a comparable variation among GCMs for both present and future climates (Figs. 7a, b). However, the variation at the base point in November is relatively larger in the future climate. This is probably because a greater temperature increase (Fig. 3) promoted the deformation to melt forms in a relatively warm temperature environment for the MIROC case. In the snow ablation period, the uncertainty in melt forms tends to increase in the future. The ratio at the top in March ranges between 10% and 20% in the present climate but between 20% and 40% in the future (Fig. 7c). The future-climate uncertainty in March is as much as the present-climate uncertainty in April. This timelag of the uncertainty could be related to the earlier start of melting period.

4.5 Sensitivity experiment

The results of sensitivity experiment are shown in figure 9 and 10. As a matter of course, the monthly-mean snow depth increased (decreased) when precipitation uniformly increased (decreased). In the present climate, for example, the snow depth on March added 30 cm more than the reference in +20% precipitation experiment (Figs. 5c, 9a). Because hoar is strongly related to the temperature gradient in snowpack, the ratio of hoar is also sensitive to precipitation (Fig. 10). However, the difference of the snow depth between present and future climate is basically not sensitive to precipitation baseline (Fig. 9b). Similarly, neither the difference of the hoar ratio nor that of melt form is sensitive (Figs. 9c,d,10). The sensitivity experiment then revealed that a systematic tendency of precipitation at a particular point on the mountain slope might only have a secondary effect to the result on the future snowpack change presented here.

5 Discussion

This study has estimated the snowpack response to the global-warming atmosphere in the timing where the global-mean temperature would increase by 2 K. According to the IPCC report (Solomon et al., 2007), the climate sensitivity is 4.3 K in

MIROC, 3.4 K in MPI's GCM, and 2.7 K in NCAR's GCM. The uncertainty in greenhouse gas emissions could also cause a large uncertainty in future surface temperature. In our strategy, fixing the temperature increase by the use of a different decade for each model, we have described the snowpack simulation in a "+2-K world." However, the uncertainty in temperature increase could be linked with the simulated points at different altitudes if the standard atmospheric lapse rate were applied. The temperature difference between hillside and top points is 3 K. Moreover the snow-depth difference between the points is about 30 cm (Figs. 5b, c). This means that a 1-K uncertainty in temperature increase approximately corresponds to a 10-cm uncertainty in monthly-mean snow depth at Mt. Annupuri.

Returning to the discussion of climate sensitivity, if we fixed the decade to the 2050s under the A1b scenario, the uncertainty in temperature among GCMs is 1 K (Solomon et al., 2007; Inatsu et al., 2015) so the uncertainty in snow depth would be 10 cm at Mt. Annupuri because a 1-K uncertainty corresponds to a 10-cm uncertainty. Similarly, by fixing the decade to the 2050s again but taking the average over GCM ensembles, the uncertainty in temperature is 0.6 K between A1b and B1 scenarios around Japan (Shin et al., 2012) so the uncertainty in snow depth would be about 6 cm.

The uncertainty in the snow depth is affected by the uncertainty not only in the temperature increase but also in precipitation change among GCMs and among the scenarios. Now, the uncertainty in the snow depth affected by the uncertainty in precipitation is also roughly estimated by the similar way to the above discussion. First, the uncertainty in precipitation change among GCMs around Japan is approximately 0% to +15% if we fixed the decade of the 2050s under the A1b scenario (Shin et al., 2012). Because a +20% uncertainty in precipitation change approximately corresponds to a 30-cm uncertainty (Fig. 9a), the uncertainty in snow depth is also about 20 cm. In spite of this relation, the uncertainty in the snow depth would not be affected by the uncertainty in precipitation change among the scenarios because its uncertainty is less than a few percent if we fixed the decade of the 2050s (Shin et al., 2012).

