NOTES AND CORRESPONDENCE

Evaluating the Role of Snow Cover in Urban Canopy Layer on the Urban Heat Island in Sapporo, Japan with a Regional Climate Model

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Abstract

Climate monitoring in urban areas is important because climate change in densely populated areas has a strong influence on society. The rate of long-term temperature increase in high-latitude snowy urban areas is relatively large due to global warming and urban heat islands. However, the influence of snow cover on urban heat islands is unclear. The purpose of this study is to assess the effect of snow cover in urban canopy layer on winter heat islands using a mesoscale atmospheric model coupled with an urban canopy model. Numerical experiments indicate that snow cover in urban areas serves to decrease surface air temperature, with a stronger decrease in daily maximum temperatures (0.4 – 0.6°C) than daily minimum temperatures (0.1 – 0.3°C). The increase in surface albedo is primarily responsible for the decrease in net shortwave radiation and sensible heat flux. In addition, increased evaporation causes a weakened sensible heat flux. The estimated snow cover effect during the day is comparable with the typical magnitude of anthropogenic heat release. In urban canopy layer, snow cover on roofs plays a significant role in reducing surface air temperature. Snow clearing on roads tends to increase nocturnal surface air temperature, especially in suburban areas because decreased snow depth increases ground heat transfer. These results indicate that snow cover in urban canopy layer reduces surface air temperature, resulting in weakened urban heat islands.

Keywords urban climate; urban heat island; snowy urban; regional climate model; urban canopy model

1. Introduction

Urban development replaces natural vegetation cover with impervious materials such as concrete and asphalt, leading to changes in the radiation budget and hydrological cycle. As a result, a local climate, known as the urban heat island, forms in urban areas in which the local air temperature is higher than that of the surrounding rural area (e.g., Oke 1987; Gartland 2008; Synnefa et al. 2008; McCarthy et al. 2010; Menon et al. 2010; Jacobson and Ten Hoeve 2012). Urban heat island intensity (UHI) or the urban–rural temperature difference depends on the scale of urban
Fig. 1. (a) Domains of the numerical experiment. (b) Land use classification around Sapporo in the inner model domain. HB, LBHD, and LB represent high-rise building areas, low-rise building areas (high-density), and low-rise building areas, respectively. Black broken lines show the Sapporo urban center, and the dotted line represents the city boundary. Black (bold) lines show the 100 m (300 m) elevation contours. Squares and circles mark the center of urban and suburban grid points in Fig. 6, respectively. (c) Diurnal variations in the anthropogenic heat ratio. Blue and red bars show energy consumed by electricity and traffic use, respectively.
areas (population and energy consumption) and urban morphology (Oke 1987; Fujibe 2011). UHI has diurnal and seasonal cycles, with the nocturnal UHI being typically higher than the daytime UHI (e.g., Oke 1987; Gartland 2008; Japan Meteorological Agency (JMA) 2011). The highest seasonal UHI depends on the size of the urban area and the climate zone (Arnfield 2003) but tends to occur in winter in Japan and elsewhere (JMA 2011).

JMA (2011) reported that the long-term increase in surface air temperature in urbanized areas in Japan is larger than that in rural areas. It also indicated that the rate of this increase is higher in winter than in summer and that the change in the daily minimum temperature is greater than that in the daily maximum temperature. In the city of Sapporo (141.33°E, 43.05°N) in northern Japan (Fig. 1a), the daily minimum winter air temperature has increased over 0.06°C yr\(^{-1}\) from 1931 to 2010, which is more than or comparable with other major Japanese cities (e.g., 0.069°C yr\(^{-1}\) in Tokyo, 0.036°C yr\(^{-1}\) in Osaka, and 0.043°C yr\(^{-1}\) in Nagoya) (JMA 2011). The properties of the winter heat island in high-latitude cold regions differ from those of the summer heat island effect because the lower solar radiation flux in winter enhances the contribution of anthropogenic warming to the heat island effect (Hinkel et al. 2003; Hamilton et al. 2009; Malevich and Klink 2011; Bohnenstengel et al. 2014). Snow cover also alters the surface energy budget (DeWalle and Rango 2011), and it is therefore critical to consider the surface energy budget over snow in any evaluation of the winter heat island effect in cold regions.

