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1	Global warming response of snowpack at mountain range in
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3	data
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16 Abstract

17 We estimate the response of snowpack to global warming along with the uncertainty of 18 the snowpack change by using a combination of multiple general-circulation models 19 (GCMs), a single regional atmospheric model, and a one-dimensional multi-layered 20 snowpack model. The target site is Mt. Annupuri in Kutchan, Hokkaido, Japan. The 21 forcing of the snowpack model is taken from dynamically downscaled data from GCMs 22 for the present climate and GCMs in a decade when the global-mean temperature has 23 increased by 2 K from present conditions. The results show that global warming would 24 decrease the monthly-mean snow depth throughout the winter season. Other salient 25 features are the decrease of snow depth by 60 cm with maximum uncertainty of 20 cm at 26 the beginning of the snow ablation period, the occurrence of the snow-depth peak a month 27 earlier, and the dominance of melt forms in an earlier season. The ratio of melt forms for 28 all snowpack layers increase with little uncertainty before the snow ablation period. The 29 ratio of hoar does not change much, even though the air temperature increases. The 30 uncertainty in snowpack evaluation is also discussed.

31

32 Keywords

33 Snowpack modeling; Snowpack change; Snow type change; Uncertainty

35 1 Introduction

36 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 37 38 global warming especially at mid-latitudes. According to the latest report of the 39 Inter-governmental Panel for Climate Change (IPCC), the emission of greenhouse gases 40 leads not only to an increase in air temperature but also changes in wind, precipitation 41 amount and intensity, and cloudiness (Stocker et al., 2014). The temperature is one of 42 the most important variables controlling the amount of snow, the properties of 43 snowpack, and the snow cover period, because snow and ice are retained below freezing 44 but rapidly melt above this temperature (López-Moreno et al., 2008, 2013). Moreover, 45 wintertime snowfall, which represents the major contribution to total snow 46 accumulation, is frequently observed in storm-track areas, such as northern Europe, 47 northeastern America, and northern Japan. Because this snowfall is often caused by 48 extratropical cyclone passage, it might be changed by global warming. The physical 49 properties of snowpack might be also changed as a result of efficient transformation 50 from solid precipitation particles to melt forms in a warmer climate. This transformation 51 of grain types may also change the occurrence of wet-avalanche because that of 52 wet-avalanche is related to the wetting of snowpack (Mitterer et al., 2011). Although the

53	wet-avalanche sometimes occurs in Japan (Akitaya et al., 2015) and changes of the
54	wet-avalanche occurrence have been pointed out in the North America (Lazar and
55	Williams, 2008) and in France (Castebrunet et al., 2014), the transformation of grain
56	types responding to the global warming in Japan is not well addressed. Moreover,
57	considering the utilization of water resources (Beniston, 2003), the mitigation of snow
58	disaster (Nakai et al., 2012), and winter tourism (Beniston, 2003; Uhlmann et al., 2009),
59	the impact of climate change on snowpack dynamics deserve to be examined (e.g.
60	Niwano et al., 2012; Mellander et al., 2007; Inatsu et al., 2016), even though there is
61	still inherent uncertainty in models of the impact of climate change.
62	We found several studies of snowpack response to climatic change, such as an
63	estimation of snowpack including its physical property along with its uncertainty by the
64	use of one-dimensional and multi-layered snowpack model (e.g. Rasmus et al., 2004;
65	Lazar et al., 2006; Rousselot et al., 2012). They consist mostly of three steps: (i) climate
66	change projection with atmosphere-ocean general-circulation models (GCMs), in which
67	atmospheric concentrations of greenhouse gases are prescribed as a function of year
68	(Solomon et al., 2007; Stocker et al., 2014); (ii) downscaling, which creates climatic
69	variables at a particular site or in a limited area with higher spatial resolution in order to