Moreover, the source of the uncertainty of snowpack change in the +2 K world may be separated into the uncertainty of temperature increase and others. The temperature increases of the three GCM’s cases approximately show a variety of 1 K throughout the season (Fig. 3). Because a 1-K uncertainty in temperature increase approximately corresponds to 10-cm uncertainty in the snow depth, the temperature variation of 1 K may produce a 10-cm uncertainty in the snow depth decrease. Now, the uncertainty in the snow depth decrease at the top point is approximately 25 cm throughout the season (Fig. 6c), so that 40% of the uncertainty is considered to be affected by the uncertainty in the temperature increase. Considering the large sensitivity of snowpack to temperature and precipitation (López-Moreno et al., 2008; 2013), residual uncertainty of 60%, i.e. 15-cm uncertainty, may be mainly produced by the uncertainty in the precipitation. Similarly, at the base and hillside points, 65% and 35% of the uncertainty may be produced by the uncertainty in the temperature and precipitation, respectively.

This study could also be applied to avalanches at the site. The wet-avalanche in Switzerland often occurs at the timing of first wetting of snowpack and the arrival of melt-water at the bottom (Mitterer et al., 2011). Because melt forms are produced after some parts of the snowpack become wet (Lehning et al., 2002b), a season when melt forms rapidly increase roughly corresponds to a season of wet-avalanche. For Mt. Annupuri, the snowpack model indicates that a season of wet-avalanche under the global warming is at hillside height after February and at the top after March, respectively, probably because melt forms are produced after some parts of the snowpack become wet. Since the dominance of melt forms arrives earlier according to our evaluation of global warming response (Fig. 7), we speculate that wet avalanches at Mt. Annupuri would be likely to occur in an earlier season. As we introduced in section 1, an earlier season of wet-avalanche has been also pointed out in the North America

(Lazar and Williams, 2008) and in France (Castebrunet et al., 2014). It should be noted that it is still uncertain whether this expected shift of wet-avalanche season can be simply applied to Japanese environment.

6 Conclusions

We have evaluated the response to global warming of snow depth and some physical properties of snowpack at the mountain range of Mt. Annupuri in Kutchan, Hokkaido, Japan (Fig. 1), by integrating a numerical snowpack model forced by DDS data with multiple GCMs. First, we validated the numerical snowpack model by comparing the results of the hindcast simulation with observation at Sapporo (17 m above sea level) in three winters of 2009/10, 2010/11, and 2011/12: in particular, we successfully reproduced snow depth at the site with bias-corrected DDS data. The numerical snowpack calculation under present and future climates suggests that monthly-mean snow depth will decrease by about 60 cm at the beginning of ablation period if the global- and local-mean temperature increases by 2 K and approximately 2.5 K, respectively (Figs. 3, 6). In addition, monthly-mean snow depth reaches its peak about one month earlier. The monthly-mean ratio of melt forms tends to increase at all sites, especially above the hillside point, while the monthly-mean ratio of hoar is likely to decrease except at the top point.