Relationships between snow and urban heat islands have been explored in previous studies (Malevich and Klink 2011; Schatz and Kucharik 2014), and the following key results have been reported. 1) The effect of snow is smaller in rural areas than in urban areas because of dirty snow and relatively large snow-free areas such as walls and bare areas formed by snow melt in urban areas. 2) Snow cover increases surface albedo and contributes to an increase in the daytime UHI. 3) Snow depth affects the nighttime UHI, but this effect varies with snow depth owing to ground heat transfer.

Previous studies have shown that urban snow decreases surface air temperature (Barker et al. 1992; Malevich and Klink 2011). It is also well known that the higher albedo of urban surfaces results in reduced surface air temperature (Synnefa et al. 2008; McCarthy et al. 2010; Menon et al. 2010; Olsen et al. 2010; Jacobson and Ten Hoeve 2012). However, few studies have quantitatively evaluated the effects of snow at an urban scale. The purpose of this study is to quantify the effects of snow cover in urban canopy layer on urban heat islands using the urban canopy model (UCM) coupled with a regional climate model.

2. Methodology

Experiments were conducted using a regional climate model; the Weather Research and Forecasting (WRF) (version 3.2.1) with the Advanced Research WRF (ARW) dynamical core (Skamarock et al. 2008) and a two-way nesting method. The domain (Fig. 1a) was composed of an outer domain (35 × 33 grid points at a 25 km interval), a middle domain (36 × 36 grid points at a 5 km interval), and an inner domain (66 × 66 grid points at a 1-km interval). Only the inner domain was used in the analysis. The time integration was from 0000 UTC on 1 December 2008 to 2 February 2009 for the outer domain, from 0000 UTC on 17 December 2008 to 2 February 2009 for the middle domain, and from 0000 UTC on 24 December 2008 to 2 February 2009 for the inner domain.

The WRF has 32 levels using sigma coordinates. The upper boundary of the model was set to 50 hPa, and the lower boundary was set to 26 m above ground level. The following physical schemes were applied: the Mellor–Yamada Nakanishi–Niino level 2.5 boundary layer scheme (Nakanishi and Niino 2004), Nakanishi–Niino PBL surface layer scheme (Nakanishi and Niino 2004), Grell 3D ensemble cumulus scheme (Grell and Devenyi 2002), WRF single-moment 6-class microphysical scheme (Hong and Lim 2006), rapid radiative transfer model for longwave parameterization (Mlawer et al. 1997), Dudhia scheme for shortwave radiation (Dudhia 1989), and Noah land surface model (LSM) (Chen and Dudhia 2001).

The same physical schemes were used in all domains except that the cumulus scheme was turned off in the inner domain. Initial and lateral boundary conditions were defined by 6-h JRA-25 reanalysis data (1.125° mesh grid point) (Onogi et al. 2007). A daily optimally interpolated sea surface temperature (Reynolds et al. 2007) with 1.125° horizontal grid spacing was employed for the ocean surface boundary condition. The topography in the model was obtained from the 30-s global land cover characterization dataset of the United States geological survey (USGS) (Fig. 1b). The USGS land use categories around Sapporo were replaced with 2009 land utilization segmented mesh data (with 100 m horizontal grid spacing) from the National Land Numerical Infor-
This dataset contains 16 land use categories including 3 urban categories. The UCM used in the WRF model calculates the surface temperature and surface energy budget of roofs, walls, and roads (Kusaka et al. 2001; Kusaka and Kimura 2004). We conducted two types of experiments, CTL and NO_SNOW_UCM, to investigate the effect of snow cover in urban canopy layer on surface air temperature. Physical parameters in the experiments are summarized in Table 1. Surface albedo, emissivity, thermal conductivity/snow depth on roofs and roads, and roof roughness length in the UCM of the CTL run varied according to snow cover fraction, snow depth, and snow water equivalent computed at each grid point using the Noah LSM scheme (Chen and Dudhia 2001). Detailed calculations for physical parameters in the Noah LSM are described in Appendix.