71	2004); and (iii) snowpack estimates, either with a physical model (e.g. Rasmus et al.,
72	2004) or with a statistical relation that has been empirically determined in advance
73	(Inoue and Yokoyama, 1998). For step (i), there are no other technical choices than
74	using GCMs. The GCM projection introduces an uncertainty in evaluating global-mean
75	temperature, however, because of differing climate sensitivities among GCMs, mainly
76	due to their physical parameterizations, and because future greenhouse gas emissions
77	depend on the socio-economic scenario (Stocker et al., 2014). The GCM projections of
78	wintertime temperature and precipitation at a particular mid-latitude site are also
79	uncertain, due to uncertain changes in storm tracks and jet streams at mid-latitudes
80	(Chang et al., 2012), wintertime Asian monsoon (Ogata et al., 2014), and the Arctic
81	Oscillation (Karpechko, 2010). These regional climate patterns certainly affect snow
82	accumulation and melt dynamics, and also snowpack dynamics. It should be noted that
83	GCMs are more or less biased, so one needs a bias correction for a particular site. One
84	then proceeds to step (ii) based on coarse-resolution GCM projection with uncertainty
85	and bias. The methods of step (ii) can be classified into dynamical downscaling (DDS)
86	(Wang et al., 2004) and statistical downscaling (Wilby et al., 2004). The former
87	provides higher-resolution climatic variables in a limited area by integrating a regional
88	atmospheric model (RAM) with the GCM output imposed as its lateral boundary

89	condition. The latter estimates a future state by simply applying a statistical relation
90	between local-site information and weather patterns such as the Siberian-Japan pattern
91	that brings much snowfall along the Japan Sea side of northern Japan (Takano et al.,
92	2008). The Siberian-Japan pattern is established based on the singular-value
93	decomposition analysis between synoptic weather pattern and local precipitation, for
94	example (Kuno and Inatsu, 2014). Recently, DDS has been widely used in the
95	community, in spite of the need for additional computation, because it has the ability to
96	produce a physically consistent dataset (e.g. Wang et al., 2004; Kuno and Inatsu, 2014;
97	Inatsu et al., 2015). Multiple RAMs, even with a single GCM imposed as the lateral
98	boundary condition, also provide uncertainty, mainly due to the variability among the
99	RAMs' physical parameterizations, but the uncertainty is not large for the extratropics in
100	winter because the DDS results are strongly controlled by lateral boundary conditions
101	(Inatsu et al., 2015; Kuno and Inatsu, 2014). In step (ii) when using DDS, a bias
102	correction should be made just before step (iii) because the DDS results have the
103	systematic biases in atmospheric variables such as temperature, precipitation, and so on,
104	due to physical parameterizations and resolution (Ishizaki et al., 2012). It should be
105	noted that an alternative choice in step (ii) is the pseudo-global warming (PGW)
106	experiment, in which observed weather time-series are added to the climatological

107	difference estimated from GCM integrations so as to form the lateral boundary
108	condition of the RAM (Kimura and Kitoh, 2007). Finally step (iii) estimates the future
109	snowpack change, which is still a challenging problem. Although Inoue and Yokoyama
110	(1998) estimated maximum snow depth and major snow type over Japan by using a
111	statistical relation between snowpack and meteorological characteristics, recent studies
112	have tended to use one-dimensional multi-layered snowpack models, such as CROCUS
113	(Brun et al., 1992), SNTHERM (Jordan, 1991), and SNOWPACK (Bartelt and Lehning,
114	2002) and its modification for wet-heavy snow (Hirashima, 2014). A one-dimensional
115	multi-layered snowpack model enables us to calculate the temporal evolution of
116	snowpack structure with multiple layers at a particular site, driven by atmospheric
117	variables, such as air temperature, precipitation, humidity, wind, and shortwave
118	radiation at the snow surface. Step (iii) is, therefore, undertaken on the basis of
119	bias-corrected atmospheric variables obtained from step (ii).

120 Several previous studies have been devoted to an evaluation of future snowpack change for particular areas, such as Switzerland (Bavay et al., 2009, 2013), 121 122 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 123 124 multiple emission scenarios of greenhouse gases. The emission scenario strongly

125	controls the global-mean temperature increase, a factor to which snowpack estimation is
126	sensitive. For example, Rousselot et al. (2012) revealed that the snow water equivalent
127	change in the A2 scenario of the Special Report on Emissions Scenarios (SRES) was
128	double to that in the B1 scenario. Bavay et al. (2009) also pointed out a great
129	discrepancy between results for the A2 and B2 scenarios. Moreover, under the same
130	emission scenario, different GCMs provide different global-mean temperature increases
131	due to climate sensitivity. The use of multiple GCMs, therefore, increases the range of
132	estimates of future snowpack (Bavay et al., 2013; Lazar et al., 2006; Rasmus et al.,
133	2004).