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Figures

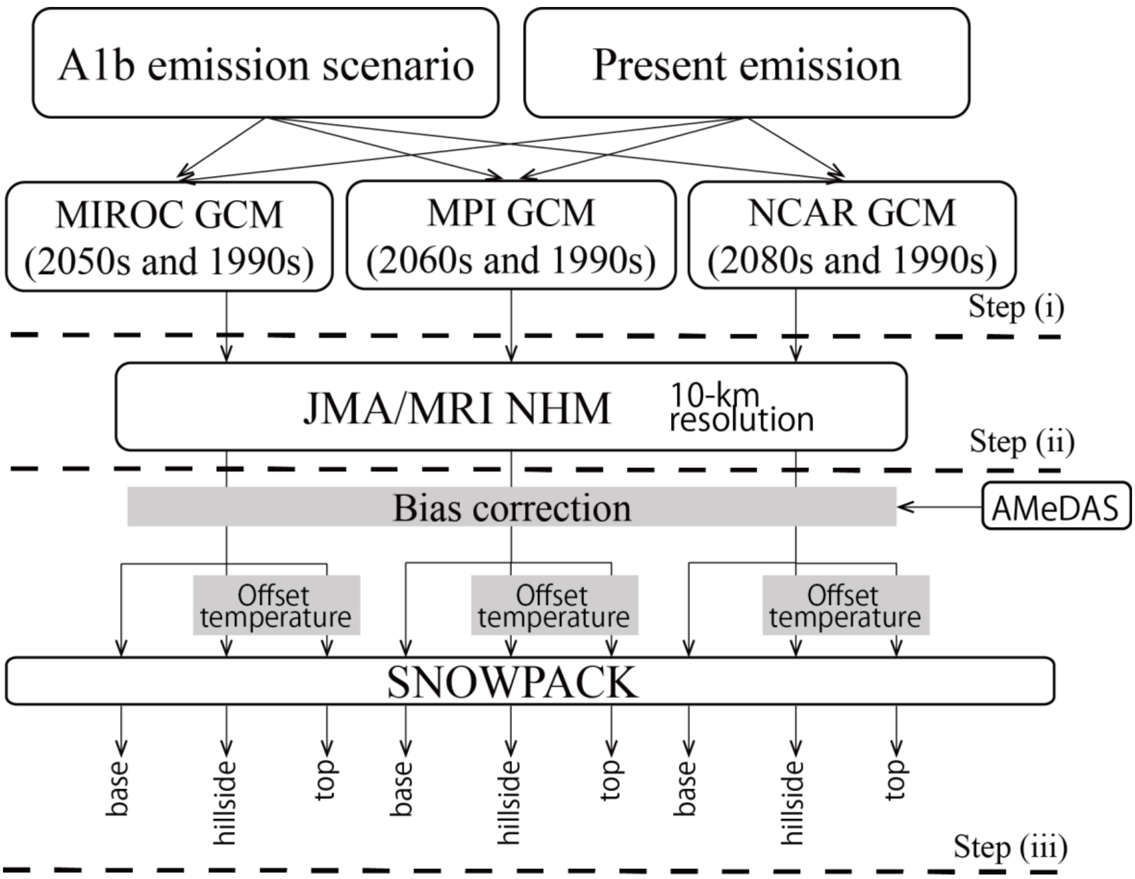

Fig. 1 Flow chart of procedure to assess the future change of snowpack used in this study.