To account for snow clearing on roads, the maximum albedo, snow depth, and snow water equivalent values and minimum thermal conductivity values were prescribed for a snow depth of 10 cm, or the depth at which snow clearing is initiated in Sapporo. Minimum moisture availability for roads and roofs in the CTL run was 0.7 when snow depth was >1 cm. Otherwise, it was set to 0.7 and 0.0 for rainfall and no-rainfall conditions, respectively (the default UCM setting). In the CTL run, if the snow cover fraction was >0.01 and surface temperatures on roofs or roads in the UCM were > 0°C, surface temperature for the corresponding facet is given by

$$T_{sfc} = (1 - S_{cv}) T_{sfc} + S_{cv} T_{freezing}. $$

where $T_{sfc}$, $T_{freezing}$, and $S_{cv}$ are surface temperature ($^\circ$C), temperature at the freezing point (i.e., 0°C), and snow cover fraction, respectively. The NO_SNOW_UCM run was conducted with the same settings as the default UCM and represents a snow-free urban environment. The same settings were used in both experiments in addition to those in the UCM.

For the land-utilization segmented mesh data, we used three types of urban category: i) high-rise building areas (HB) including industrial areas; ii) high-density low-rise building areas (LBHD); and iii) low-rise building areas (LB). The urban grid in the WRF model is composed of impervious surfaces (urban areas) and pervious surfaces (natural areas). The prescribed urban area ratios (or the ratio of impervious to pervious surface areas in each urban grid) for the HB, LBHD, and LB urban types were set to 0.95, 0.9, and 0.8, respectively. Surface skin

<table>
<thead>
<tr>
<th></th>
<th>Albedo ($\alpha$)</th>
<th>Emissivity ($\epsilon$)</th>
<th>Thermal conductivity (J m$^{-1}$ s$^{-1}$ K$^{-1}$) ($\lambda$)</th>
<th>Roughness length (m) (HB, LBHD, LB) ($z_0$)</th>
<th>Snow depth (cm) ($d_{snow}$)</th>
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<tbody>
<tr>
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<td>$0.9 \leq \epsilon \leq 0.95$</td>
<td>$\lambda \leq 0.67$</td>
<td>$z_0 \leq 0.22, z_0 \leq 0.126, z_0 \leq 0.016$</td>
<td>$0 \leq d_{snow}$</td>
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<td>N-A</td>
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<tr>
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<td>$\epsilon = 0.95$</td>
<td>$\lambda = 0.67$</td>
<td>$z_0 = 0, z_0 = 0.126, z_0 = 0.016$</td>
<td>$d_{snow} = 0$</td>
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<tr>
<td>Road</td>
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<td>$\epsilon = 0.95$</td>
<td>$\lambda = 0.4004$</td>
<td>N-A</td>
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<tr>
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<td>$0.9 \leq \epsilon \leq 0.95$</td>
<td>$\lambda \leq 0.67$</td>
<td>$z_0 \leq 0.22, z_0 \leq 0.126, z_0 \leq 0.016$</td>
<td>$0 \leq d_{snow}$</td>
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<tr>
<td>Road</td>
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<td><strong>CL20</strong></td>
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<tr>
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<td>$0.9 \leq \epsilon \leq 0.95$</td>
<td>$\lambda \leq 0.67$</td>
<td>$z_0 \leq 0.22, z_0 \leq 0.126, z_0 \leq 0.016$</td>
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<td>$0 \leq d_{snow}$</td>
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<td>Road</td>
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<td>$\epsilon = 0.95$</td>
<td>$\lambda \leq 0.4004$</td>
<td>N-A</td>
<td>$0 \leq d_{snow}$</td>
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October 2015 K. MORI and T. SATO