134 When considering the effect of climate change on snowpack, one may desire 135 to separate the changes due to differences in global-mean temperature from those due to 136 changes of synoptic-scale climate around a particular site. However, since the 137 temperature increase affects snowpack estimation quite strongly, it is difficult to 138 determine the uncertainty arising from changes of meso-scale convection, storm tracks, 139 and quasi-stationary pressure patterns by using a set of arbitrarily chosen GCMs. 140 From the point of view of numerical snowpack experiments, biased input data 141 can cause problems. For example, a warm bias would shorten the snow season and a dry

142 bias would effectively decrease the snow depth. Hence, we need a rational treatment for

143 bias in GCMs and downscaled data. One way to ameliorate this problem is to offset 144 climatological differences between present and future conditions assuming that the 145 model biases are stationary. This approach has been used in several studies. Rasmus et 146 al. (2004) calculated the snowpack difference for present and future climates using DDS 147 without any bias correction of input data. Bavay et al. (2009; 2013) and Nakamura et al. 148 (2011) estimated a change in snowpack using the PGW strategy. It is also expected that 149 statistical downscaling will correct for any bias without requiring additional procedures 150 (Rousselot et al., 2012; Lazar et al., 2006).

151 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 152 153 one-dimensional snowpack model (Fig. 1). The analysis is based on the idea proposed 154 in Inatsu et al. (2015), in which the synoptic-scale response was successfully separated 155 from global temperature increase by performing DDS for a decade during which the 156 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 157 158 pre-processing before the snowpack calculation, we make bias corrections for 159 temperature and precipitation and height correction for temperature in order to discuss 160 differences of the snowpack response with altitude. After pre-processing, the numerical

161 snowpack calculation is performed for a particular mountain range at Mt. Annupuri in
162 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.

170

171 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.

179	Takano et al., 2008). The climatological feature holds the snow-cover period exceeding
180	4 months and a maximum snow depth of 190 cm even at the mountain base. Second,
181	this site encompasses the mountain top, at 1,308 m above sea level, down to a wide
182	steep hill with its base around 200 m above sea level (Fig. 2b). The large difference of
183	height at a single mountain enables us to facilitate the discussion on snowpack change
184	with different temperature baselines. For the estimation, we considered three locations:
185	the top (1,300 m), the hillside (800 m), and the base (173 m), which is the level of the
186	JMA's meteorological station (Fig. 2b). The site is the downwind side of winter
187	monsoon that brings heavy snowfall. The slope of Mt. Annupuri directs from southwest
188	to northeast and the part of mountain area is leeward, but we do not consider the effect
189	of such small-scale topography on the mountain slope. Finally, there is a social demand
190	for the estimation because a famous ski resort with high-quality snow is located at the
191	site.
192	
193	3 Data and Methods
194	3.1 Data
195	3.1.1 Downscaled data

196 We used a dataset of DDS results provided by Kuno and Inatsu (2014). For

197	this dataset, the 1990s in the 20th century experiment (20C3M) was chosen as a period
198	of present climate. Periods of future climate were the decades in which each GCM
199	estimated the global-mean surface air temperature increase by 2 K under the SRES A1b
200	condition compared with the present climate. This selection of the different decades
201	may distinguish the uncertainty due to changes in synoptic phenomena from the
202	uncertainty due to the climate sensitivity and emission scenario (Inatsu et al., 2015).
203	In the DDS, three GCMs of the Coupled Model Inter-comparison Project
204	phase 3 (CMIP3) were chosen as initial and boundary conditions for a RAM of the
205	JMA/Meteorological Research Institute (JMA/MRI) nonhydrostatic model (Saito et al.,
206	2006). The chosen GCMs were the high-resolution version of the Model for
207	Interdisciplinary Research on Climate 3.2 (MIROC; Hasumi and Emori, 2004), the
208	fifth-generation atmospheric GCM of the Max-Planck-Institut für Meteorologie
209	(ECHAM5/MPI; Roeckner et al., 2003), and version 3 of the Community Climate
210	System Model of the National Center for Atmospheric Research (CCSM3/NCAR;
211	Collins et al., 2006). These three GCMs were able to reproduce the present climate
212	around Hokkaido (Kuno and Inatsu, 2014). As for the RAM, the spatial resolution was
213	10 km and the domain size was ~2.1×10 ⁶ km ² , ranging from 135°E to 150°E and 39°N
214	to 49°N. Mt. Annupuri is not resolved in the topography of the RAM (Fig. 2a). The