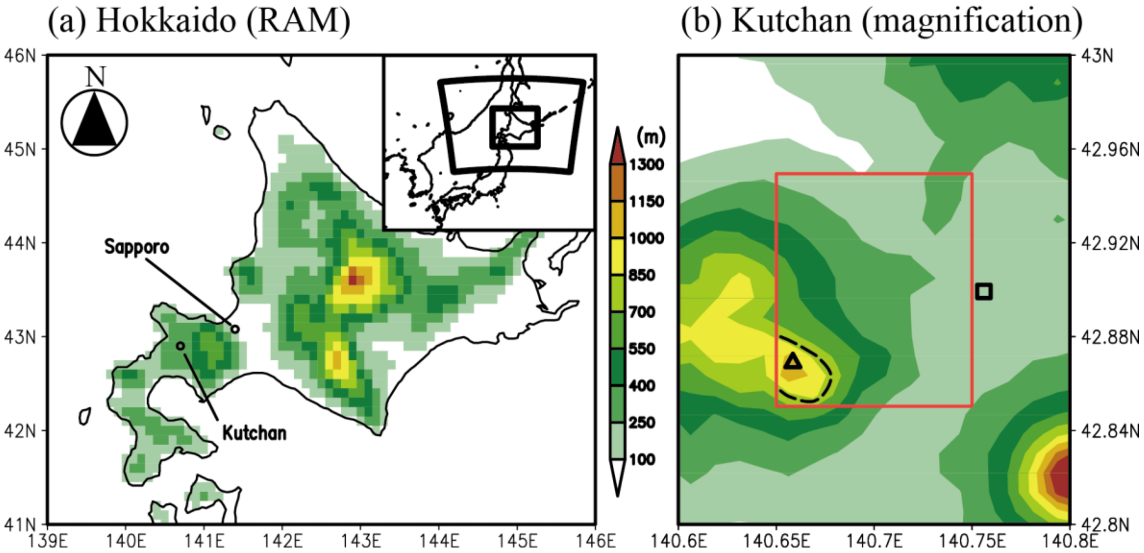

Fig. 2 (a) Surface height above the sea level given to the regional atmospheric model (RAM) and the location of Hokkaido in the upper right and (b) realistic topography with about 1-km resolution. Outer and inner black solid line in a window of (a) shows RAM's domain calculated and a domain of (a), respectively. The locations of top of Mt. Annupuri (the top; 1,300 m a.s.l.) and AMeDAS station (the base; 173 m a.s.l.) are respectively indicated with an open triangle and square in (b). Black dashed line in (b) shows an 800 m level of height corresponding to the hillside of the mountain slope. Red rectangle in (b) shows the RAM's grid cell of which meteorological data are imposed to the SNOWPACK model. The color-scale is shown between the panels.

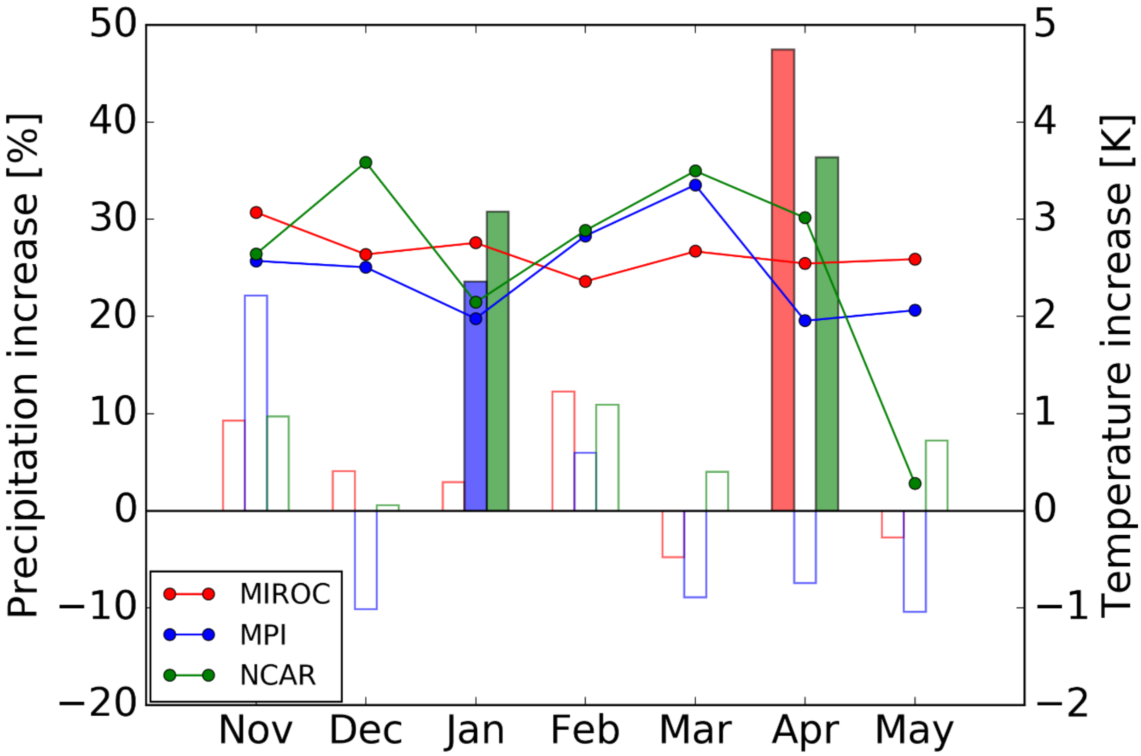

Fig. 3 Global warming response at Kutchan for November to May, based on the dynamical downscaling (DDS) results from (red) the high-resolution version of the Model for Interdisciplinary Research on Climate 3.2 (MIROC), (blue) the fifth-generation atmospheric general-circulation model (GCM) of the Max-Planck-Institut für Meteorologie (ECHAM5/MPI), and (green) version 3 of the Community Climate System Model of the National Center for Atmospheric Research (CCSM3/NCAR). The bar graph shows the increasing rate of monthly precipitation [%; scale on the left] with filled bins denoting a precipitation increase statistically significant at the 10% level. The line graph shows the increase in monthly-mean temperature [K;