Anthropogenic heat in the UCM was estimated using traffic and electricity consumption data. The diurnal variation in anthropogenic heat from traffic was computed on the basis of the FY2010 road traffic census results of the general traffic volume surveys (Ministry of Land, Infrastructure, Transport and Tourism, in Japanese, available at http://www.mlit.go.jp/road/census/h22-1/) and gasoline sales data in Hokkaido (Hokkaido Bureau of Economy, Trade and Industry, in Japanese, available at http://www.pref.hokkaido.lg.jp/kz/kke/sekiyu/sekiyudoukou.htm). The diurnal variation in electrical energy consumption was estimated using electricity consumption data from Hokkaido (Hokkaido Electric Power Company, in Japanese, available at http://denkiyoho.hepco.co.jp/forecast.html) and Sapporo city (in Japanese, available at http://www.city.sapporo.jp/toukei/tokeisyo/documents/h2113.pdf). Electricity consumption data are expected to represent building energy consumption including gas and kerosene with the following two assumptions: 1) the primary contribution of gas and kerosene consumption to anthropogenic heat release is space heating, and 2) the consumption of gas and kerosene for space heating has a similar diurnal cycle to that of electricity consumption. Anthropogenic heat release in the UCM at each time step was calculated by multiplying anthropogenic heat intensity (Fig. 1c) by background anthropogenic heat defined for each urban type (40, 20, and 10 W m\(^{-2}\) for HB, LBHD, and LB, respectively). Background anthropogenic heat was based on Kayaba et al. (2010) by considering the ratio of winter (January) to summer (August) energy consumption estimated from electricity and traffic data.

UCM parameters in both experiments are identical to the default values in the original UCM (Chen et al. 2011) (shown in Supplementary Table 1) because of no available data, aside from surface physical parameters in the CTL run, anthropogenic heat, and the urban area ratio.

3. Results

Figure 2 shows the horizontal distribution of monthly mean surface air temperature in January 2009 from Sapporo Information Network (SNET) hourly surface air temperature observations at 46 stations (http://www.sapporotenki.jp). The UHI between the representative urban observation station (star in Fig. 2; 43.06°N, 141.37°E) and rural observation station (circle in Fig. 2; 43.06°N, 141.55°E) is 2.9°C, 4.0°C, and 1.4°C for the daily mean temperature (T\(_{\text{mean}}\)), daily minimum temperature (T\(_{\text{min}}\)), and daily maximum temperature (T\(_{\text{max}}\)), respectively. In the CTL run, the UHI at corresponding urban and rural grid points are 2.2°C (T\(_{\text{mean}}\)), 5.3°C (T\(_{\text{min}}\)), and 0.9°C (T\(_{\text{max}}\)). Surface air temperature for T\(_{\text{mean}}\), T\(_{\text{min}}\), and T\(_{\text{max}}\) from observations and the WRF output are summarized in Supplementary Table 2. In both observations and the CTL run, the UHI is larger for T\(_{\text{min}}\) than for T\(_{\text{max}}\), which is a common characteristic of urban heat islands (e.g., Gartland 2008).