215	DDS was performed for the present climate of 1990s for all three GCMs. The DDS was
216	also performed for the 2050s for MIROC, the 2060s for the MPI model, and the 2080s
217	for the NCAR model, these being the decades in which each GCM estimated that the
218	global-mean surface temperature would have increased by 2 K. For the snowpack
219	calculation, the DDS data corresponding to the nearest grid point to the study area
220	approximately including the three locations of the top, hillside, and base in the same
221	single grid, are used as atmospheric forcing (Fig. 2b). Although the selected grid do not
222	include the base point (Fig. 2b), it is used as the forcing because the same observation
223	data should be used for the bias correction of the following section 3.2. Forced variables
224	are temperature, precipitation, relative humidity, incoming shortwave radiation, and
225	wind.
226	
227	3.1.2 Observed data
228	The temperature, precipitation, relative humidity, incoming shortwave
229	radiation, wind, and snow depth observed with the Automated Meteorological Data
230	Acquisition System (AMeDAS) operated by the JMA are basically used for validation
231	of snowpack modeling. The validation was done at Sapporo, because all the
232	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.

240

241 3.2 Pre-processing of the forcing data

242 Bias corrections for DDS precipitation and temperature of data are made by 243 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 244 245 climatology, and the bias is simply subtracted from the hourly DDS data. The 246 temperatures at the hillside point and at the mountain top point are estimated by the 247 temperature difference from the base decreased by means of the standard lapse rate of 248 6.5 K/km. As for precipitation, the scaling factor to correct the DDS data is determined 249 month by month from the observed climate at the base point. This scaling factor is 250 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.

- 254
- 255 3.3 SNOWPACK model setup

256 We used version 3.2.1 of SNOWPACK for step (iii) in the procedure (Fig. 1). 257 SNOWPACK is based on a one-dimensional multi-layered snowpack model and solves 258 the mass balance for water vapor, liquid water, and snow, and the energy balance for 259 snowpack. See Bartelt and Lehning (2002) for more details. This model has some 260 achievement to be applied to Japan and have been suitable for cold regions including 261 Hokkaido (Hirashima et al., 2004; Nakamura et al., 2011; Nishimura et al., 2005). This 262 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 263 264 meteorological data of air temperature, relative humidity, wind speed, incoming shortwave radiation, and precipitation at the snow surface. Snowfall is discriminated 265 266 from rainfall according to a threshold of 1.2 °C in surface air temperature and a 267 threshold of 50% in relative humidity in the model. The volume of precipitation particles is estimated from precipitation with a snow density parameterization slightly 268

269	modified from that in Lehning et al. (2002a), but this modification is unpublished. Net
270	longwave radiation is estimated from externally given air temperature and snow surface
271	temperature as calculated in the model, because incoming longwave radiation is not
272	prescribed (Lehning et al., 2002a). The soil temperature is fixed at 0 °C. Latent and
273	sensible heat fluxes from snow surface to air are calculated under the Monin-Obukhov
274	bulk formulation (Monin and Obukhov, 1954). The integration period for a single
275	season is from 1 October of a year to 25 June in the following year, and the 10-season
276	integration is done for bias-corrected downscaled data with a particular GCM under
277	present or future climate.

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278 In SNOWPACK, snow grains are classified into eight types with majority and 279 minority forms: precipitation particles, decomposed precipitation particles, rounded 280 grains, faceted crystals, depth hoar, surface hoar, ice formations, and melt forms 281 (Lehning et al., 2002b). The snow grain type is determined by four parameters in the 282 model: dendricity, sphericity, grain size, and grain type history. Precipitation particles 283 have higher dendricity; rounded grains are characterized by higher sphericity, while 284 faceted crystals are characterized by low sphericity. This study regards faceted crystals, 285 depth hoar, surface hoar, and their mixed forms, which were originally differentiated in 286 the model, as a single type of hoar. We also ignore the minority forms. This study

287	focuses on only the ratios of melt forms and hoar as traditional indices of snowpack
288	property, which can be readily validated by a comparison with the snow pit observation
289	because number of layers observed is different with that of layers calculated. The ratio
290	of melt form is useful to diagnose wet avalanches (Techel and Pielmeier, 2010), while
291	the hoar is recognized as one of the factors for dry avalanches at Mt. Annupuri
292	(Nishimura et al., 2005). In the following sections, we define the date of maximum
293	snow depth as the boundary between the "accumulation period" and the "ablation
294	period".