scale on the right].

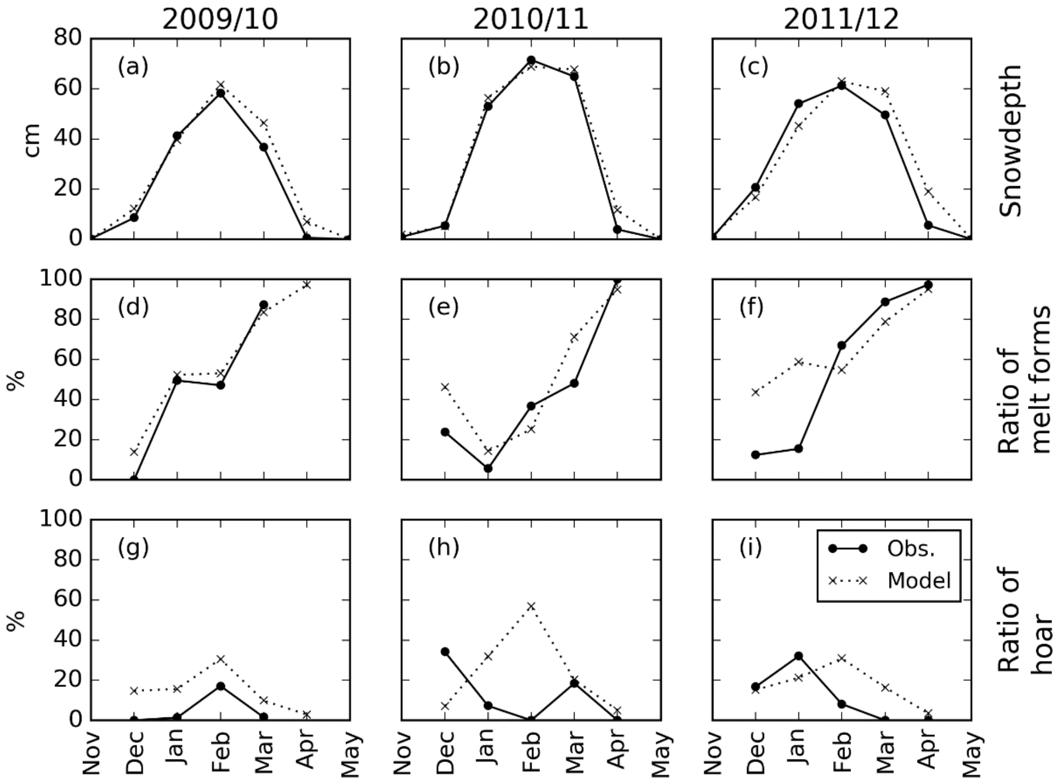

Fig. 4 Monthly-mean (a–c) snow depth and the ratios of (d–f) melt forms and (g–i) hoar for all snowpack layers at Sapporo, averaged over the winters of (a, d, g) 2009/10, (b, e, h) 2010/11, and (c, f, i) 2011/12. Solid lines with circles show (a-c) AMeDAS and (d–i) snow pit observations; dotted lines with crosses show the SNOWPACK model results.