In general, the diurnal variations in UHI and the spatial distribution of air temperature agree well with observed values. The root mean square errors between observations and the model output are 1.1°C (T\(_{\text{mean}}\)), 2.2°C (T\(_{\text{min}}\)), and 0.9°C (T\(_{\text{max}}\)). The Spearman rank correlation coefficients between observations and model outputs are 0.69 (T\(_{\text{mean}}\)), 0.64 (T\(_{\text{min}}\)), and 0.53 (T\(_{\text{max}}\)), which are significant at the 99 % confidence level (\(p < 0.01\)). The WRF overestimates the observed T\(_{\text{mean}}\) and T\(_{\text{min}}\) by < 0.5°C in the central urban zone (broken lines in Fig. 2) and underestimates T\(_{\text{max}}\) by 0.5–1.0°C. The latter result is likely because the model underestimates surface downward shortwave radiation; this was 10–40 W m\(^{-2}\) lower during 0700–1700 Japan Standard Time (JST = UTC + 0900) than observations at the JMA Sapporo observatory. Weaker shortwave radiation and the associated cold bias in T\(_{\text{max}}\) is likely to be caused by more cloud cover in WRF, which is consistent with the findings of Ishizaki et al. (2012) who reported that the WRF model has a cold bias due to overestimation of precipitation in the cold season in northern Japan.

Figure 3 shows the monthly mean surface air temperature difference between the CTL and NO_SNOW_UCM runs. All T\(_{\text{mean}}\), T\(_{\text{min}}\), and T\(_{\text{max}}\) differences between the CTL and NO_SNOW_UCM runs are negative in urban areas, which illustrates the effect of snow cover in terms of reducing surface air temperature in urban areas. The temperature differences are 0.2–0.3°C (T\(_{\text{mean}}\)), 0.1–0.3°C (T\(_{\text{min}}\)), and 0.4–0.6°C (T\(_{\text{max}}\)), indicating that the snow cover effect is stronger for T\(_{\text{max}}\) than for T\(_{\text{min}}\). Furthermore, T\(_{\text{min}}\) and T\(_{\text{max}}\) in the area to the east of the urban center (black broken lines in Fig. 3a) are 0.1°C higher than that of the urban center itself (Fig. 3b, c), which is associated with changes in surface wind patterns (not shown). The temperature difference between the two runs in urban areas was > 1.5°C at 1200
Fig. 2. Monthly mean of daily mean, minimum, and maximum surface air temperature in January 2009 derived from observations (panels a, c, and e) and the CTL run (panels b, d, and f). Squares indicate the location of the observational station. Stars indicate the location of the representative urban observation station (43.06°N, 141.37°E). Circles indicate the location of the representative rural observation station (43.06°N, 141.55°E).
JST, which is around the time when $T_{\text{max}}$ occurs on a typical cloud-free day (not shown).

Although the snow cover effect reduces urban temperatures, downward radiation (both longwave and shortwave) is similar in the CTL and NO_SNOW_UCM runs, suggesting that the effect of snow cover on cloud cover is negligible. Therefore, surface cooling is likely to be caused by changes in surface radiation and the energy balance associated with snow cover.

Figure 4 shows the grid scale surface energy budget in the two runs. Snow cover has a pronounced effect on net shortwave radiation and a minor effect on net longwave radiation. At 1200 JST, net shortwave radiation in the CTL run is 25 % lower than in the NO_SNOW_UCM run (a decrease of 50 W m$^{-2}$) because of increased albedo. Weak net shortwave radiation moderates ground surface heating, and thus the sensible and ground heat flux at the surface become smaller. The daytime latent heat flux increases in the CTL run owing to an increase in minimum moisture availability. At 1200 JST, the latent heat flux in the CTL run is 58 % larger than in the NO_SNOW_UCM run (a 16.2 W m$^{-2}$ increase). An increased latent heat flux leads to a decrease in the sensible and ground heat fluxes at the surface. These differences contribute to cooling of surface air temperature by reducing the sensible heat flux and heat storage in buildings and roads. A large decrease in net shortwave radiation plays a particularly important role in determining surface air temperature change. As a result, the consideration of snow covered situation in the urban surface scheme makes the estimation of UHI lower.