295

296 3.4 Sensitivity experiment

297 This paper has basically excluded the effect of small-scale topography on the 298 mountain slope and assumed the uniform precipitation field among top, hillside and 299 base points in the pre-processing (section 3.2), mainly because we have no reference 300 data of precipitation at top and hillside points. However, even the small-scale 301 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 302 303 additional sensitivity experiment to precipitation at the top in order to discuss an 304 influence of the possible orographic effects on the snowpack estimation. In this 305 sensitivity experiment, the snowpack model ran with the same downscaled data except306 for hourly precipitation data increasing or decreasing by 20%.

307

308 4 Results

309 4.1 Atmospheric changes

310 Figure 3 shows the global warming response of the monthly-mean 311 temperature and monthly precipitation at Kutchan. Although the future climate is 312 defined as the decades when global-mean temperature has increased by 2 K compared 313 with 1990s, the DDS results showed a temperature increase of about 2.5 K, probably 314 because of land-sea contrast in the Northern Hemisphere. Remember that DDS with 315 MIROC, MPI, and NCAR GCMs was performed for the future periods of 2050s, 2060s, 316 and 2080s, respectively, and they are compared with the reference of present climate. 317 The DDS results from MIROC show the least month-to-month variation while the 318 NCAR GCM shows the most; the amount of increase is slightly smaller in the MPI case. 319 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 320 321 precipitation in January and April, however.

323 4.2 Validation of the model

We validated the SNOWPACK model by comparing the calculations enforced 324 325 by the atmospheric variables observed at Sapporo with the snow data observed at 326 Sapporo. Figure 4 shows some snowpack properties at Sapporo for three winters of 327 2009/2010, 2010/2011, and 2011/2012 (The winter season is between December and the 328 following May). Snow depth is well simulated through the season, except for a slight 329 overestimation in March and April 2011 (Figs. 4a-c). The ratio of melt forms for all 330 snowpack layers is also well simulated, but the simulated ratio in December and January 331 2011/12 is twice larger than the observed ratio (Figs. 4d-f). The ratio of hoar is 332 overestimated as well, especially in 2010/11 (Figs. 4g-i). SNOWPACK cannot 333 reproduce the hoar realistically as in our experiments, presumably because snowpack 334 surface temperature tends to decrease in the model. It is remarked, however, that we will 335 show the results of hoar by regarding it as the diagnosed quantity that is a function of 336 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

341	monthly-mean snow depth is highly correlated with the observations, though it is
342	slightly underestimated in February to March and overestimated in April (Fig. 5a). This
343	overestimation is consistent with an earlier report on the snow ablation period in the
344	Japanese snowy area (Yamaguchi et al., 2004). It is also remarked that, though the
345	validation of the snow depth at the hillside and top points was basically difficult for the
346	paucity of observation, the special observation at the hillside by Nishimura et al. (2005)
347	did not give much difference from our simulated result described below.
348	
349	4.3 Snowpack estimation
350	Figure 5 shows the snowpack estimations based on DDS under the present
351	climate. The monthly-mean snow depth attains seasonal maximums of 130 cm at the
352	base point, 190 cm at the hillside point, and 220 cm at the top point. The maximum
353	monthly-mean snow depth increases by about 8 cm per 100 m in altitude. In addition,
354	the greatest depth occurs later at higher altitudes because the freezing environment
355	extends the snowfall period.
356	Global warming significantly decreases monthly-mean snow depth at all
357	points throughout the season (Fig. 6). From December to February, the snow depth
358	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).