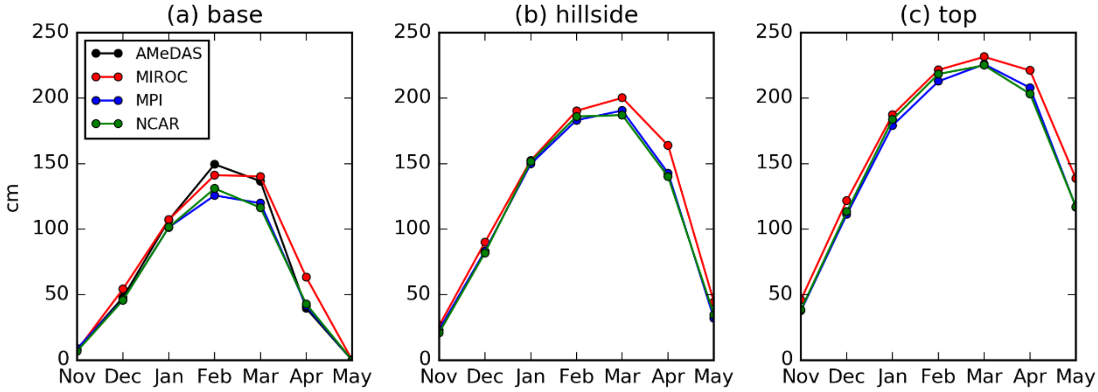

Fig. 5 Snowpack simulation results with DDS data under the present climate for (red) MIROC, (green) MPI and (blue) NCAR GCMs. Panels show monthly-mean snow depth on Mt. Annupuri at (a) the base point at 173 m above sea level, (b) the hillside point at 800 m, and (c) the top point at 1,300 m. Snow depth observed at the JMA's site at Kutchan is superimposed on (a) in black.

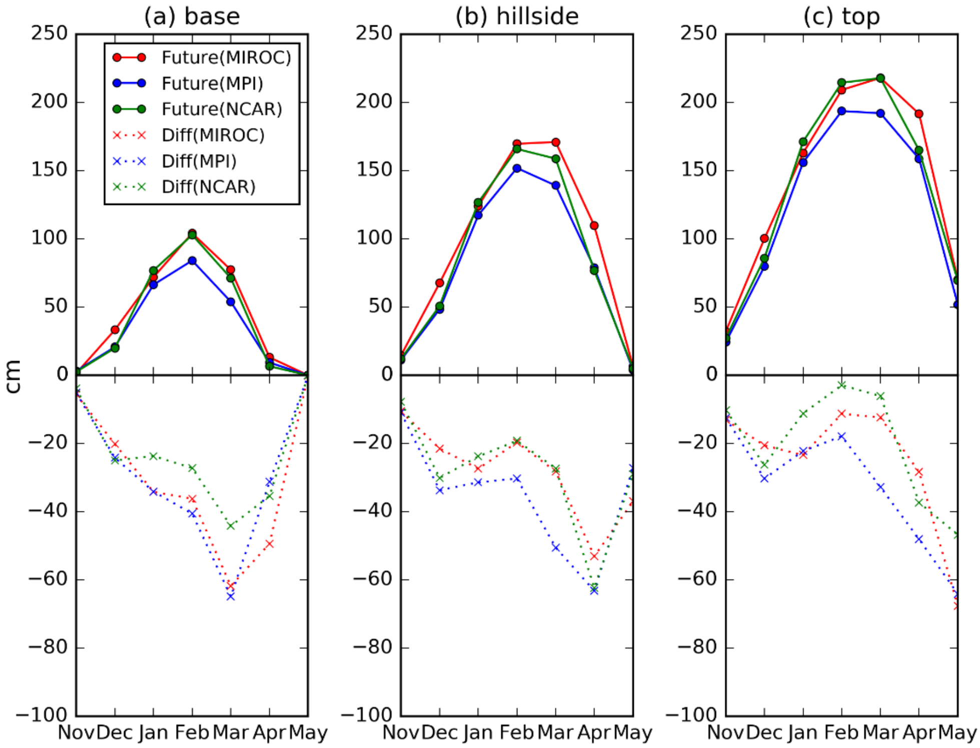

Fig. 6 Monthly-mean snow depth in future climate (solid lines) and the difference between present and future climates (dotted lines) at (a) base, (b) hillside, and (c) top points on Mt. Annupuri, based on the DDS data for (red) MIROC, (green) MPI and (blue) NCAR GCMs.