The diurnal variation in the sensible heat flux shown in Fig. 4 is consistent with the result that the snow cover effect is larger for $T_{\text{max}}$ than for $T_{\text{min}}$. Figure 4 indicates that the difference in the sensible heat flux between the NO_SNOW_UCM and CTL runs is larger in the daytime than overnight. This result suggests that the effect of snow cover on the ground heat flux in buildings and roads has less impact on surface air temperature decrease than on the sensible heat flux, because the ground heat flux change is likely related to the nocturnal sensible heat flux change and subsequent surface air temperature change. Since the simulated sensible heat flux at each time step is calculated as the total of sensible heat fluxes from roofs, walls, roads, and anthropogenic heat, the contributions of all heat flux components influenced by the snow cover effect were evaluated.

To elucidate the influence of the snow cover effect on the sensible heat flux, Fig. 5 shows the monthly

<table>
<thead>
<tr>
<th>Variable</th>
<th>Symbol</th>
<th>Unit</th>
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<tbody>
<tr>
<td>Snow depth</td>
<td>$d_{\text{snow}}$</td>
<td>(m)</td>
</tr>
<tr>
<td>Thickness of background upper soil layer</td>
<td>$d_{s1}$</td>
<td>(m)</td>
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<tr>
<td>Snow water equivalent</td>
<td>$SWE$</td>
<td>(m)</td>
</tr>
<tr>
<td>Snow density</td>
<td>$\rho_s$</td>
<td>(g cm$^{-3}$)</td>
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<tr>
<td>Snow cover fraction</td>
<td>$S_{cv}$</td>
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<tr>
<td>Elapsed time after prior snowfall</td>
<td>SNOWTIME</td>
<td>(s)</td>
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<tr>
<td>Snow surface albedo</td>
<td>$\alpha_{\text{snow}}$</td>
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<tr>
<td>Maximum snow surface albedo</td>
<td>$\alpha_{\text{Max_snow}}$</td>
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<tr>
<td>Background surface albedo</td>
<td>$\alpha_0$</td>
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<tr>
<td>Thermal conductivity of snow</td>
<td>$\lambda_{\text{snow}}$</td>
<td>(J m$^{-1}$ s$^{-1}$ K$^{-1}$)</td>
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<tr>
<td>Thermal conductivity of background upper soil layer</td>
<td>$\lambda_{s1}$</td>
<td>(J m$^{-1}$ s$^{-1}$ K$^{-1}$)</td>
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<tr>
<td>Snow surface emissivity</td>
<td>$\varepsilon_{\text{snow}}$</td>
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<tr>
<td>Background surface emissivity</td>
<td>$\varepsilon_0$</td>
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<tr>
<td>Roughness length above snow</td>
<td>$z_0_s$</td>
<td>(m)</td>
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<tr>
<td>Roughness length above background surface</td>
<td>$z_0_0$</td>
<td>(m)</td>
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<tr>
<td>Effective roughness length</td>
<td>$z_{0\text{eff}}$</td>
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mean diurnal variation in the sensible heat flux from the UCM (excluding anthropogenic heat). The total sensible heat flux in the CTL run is smaller than in the NO_SNOW_UCM run throughout the day. The total sensible heat fluxes at 1200 JST in the NO_SNOW_UCM and CTL runs are 83.3 and 32.7 W m$^{-2}$, respectively. The total sensible heat flux in the CTL run decreases to 39 % of the NO_SNOW_UCM run (a 50.6 W m$^{-2}$ decrease), indicating that the net loss of sensible heat from snow cover due to increased upward shortwave radiation is comparable with or larger than the amount of anthropogenic heat released from urban areas. The change in the sensible heat flux from roofs, walls, and roads is $-43.1$, $3.7$, and $-11.2$ W m$^{-2}$, respectively. Thus, in snow-covered areas, roofs account for 85 % of the total sensible heat flux decrease.