365 Figure 7 shows the monthly-mean ratio of melt forms for all snowpack layers 366 in the simulation. Under the present climate, at the base point, melt forms occupy about 367 20% of all snowpack in November, and the ratio gradually increases throughout the 368 season. This pattern does not change substantially under the future climate, but the 369 increase is slightly faster and the melt forms become dominant about a month earlier. 370 Under the present climate, the hillside and top points have a ratio of melt forms that is 371 about 10–20% during the snow accumulation season. Since snow ablation starts earlier 372 in the future, melt forms are dominant at the hillside point in March, and at the top point 373 between March and April. This is also about a month earlier than in the present-climate 374 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

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

389

390 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

395	at the base, primarily because precipitation change is insignificant in the MIROC case
396	and +25% in the MPI and NCAR cases (Fig. 3). Although the given climatic variables
397	in February have less variation among GCM cases, the snow accumulation process may
398	increase the variation of the snow-depth difference; it becomes largest at the snow
399	ablation period (Fig. 6a). The uncertainty in the snow-depth difference at the hillside
400	and top points is also noticeable after March (Figs. 6b, c).
401	The ratio of melt forms from base to top points before the snow ablation
402	period has a comparable variation among GCMs for both present and future climates
403	(Figs. 7a, b). However, the variation at the base point in November is relatively larger in
404	the future climate. This is probably because a greater temperature increase (Fig. 3)
405	promoted the deformation to melt forms in a relatively warm temperature environment
406	for the MIROC case. In the snow ablation period, the uncertainty in melt forms tends to
407	increase in the future. The ratio at the top in March ranges between 10% and 20% in the
408	present climate but between 20% and 40% in the future (Fig. 7c). The future-climate
409	uncertainty in March is as much as the present-climate uncertainty in April. This timelag
410	of the uncertainty could be related to the earlier start of melting period.
411	

412 4.5 Sensitivity experiment

413 The results of sensitivity experiment are shown in figure 9 and 10. As a matter 414 of course, the monthly-mean snow depth increased (decreased) when precipitation 415 uniformly increased (decreased). In the present climate, for example, the snow depth on 416 March added 30 cm more than the reference in +20% precipitation experiment (Figs. 5c, 417 9a). Because hoar is strongly related to the temperature gradient in snowpack, the ratio 418 of hoar is also sensitive to precipitation (Fig. 10). However, the difference of the snow 419 depth between present and future climate is basically not sensitive to precipitation 420 baseline (Fig. 9b). Similarly, neither the difference of the hoar ratio nor that of melt 421 form is sensitive (Figs. 9c,d,10). The sensitivity experiment then revealed that a 422 systematic tendency of precipitation at a particular point on the mountain slope might 423 only have a secondary effect to the result on the future snowpack change presented here. 424

425 5 Discussion

426

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

430	MIROC, 3.4 K in MPI's GCM, and 2.7 K in NCAR's GCM. The uncertainty in
431	greenhouse gas emissions could also cause a large uncertainty in future surface
432	temperature. In our strategy, fixing the temperature increase by the use of a different
433	decade for each model, we have described the snowpack simulation in a "+2-K world."
434	However, the uncertainty in temperature increase could be linked with the simulated
435	points at different altitudes if the standard atmospheric lapse rate were applied. The
436	temperature difference between hillside and top points is 3 K. Moreover the snow-depth
437	difference between the points is about 30 cm (Figs. 5b, c). This means that a 1-K
438	uncertainty in temperature increase approximately corresponds to a 10-cm uncertainty
439	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.

447

The uncertainty in the snow depth is affected by the uncertainty not only in

448	the temperature increase but also in precipitation change among GCMs and among the
449	scenarios. Now, the uncertainty in the snow depth affected by the uncertainty in
450	precipitation is also roughly estimated by the similar way to the above discussion. First,
451	the uncertainty in precipitation change among GCMs around Japan is approximately 0%
452	to +15% if we fixed the decade of the 2050s under the A1b scenario (Shin et al., 2012).
453	Because a +20% uncertainty in precipitation change approximately corresponds to a
454	30-cm uncertainty (Fig. 9a), the uncertainty in snow depth is also about 20 cm. In spite
455	of this relation, the uncertainty in the snow depth would not be affected by the
456	uncertainty in precipitation change among the scenarios because its uncertainty is less
457	than a few percent if we fixed the decade of the 2050s (Shin et al., 2012).
458	Moreover, the source of the uncertainty of snowpack change in the +2 K
459	world may be separated into the uncertainty of temperature increase and others. The
460	temperature increases of the three GCM's cases approximately show a variety of 1 K
461	throughout the season (Fig. 3). Because a 1-K uncertainty in temperature increase
462	approximately corresponds to 10-cm uncertainty in the snow depth, the temperature
463	variation of 1 K may produce a 10-cm uncertainty in the snow depth decrease. Now, the
464	uncertainty in the snow depth decrease at the top point is approximately 25 cm
465	throughout the season (Fig. 6c), so that 40% of the uncertainty is considered to be