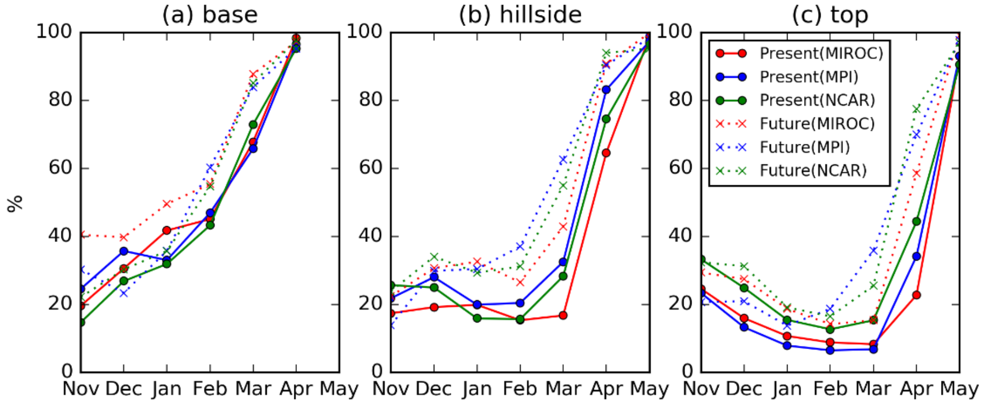

Fig. 7 The ratio of melt forms at (a) base, (b) hillside, and (c) top points, based on the DDS data under (solid line) present and (dotted) future climates.

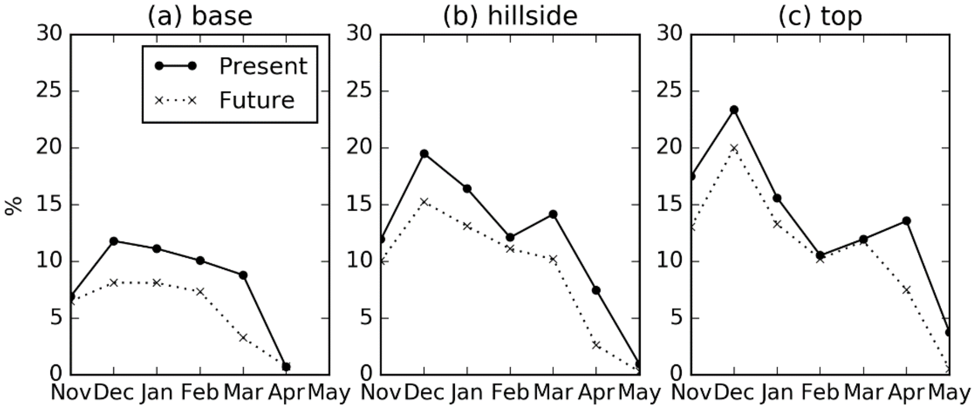

Fig. 8 The ratio of hoar at (a) base, (b) hillside, and (c) top, based on the DDS data averaged over all GCM cases under (solid line) present and (dotted) future climates.