The total sensible heat fluxes at 0600 JST in the NO_SNOW_UCM and CTL runs are $-18.8$ and $-21.4$ W m$^{-2}$, respectively (i.e., the sensible heat flux from the atmosphere to the surface increases by 2.6 W m$^{-2}$ in the CTL run). The changes in the sensible heat flux from roofs, walls, and roads due to snow cover are $-1.3$, $-0.2$, and $-1.1$ W m$^{-2}$, respectively. The influence of urban snow cover on the sensible heat flux is smaller overnight or in the early morning than in the daytime. This result suggests that changes in daytime heat fluxes affect the sensible heat flux decrease overnight. As a result, the $T_{\text{max}}$ of urban areas decreases in the CTL run by 0.4–0.6°C, whereas the $T_{\text{min}}$ of...
urban areas decreases by 0.1–0.3°C. In the NO_SNOW_UCM run, the sensible heat flux increase in the morning (0700–1200 JST) and the decrease in the early afternoon (1200–1700 JST) are largest over roofs, followed by roads and walls.

These results show that snow cover in urban areas serves to reduce the sensible heat flux as a result of reduced daytime net shortwave radiation at the surface. This decrease in net shortwave radiation is caused mainly by the higher albedo over roofs. However, the higher albedo over road surfaces increases the absorption of daytime net shortwave radiation on walls, leading to rising wall temperatures and an increased sensible heat flux on wall surfaces.

4. Discussion and conclusions

To elucidate the effect of snow cover in urban canopy layer on winter heat islands in urban environments, numerical experiments were conducted using an UCM coupled with the WRF model. Consideration of the snow cover situation in urban areas results in decreased urban surface air temperatures and weakened temperature gradients away from the urban center, especially during the daytime. Decreased urban surface air temperatures are caused primarily by a decrease in absorbed shortwave radiation and secondarily by an increase in the latent heat flux during daytime. The contribution of solar insolation to winter heat islands is less than that to summer heat islands (Oleson et al. 2010). Moreover, the contribution of anthropogenic heat in winter is larger than in other seasons because of the weaker solar radiation flux (Hinkel et al. 2003; Hamilton et al. 2009; Malevich and Klink 2011; Bohnenstengel et al. 2014). Therefore, the effect of heat storage on maintaining nocturnal urban surface air temperatures is relatively small in winter, indicating that nocturnal temperature distribution depends heavily on nocturnal anthropogenic heat release. Further improvement in prescribed diurnal variations in anthropogenic heat would lead to more realistic simulations of urban heat islands in cold regions.

Snow cover decreases the monthly mean sensible heat flux by 11 W m$^{-2}$, which is larger than the 8.8 W m$^{-2}$ of monthly mean anthropogenic heat release averaged across all urban grid points. Moreover, the most significant decrease in the sensible heat flux is 50.6 W m$^{-2}$ at 1200 JST, which is 1.6 times larger than the anthropogenic heat release in HB (31.2 W m$^{-2}$) at 1200 JST. This result indicates that snow cover weakens the daytime urban temperature gradient (i.e., heat island intensity) more than anthropogenic heat strengthens the gradient. Therefore, snow cover in urban areas is of comparable importance to anthropogenic heat release when simulating the spatial air temperature contrast between urban and surrounding areas.

Schatz and Kucharik (2014) reported snow depth effects on nocturnal surface air temperature through ground heat transfer change. To confirm the effect of snow depth and road snow clearing on urban surface air temperature, we conducted sensitivity experiments as the CL0, CL20, and NO_CL runs (Table 1). These runs used the same settings as the CTL run except for maximum snow depth on the road in the UCM (i.e., to represent different snow clearing regulations). Maximum snow depths on the road in the CL0 and CL20 runs were 0 and 20 cm, respectively, whereas in the NO_CL run there was no road snow clearing.