466 affected by the uncertainty in the temperature increase. Considering the large sensitivity 467 of snowpack to temperature and precipitation (López-Moreno et al., 2008; 2013), 468 residual uncertainty of 60%, i.e. 15-cm uncertainty, may be mainly produced by the 469 uncertainty in the precipitation. Similarly, at the base and hillside points, 65% and 35% 470 of the uncertainty may be produced by the uncertainty in the temperature and 471 precipitation, respectively.

472 This study could also be applied to avalanches at the site. The wet-avalanche 473 in Switzerland often occurs at the timing of first wetting of snowpack and the arrival of 474 melt-water at the bottom (Mitterer et al., 2011). Because melt forms are produced after 475 some parts of the snowpack become wet (Lehning et al., 2002b), a season when melt 476 forms rapidly increase roughly corresponds to a season of wet-avalanche. For Mt. 477 Annupuri, the snowpack model indicates that a season of wet-avalanche under the 478 global warming is at hillside height after February and at the top after March, 479 respectively, probably because melt forms are produced after some parts of the 480 snowpack become wet. Since the dominance of melt forms arrives earlier according to 481 our evaluation of global warming response (Fig. 7), we speculate that wet avalanches at 482 Mt. Annupuri would be likely to occur in an earlier season. As we introduced in section 483 1, an earlier season of wet-avalanche has been also pointed out in the North America 484 (Lazar and Williams, 2008) and in France (Castebrunet et al., 2014). It should be noted
485 that it is still uncertain whether this expected shift of wet-avalanche season can be
486 simply applied to Japanese environment.

487

488 6 Conclusions

489 We have evaluated the response to global warming of snow depth and some 490 physical properties of snowpack at the mountain range of Mt. Annupuri in Kutchan, 491 Hokkaido, Japan (Fig. 1), by integrating a numerical snowpack model forced by DDS 492 data with multiple GCMs. First, we validated the numerical snowpack model by 493 comparing the results of the hindcast simulation with observation at Sapporo (17 m 494 above sea level) in three winters of 2009/10, 2010/11, and 2011/12: in particular, we 495 successfully reproduced snow depth at the site with bias-corrected DDS data. The 496 numerical snowpack calculation under present and future climates suggests that 497 monthly-mean snow depth will decrease by about 60 cm at the beginning of ablation 498 period if the global- and local-mean temperature increases by 2 K and approximately 499 2.5 K, respectively (Figs. 3, 6). In addition, monthly-mean snow depth reaches its peak 500 about one month earlier. The monthly-mean ratio of melt forms tends to increase at all 501 sites, especially above the hillside point, while the monthly-mean ratio of hoar is likely

502 to decrease except at the top point.

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702 Figures



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



707 Fig. 2 (a) Surface height above the sea level given to the regional atmospheric model 708 (RAM) and the location of Hokkaido in the upper right and (b) realistic topography with 709 about 1-km resolution. Outer and inner black solid line in a window of (a) shows RAM's 710 domain calculated and a domain of (a), respectively. The locations of top of Mt. Annupuri 711 (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 800 712 m level of height corresponding to the hillside of the mountain slope. Red rectangle in (b) 713 714 shows the RAM's grid cell of which meteorological data are imposed to the SNOWPACK 715 model. The color-scale is shown between the panels. 716



717

Fig. 3 Global warming response at Kutchan for November to May, based on the 718 719 dynamical downscaling (DDS) results from (red) the high-resolution version of the Model for Interdisciplinary Research on Climate 3.2 (MIROC), 720 (blue) the 721 fifth-generation atmospheric general-circulation of model (GCM) the 722 Max-Planck-Institut für Meteorologie (ECHAM5/MPI), and (green) version 3 of the 723 Community Climate System Model of the National Center for Atmospheric Research 724 (CCSM3/NCAR). The bar graph shows the increasing rate of monthly precipitation [%; 725 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; 726

scale on the right].









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

750 DDS data under (solid line) present and (dotted) future climates.

751





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