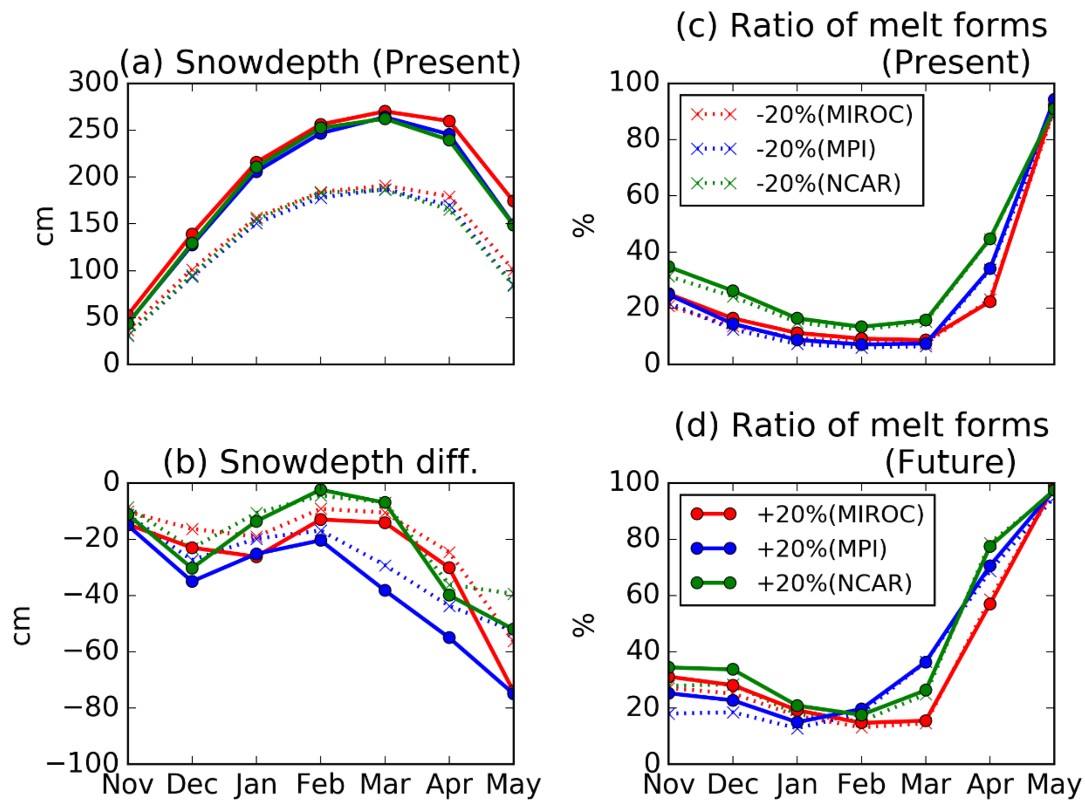

Fig. 9 (a) The monthly-mean snow depth in present climate in (solid line) +20% and (dotted) −20% precipitation experiments. Red, blue, and green lines indicate MIROC's, MPI's, and NCAR's case, respectively. (b) The difference of the snow depth between present and future climate. (c,d) The ratio of melt forms in (c) present and (d) future climate.

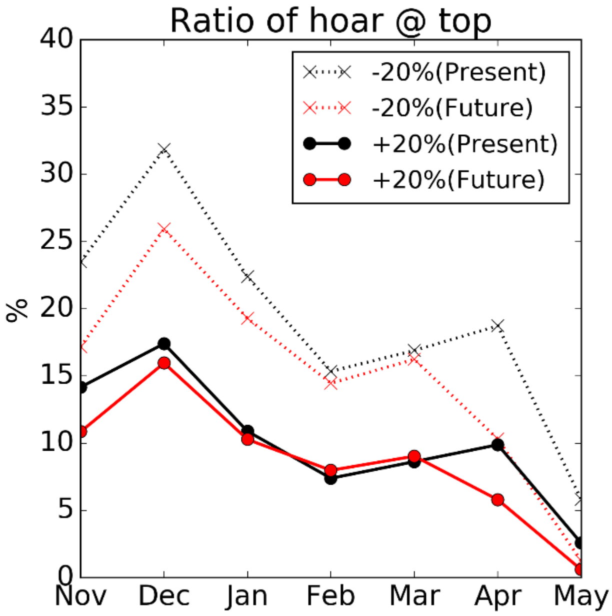

Fig. 10 The ratio of hoar averaged over all GCMs in (black) present and (red) future climate in (solid line) +20% and (dotted line) −20% precipitation experiments.