These experimental results show that road snow clearing has a stronger effect on $T_{\text{min}}$ than on $T_{\text{max}}$ (not shown), which is consistent with the results of Schatz and Kucharik (2014). This result also indicates that the decrease in $T_{\text{max}}$ is attributable mainly to snow cover on roofs. Figure 6 shows the difference in monthly mean daily $T_{\text{min}}$ between experiments with and without road snow clearing. The difference in $T_{\text{min}}$ averaged across suburban grid points (circles in Fig. 1b) ranges from 0.1°C to 0.24°C, whereas the difference for average values centered on urban grid
Matsumura and Sato (2011) projected a snow depth decrease in northern Japan due to global warming. Therefore, a detailed investigation on the influence of reduced snow cover in urban areas is necessary because it can enhance the future warming in snowy urban areas through positive feedbacks. It will be therefore necessary to develop the model further to represent snow accumulation processes in urban areas.

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Supplementary Materials

Supplementary Table 1 lists the physical properties of building surfaces and walls in the UCM. Supplementary Table 2 provides a summary of surface air temperature and UHI for $T_{\text{mean}}$, $T_{\text{min}}$, and $T_{\text{max}}$ for both observations and the WRF output.

Appendix

This appendix describes methods for calculating the physical parameters of snow-covered surfaces (surface albedo, surface emissivity, thermal conductivity, and roughness length) in the Noah LSM (Chen and Dudhia 2001), which is also used to compute parameters for roofs and roads in the UCM. The Noah LSM employs a single-layer snowpack model in which snow depth $d_{\text{snow}}$ and snow water equivalent $\text{SWE}$ are defined for a single snow column composed of old and new snow layers. Snow density $\rho_s$ is computed as $\text{SWE}$ divided by $d_{\text{snow}}$, where the temporal change in $d_{\text{snow}}$ is expressed by considering the accumulation of fresh snow, melting, and loading. Table 2 lists the parameters in the snowpack model. The snow cover fraction $S_{cv}$ ranges from 0 to 1 and is estimated as follows:

$$\begin{align*}
S_{cv} &= \begin{cases} 
1.0 - \exp(-\frac{2.6\text{SWE}}{0.04}) + \frac{\text{SWE}}{0.04} \exp(-2.6) & (\text{SWE} \leq 0.04) \\
1 & (\text{SWE} > 0.04)
\end{cases}
\end{align*}$$

Snow-covered surface physical parameters (surface albedo $\alpha$, surface emissivity $\varepsilon$, thermal conductivity...
\( \lambda \) and roughness length \( z_0 \) are expressed as weighted mean values for snow and background surface by \( S_{sv} \). Snow covered surface albedo \( \alpha \) is given by

\[
\alpha = (1 - S_{sv})\alpha_0 + S_{sv}\alpha_{snow}, \quad \text{and}
\]

\[
\alpha_{snow} = \alpha_{Max\_snow} \times 0.94^{(\text{SNOTIME/86400})^{0.58}}.
\]

Snow-covered surface emissivity \( \varepsilon \) is expressed as

\[
\varepsilon = (1 - S_{sv})\varepsilon_0 + S_{sv}\varepsilon_{snow}.
\]

The thermal conductivity of a snow-covered surface \( \lambda \) is computed by

\[
\lambda = (1 - S_{sv})\lambda_0 + S_{sv}\frac{\lambda_0 + \lambda_{snow}}{d_0 + d_{snow}}, \quad \text{and}
\]

\[
\lambda_{snow} = 0.07623 \times 10^{2.25 u_s}.
\]

Roughness length \( z_0 \) for a snow-covered surface is given by

\[
z_0 = (1 - S_{sv})z_0 + S_{sv}z_{0\_eff},
\]

\[
z_{0\_eff} = \begin{cases} z_0, & (7.0 \leq z_0 - d_{snow} \leq 0.0007) \\ (7.0 \leq z_0 - d_{snow} < 0.0007) & (7.0 \leq z_0 - d_{snow} > 0.0007) \end{cases}
\]

References


McCarthy, M. P., M. J. Best, and R. A. Betts, 2010:


