Thermal Effects of Urban Canyon Structure on the Nocturnal Heat Island: Numerical Experiment Using a Mesoscale Model Coupled with an Urban Canopy Model

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ABSTRACT
A single-layer urban canopy model is incorporated into a simple two-dimensional atmospheric model in order to examine the individual impacts of anthropogenic heating, a large heat capacity, and a small sky-view factor on mesoscale heat island formation. It is confirmed that a nocturnal heat island on a clear, calm summer day results from the difference in atmospheric stability between a city and its surroundings. The difference is caused by anthropogenic heating and the following two effects of urban canyon structure: (i) a larger heat capacity due to the walls and (ii) a smaller sky-view factor. Sensitivity experiments show that the anthropogenic heating increases the surface air temperature though the day. (This factor strongly affects the nocturnal temperature, and the maximum increase of 0.67 °C occurs at 0500 LST.) The larger heat capacity due to the walls decreases the daytime temperature and increases the nocturnal temperature. (The maximum increase of 0.39 °C occurs at 0600 LST.) The smaller sky-view factor increases the temperature though the day, particularly during the first several hours after sunset. (The maximum increase of 0.52 °C occurs at midnight.) In urban areas, this factor results in uniform cooling that occurs at a constant rate. The impact of the canyon structure is shown to be as significant as anthropogenic heating.

1. Introduction
The surface layer in cities is generally warmer than that of the surrounding areas. Near a city, the surface isotherms look like the topographic contours around an island. Thus, this phenomenon has become known as the urban heat island, which is clearly observed under atmospheric conditions of a clear sky and light wind. Many observational studies have provided us with essential ideas about the urban heat island phenomenon. These earlier studies have been summarized by Yoshino (1975), Landsberg (1981), and Oke (1987). Numerical simulations have also been conducted using boundary layer and mesoscale models in order to investigate the mechanism of the urban heat island formation (e.g., Atwater 1972; Kimura and Takahashi 1991; Avissar 1996; Taha 1999; Kusaka et al. 2000). To better understand the role of the various factors, multilayer urban canopy models have also been developed. The approach is typically used to apply the effective fluid volume in order to consider the volume occupied by obstacles and to add the drag-force term in order to account for the drag force induced by the obstacles. Uno et al. (1989) added a drag-force term to the atmospheric boundary layer models. Ca et al. (2002), Martilli et al. (2002), and Kondo et al. (2004) developed multilayer urban canopy models, considering thermal effects such as shading and the radiation trapping by the buildings, as well as the mechanical effects. Their models are very powerful and are suitable for studying the temperature and wind profiles within the urban canopy layer or boundary layer. However they require higher vertical resolution than that of current mesoscale models, such as the fifth-generation Pennsylvania State University–National Center for Atmospheric Research (NCAR) Mesoscale Model (MM5), and next-generation models such as the Weather Research and Forecasting Model developed by the collaboration among NCAR, the National Centers for Environmental Prediction, the National Oceanic and Atmospheric Administration Forecast Systems Laboratory, and the U.S. Air Force Weather Agency, for example (Dudhia 1993; Grell et al. 1994; NCAR 2002). The reason for this is that a mesoscale model has to set its first atmospheric level greater than 10 times that of a roughness length. Single-layer urban canopy models are proposed by Masson (2000), Kusaka et al. (2001), and

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Kusaka and Kimura (2004). Their models can be easily applied to mesoscale models without modifying the dynamic core, which is required by the developers of mesoscale models and subsequent users. Kusaka et al. (2001) performed a validation of simulated surface temperatures and net radiation against observed data. Furthermore, they confirmed that the performance of their model is approximately that of multilayer models for the surface fluxes, which are important for the study of mesoscale urban heat islands. Kusaka and Kimura (2004) showed that their coupled mesoscale model with their canopy model could produce better results than a standard mesoscale model. These results indicate the necessity of an urban-specific surface exchange for mesoscale models to better understand the mechanism of mesoscale heat island formation.

When studying heat island formation, a clear, calm winter day—the most pronounced weather condition—has been traditionally focused on. One of the current interests is to examine important factors of the nocturnal mesoscale heat island formation on a clear, calm summer day, which is recognized in relation to environmental problems and to cooling needs associated with uncomfortable hot and humid conditions (e.g., Kimura and Takahashi 1991; Ichinose et al. 1999). Kimura and Takahashi (1991) conducted sensitivity experiments to investigate the heat island of Tokyo, Japan, under clear, calm summertime synoptic conditions using a mesoscale model with a standard slab urban model; one simulation included artificial land cover and anthropogenic heat, another simulation ignored anthropogenic heat, and another had neither. A comparison of these simulations established that the daytime heat island of Tokyo is mainly formed by the relatively small latent heat release from the artificial land surface, and the nocturnal heat island is caused by anthropogenic heat release. Ichinose et al. (1999) also came to the same conclusion for the heat island of Tokyo using a mesoscale model with very a detailed anthropogenic heat map. However, their numerical experiments do not discuss the thermal effects of the urban canyon structure suggested by the earlier laboratory and field experiments (e.g., Oke 1981; Sugawara et al. 2001) and the recent numerical experiments using the multilayer urban canopy models (Martilli et al. 2002). In the present study, we will individually examine the thermal effects of the small sky-view factor and large heat capacity on a mesoscale nocturnal heat island formation under clear, calm summertime synoptic conditions, using a mesoscale model coupled with a single-layer urban canopy model. Furthermore, these factors will be compared with another thermal factor—anthropogenic heating. This is required for aid in urban planning to solve the environmental and energy problems, to progress urban modeling, and to contribute to meteorological interests. It is noteworthy that we focus on the impacts on the mesoscale heat island over a metropolitan area like Tokyo and its surroundings, not on a microscale to local-scale thermal environment in a city. A brief description of the mesoscale model and single-layer urban canopy model is given in section 2. Conditions and the results of simulations are shown in sections 3 and 4, respectively. Conclusions and remarks follow in sections 5.

2. Model description

a. Dynamic core

The numerical model used in the present study is based on the two-dimensional, hydrostatic mesoscale model (e.g., Kimura and Kuwagata 1993; Lee and Kimura 2001), but the anelastic approximation is applied to the dynamic core and incorporates the single-layer urban canopy model (Kusaka and Kimura 2004). The anelastic equations of motion, based on the hydrostatic assumption by Ogura and Phillips (1962), are

\[
\frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} + w \frac{\partial u}{\partial z} = f v - \frac{\partial \pi'}{\partial x} + \frac{\partial}{\partial x} \left( K_H \frac{\partial u}{\partial x} \right) + \frac{\partial}{\partial z} \left( K_H \frac{\partial w}{\partial z} \right) - \frac{\partial}{\partial x} \left( K_m \frac{\partial v}{\partial x} \right) + \frac{\partial}{\partial z} \left( K_m \frac{\partial w}{\partial z} \right)
\]

(1)

\[
\frac{\partial v}{\partial t} + u \frac{\partial v}{\partial x} + w \frac{\partial v}{\partial z} = -f (u - u_0) + \frac{\partial}{\partial x} \left( K_H \frac{\partial v}{\partial x} \right) + \frac{\partial}{\partial z} \left( K_m \frac{\partial w}{\partial z} \right) + \frac{\partial}{\partial x} \left( K_m \frac{\partial v}{\partial x} \right) + \frac{\partial}{\partial z} \left( K_m \frac{\partial w}{\partial z} \right) + \frac{\partial}{\partial x} \left( K_m \frac{\partial w}{\partial x} \right) + \frac{\partial}{\partial z} \left( K_m \frac{\partial w}{\partial z} \right) + \frac{\partial}{\partial x} \left( K_m \frac{\partial w}{\partial x} \right) + \frac{\partial}{\partial z} \left( K_m \frac{\partial w}{\partial z} \right)
\]

(2)

The hydrostatic equation is

\[
\frac{\partial \pi'}{\partial z} = g \frac{\theta'}{\Theta^2}
\]

(3)

and the continuity equation is

\[
\frac{\partial \rho_{lu}}{\partial x} + \frac{\partial \rho_{lw}}{\partial z} = 0
\]

(4)

Here, the symbols \( u, v, w, x, z, f, g, \) and \( \theta \) have their conventional meaning; \( u_0 \) is the east–west component of geostrophic wind velocity; \( K_H \) is the horizontal exchange coefficient, and \( K_m \) is the vertical exchange coefficient of momentum; and \( \theta' \) is the perturbation potential temperature from mean potential temperature \( \Theta \). Here \( \pi \) is the Exner’s function, and \( \pi' \) is its perturbation from mean Exner’s function \( \Pi \); \( \rho_l \) is the mean fluid density.

The thermodynamic equation is expressed as

\[
\frac{\partial \theta'}{\partial t} + u \frac{\partial \theta'}{\partial x} + w \frac{\partial \theta'}{\partial z} = \frac{\partial}{\partial x} \left( K_H \frac{\partial \theta'}{\partial x} \right) + \frac{\partial}{\partial z} \left( K_m \frac{\partial \theta'}{\partial z} \right) + Q_{cool}
\]

(5)

Here, \( K_s \) is the vertical exchange coefficient of heat and
Fig. 1. Schematic of the single-layer urban canopy model: $T_a$ is the air temperature at reference height $z_a$, $T_R$ is the building roof temperature, $T_W$ is the building wall temperature, $T_G$ is the road temperature, $T_S$ is the temperature defined at $z_T + d$, $H$ is the sensible heat exchange at the reference height, $H_a$ is the sensible heat flux from the canyon space to the atmosphere, $H_W$ is that from wall to the canyon space, $H_G$ is that from road to the canyon space, and $H_R$ is that from roof to the atmosphere.

Fig. 2. Radiation of the single-layer urban canopy model; $S_a$ is the direct solar radiation incident on a horizontal surface, $l_{road}$ is the normalized road width, $h_c$ is the normalized building height ($l_{roof} = 1$), and $l_{shadow}$ is the normalized shadow length on the road.

Q_{cool} is the radiative cooling term. The equation for specific humidity $q_v$ is expressed as

$$\frac{\partial q_v}{\partial t} + u \frac{\partial q_v}{\partial x} + w \frac{\partial q_v}{\partial z} = \frac{\partial}{\partial x} \left( K_u \frac{\partial q_v}{\partial x} \right) + \frac{\partial}{\partial z} \left( K_v \frac{\partial q_v}{\partial z} \right). \quad (6)$$

Exner’s function is defined as

$$\pi = c_p \left( \frac{p}{p_\infty} \right)^{\gamma / c_p}. \quad (7)$$

Here, $p_\infty$ is the reference pressure, $R_g$ is the gas constant for air, and $c_p$ is the specific heat of air. The vertical coordinate of the model is transformed from $z$ to $z^*$, and is written in a terrain-following coordinate system in order to consider the terrain effect in the model. The vertical coordinate $z^*$ is defined as

$$z^* = z_{TOP} - \frac{z_G}{z_{TOP} - z_G}. \quad (8)$$

Here, $z_G$ and $z_{TOP}$ are ground elevation and height of the model atmosphere top, respectively.

b. PBL scheme

The vertical exchange coefficients are calculated in the turbulent closure model at level 2, developed by Mellor and Yamada (1974), which has an approximate balance between production and dissipation terms of the turbulence kinetic energy equation. It is considered that the level-2 model is applicable for the spatial resolution in this study.

c. Radiation scheme

The downward shortwave radiation is calculated according to Kimura and Arakawa (1983), which provides diurnally varying shortwave radiation and is distinct, direct solar radiation rather than a diffused one.

The downward longwave radiation $L_{\downarrow}$ is given by an empirical form (Kondo et al. 1991), estimated from Yamamoto’s radiation chart (Yamamoto 1952), and much data on clear-sky days during summer (1979–86). The form depends on atmospheric column-integrated water vapor and surface air temperature calculated in the model. Newtonian cooling for atmospheric cooling is also assumed in the lower atmosphere when the surface temperature is less than the atmospheric temperature. The cooling coefficient is assumed to be $10^{-5}$ s$^{-1}$ after Yamamoto et al. (1973).

d. Land surface models

The land surface model consists of two parts: the slab grassland model and the single-layer urban canopy model. Here, the slab grassland model is a standard surface layer scheme using the Monin–Obukhov similarity theory for bulk transfer coefficients, a heat balance equation for surface skin temperature, and a vertical diffusion equation for soil temperature. The single-layer urban canopy model (Fig. 1) includes (a) street canyons that are parameterized to represent the urban geometry, (b) shadowing from buildings and the reflection of radiation (Fig. 2), (c) the canyon orientation and diurnal change of solar azimuth angle, (d) a surface that consists of several canyons with different orientation, (e) an exponential wind profile in the canopy layer (Inoue 1963), (f) the multilayer heat equation for the roof, wall, and road interior temperatures, (g) anthropogenic heat, and (h) the very thin bucket model for hydrological processes. The moisture availability is time independent in the present study. This urban canopy model estimates both the surface temperatures of, and heat fluxes from,
the roof, wall, and road; it also calculates the energy and momentum exchange between the urban surface and the atmosphere. More details about the present canopy model are described in the appendix. Surface fluxes from artificial and natural surfaces are calculated from the slab grassland model and the urban canopy model, respectively. Afterward, total fluxes are calculated using subgrid parameterization for heterogeneous surfaces proposed by Kimura (1989).

e. Boundary conditions

The top boundary is controlled by the wave radiation condition discussed in Klemp and Durran (1983), in order to avoid the reflection of gravity waves generated in lower layers. For the lateral boundaries, periodic boundary conditions are applied. The lower boundary is given by the surface fluxes from the land surface to environmental problems and cooling needs associated with uncomfortable hot and humid conditions. On the other hand, the land cover was set up, referring to the Tokyo metropolitan area and its surroundings. However, a flat topography is assumed in order to avoid the effects of mesoscale circulation induced by terrain and the contrast of the land and sea, which overcome urban effects. The Coriolis parameter is also set to zero to simplify the simulation. Simulations using a 2D model with idealized land cover and terrain readily provide us the comprehensible results necessary for examining the “potential impact,” although it has disadvantages, such as producing strong heat island circulation and not simulating a real case. Indeed, many simulations for idealized cases have been carried out in order to focus on the heat island phenomenon under such conditions (e.g., Yoshikado 1992, 1994; Avissar 1996; Martilli 2002; Ohashi and Kida 2002a, b).

Each simulation is run within a domain that is 300 km in the zonal (x) direction and 4 km in the vertical (z). The horizontal grid spacing is 3 km and the vertical grid spacing is 20 m near the ground increasing to 400 m at the top. The urban area of 30 km is located in the middle of the domain. For each grid box in the urban area, it is assumed that 80% of the grid box is covered by artificial surface and 20% by natural surface. The artificial and natural surfaces are distributed randomly within each grid box. The surrounding rural areas are assumed to be only the natural surface. Canyon dimensions and surface parameters are set up using Kusaka et al. (2000), Sugawara (2001), and Kusaka and Kimura (2004) as references. These are almost the same as averages of the Tokyo metropolitan area (Tables 1 and 2). Anthropogenic heat is also added to the lowest grid points of the atmospheric layer over urban areas for the control run. Area-averaged daily mean anthropogenic heat is assumed to be 25 W m\(^{-2}\) using the

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### Table 1. Parameters for artificial surface.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Symbol</th>
<th>Value</th>
<th>Unit</th>
</tr>
</thead>
<tbody>
<tr>
<td>Roof level (building height)</td>
<td>(z_r)</td>
<td>6</td>
<td>(m)</td>
</tr>
<tr>
<td>Normalized building height</td>
<td>(h_1)</td>
<td>0.35</td>
<td></td>
</tr>
<tr>
<td>Normalized roof width</td>
<td>(l_{ro} )</td>
<td>0.3</td>
<td></td>
</tr>
<tr>
<td>Normalized road width</td>
<td>(l_{ro} )</td>
<td>0.7</td>
<td></td>
</tr>
<tr>
<td>Volumetric heat capacity</td>
<td>(\rho c)</td>
<td>(2.01 \times 10^3)</td>
<td>(J m(^{-2}) K(^{-1}))</td>
</tr>
<tr>
<td>Thermal conductivity</td>
<td>(\kappa)</td>
<td>2.28</td>
<td>(W m(^{-1}) K(^{-1}))</td>
</tr>
<tr>
<td>Sublayer Stanton number</td>
<td>(B_{st})</td>
<td>6</td>
<td></td>
</tr>
<tr>
<td>Roughness length above city</td>
<td>(z_{cm})</td>
<td>0.5</td>
<td>(m)</td>
</tr>
<tr>
<td>Roughness length above canyon</td>
<td>(z_{ca})</td>
<td>0.667</td>
<td>(m)</td>
</tr>
<tr>
<td>Roughness length above roof</td>
<td>(z_{ca})</td>
<td>0.005</td>
<td>(m)</td>
</tr>
<tr>
<td>Zero-plane displacement height</td>
<td>(d)</td>
<td>2</td>
<td>(m)</td>
</tr>
<tr>
<td>Roof surface albedo</td>
<td>(\alpha_r)</td>
<td>0.2</td>
<td></td>
</tr>
<tr>
<td>Wall surface albedo</td>
<td>(\alpha_w)</td>
<td>0.2</td>
<td></td>
</tr>
<tr>
<td>Road surface albedo</td>
<td>(\alpha_c)</td>
<td>0.2</td>
<td></td>
</tr>
<tr>
<td>Roof surface emissivity</td>
<td>(\epsilon_r)</td>
<td>0.97</td>
<td></td>
</tr>
<tr>
<td>Wall surface emissivity</td>
<td>(\epsilon_w)</td>
<td>0.97</td>
<td></td>
</tr>
<tr>
<td>Road surface emissivity</td>
<td>(\epsilon_c)</td>
<td>0.97</td>
<td></td>
</tr>
<tr>
<td>Moisture availability for the artificial surface</td>
<td>(\beta_r)</td>
<td>0</td>
<td></td>
</tr>
<tr>
<td>Canyon orientation</td>
<td>(\theta_{can})</td>
<td>n m/8 ((n = 0–7))</td>
<td>(rad)</td>
</tr>
</tbody>
</table>

### Table 2. Parameters for natural surface.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Symbol</th>
<th>Value</th>
<th>Unit</th>
</tr>
</thead>
<tbody>
<tr>
<td>Volumetric heat capacity</td>
<td>(\rho c)</td>
<td>(2.01 \times 10^3)</td>
<td>(J m(^{-2}) K(^{-1}))</td>
</tr>
<tr>
<td>Thermal conductivity</td>
<td>(\kappa)</td>
<td>2.28</td>
<td>(W m(^{-1}) K(^{-1}))</td>
</tr>
<tr>
<td>Sublayer Stanton number</td>
<td>(B_{st})</td>
<td>6</td>
<td></td>
</tr>
<tr>
<td>Roughness length</td>
<td>(z_{o})</td>
<td>0.1</td>
<td>(m)</td>
</tr>
<tr>
<td>Surface albedo</td>
<td>(\alpha)</td>
<td>0.2</td>
<td></td>
</tr>
<tr>
<td>Grass surface emissivity</td>
<td>(\epsilon)</td>
<td>0.97</td>
<td></td>
</tr>
<tr>
<td>Moisture availability</td>
<td>(\beta)</td>
<td>0.3</td>
<td></td>
</tr>
</tbody>
</table>
energy consumption data of the Tokyo metropolitan area in August 1994. This is comparable to Kimura and Takahashi (1991) and NIRE (1997). The diurnal cycle presented in Fig. 3 is essentially the same as that in Kimura and Takahashi (1991). However, the values at 1300, 1800, and 1900 local solar time (LST) have been slightly modified to preserve continuity. This diurnal variation is comparable to Kanda et al. (2001). The maximum anthropogenic heat is 39% greater than the mean, while the minimum is 42% less than the mean. It is considered that the simplified, idealized urban structure readily provides us with comprehensible results, as well as a simplification of the simulation.

b. Initial condition

Numerical integration of all simulations started at 0300 local solar time (LST) on the idealized summer day under the clear, calm synoptic conditions, which are considered to be pronounced for uncomfortable hot and humid conditions. Initial conditions were created from observations under the conditions (Kusaka et al. 2000). A potential temperature lapse rate of 0.005 K km\(^{-1}\) was assumed in the model atmosphere, and a relative humidity of 70% and a wind speed of 0 m s\(^{-1}\) were assumed at all levels. The model atmosphere is assumed to be horizontally homogeneous at initialization. In nature, large-scale winds and terrain-forced winds periodically advect the heat, and consequently produce the typical atmospheric condition like an initial condition assumed in the present study. In the model, however, such an effect could not be included. Thus, the model was run for 36 h. The purpose of this study is to examine the several factors of the nocturnal heat island with sensitivity experiments using a numerical model. Thus, it is desirable to set the same initial and boundary conditions in all simulations.

4. Results

a. Control run

Figure 4 shows the potential temperature above urban and rural sites. A mixed layer with a height up to 900 m is formed over the rural areas at 1500 LST. Over the urban area, a deeper mixed layer of 1200-m depth develops due to the higher surface temperature and the upward flow from the heat island circulation (Fig. 5). Surface potential temperature in the urban area is about 2 K higher than that in the rural area (Fig. 5).

b. Streamlines and potential temperature

The streamlines and potential temperature at 1500 LST are shown in Fig. 5. The streamlines indicate the flow pattern due to the heat island circulation. The potential temperature contour shows the temperature variation in the atmosphere. The maximum potential temperature is found in the upper part of the mixed layer, and the minimum is found in the lower part. This indicates that the heat island circulation is effective in transporting warm air from the urban area to the surrounding rural areas.

In summary, the numerical simulations demonstrate that the simplified, idealized urban structure readily provides comprehensible results. The diurnal variation of anthropogenic heat is comparable to that observed in the Tokyo metropolitan area. The initial condition was created from observations under the clear, calm synoptic conditions. The model atmosphere is assumed to be horizontally homogeneous at initialization. Large-scale winds and terrain-forced winds periodically advect the heat, producing the typical atmospheric condition like an initial condition assumed in the present study. The purpose of this study is to examine the several factors of the nocturnal heat island with sensitivity experiments using a numerical model.
304 K, and the heat island intensity is about 1.4 K. At 0300 LST, a stable layer develops at the urban site, but for the near-urban surface (Fig. 6). A stable layer develops because of radiative cooling at the rural site. The difference in the surface air temperature between urban and rural sites shows that the heat island intensity is about 2.4 K, which is stronger than that of the daytime heat island. This feature qualitatively agrees with the typical temperature profile observed above the central part of a metropolitan area, for instance, Tokyo (Saito 1977).

When the results of the heat budget are shown with respect to the heat island intensity, the mechanism of the heat island formation becomes clear. Figures 7 and 8 show some differences in a diurnal variation of surface energy budget between urban and rural sites. One of the major differences is that the latent heat flux is much smaller during the daytime at the urban site. This is due to the difference in the moisture availability between urban and rural sites. Persistence of the daytime heating by the warmer urban surface helps to produce and maintain the daytime heat island. Another difference is that the heat storage is much larger during the daytime at the urban site, and, thus, the sensible heat flux remains positive until midnight at the warmer urban surface. This maintains the mixed layer near the surface and the nocturnal heat island. Duration of the heating depends on the canyon geometry, urban scale, location, weather conditions, season, and human activity (e.g., Oke 1981; Kimura and Takahashi 1991; Ichinose et al. 1999; Kusaka et al. 2000; Martilli 2002). However, observations show that the large heat storage during the daytime and positive sensible heat flux during several hours after sunset are an essential features of the urban surface heat...
b. Sensitivity experiments

Figure 9 shows the diurnal variation of the surface skin temperatures of the roof, wall, and road. These are an average temperature of eight types of canyons. The wall surface skin temperature peaks between 1600 and 1700 LST, whereas the roof and road surface temperatures peak around 1400 and 1300 LST, respectively. It is noted that the cooling rate of the wall surface is much smaller than that of the roof and road surfaces. Because of the lower cooling rate, the temperature of the wall surface is high, even at 0500 LST. These indicate that the urban canyon structure plays the important role in nocturnal heat island formation, as well as anthropogenic heating. In the present study, we investigate the thermal effects of the anthropogenic heat and urban canyon structure, that is, a larger volumetric heat capacity and smaller sky-view factor due to the existence of walls.

First, we investigate the impact of anthropogenic heating on the surface layer air temperature. Anthropogenic heating increases the temperature through the day (Figs. 10 and 11). However, the heating has a greater effect on the nocturnal temperature. The results from cases 1 and 2 show that the temperature difference decreases from 0.43°C at 1200 LST to 0.24°C at 1900 LST. Table 3 defines the cases. Following 1900 LST, it steadily increases and reaches 0.67°C at 0500 LST.

Equation (5) can be described as follows, neglecting the cooling term and the horizontal diffusion term, and converting to potential temperature:

$$
\frac{\partial \theta}{\partial x} + \frac{\partial \theta}{\partial z} = -\frac{1}{\rho_c \partial \theta \partial z}.
$$

Here, $H/\rho_c = -(K \partial \theta \partial z)$. At 1500 LST, a mixed layer is developed both over the rural and urban sites as shown.
in Fig. 4. We obtain Eq. (10) by integrating Eq. (9) with respect to \( z \) from the surface to the top of the urban boundary layer \( h_u \), and \( x \) from the rural site to the urban site of the central grid point \( x_u \):

\[
\rho c_p \Delta \theta_m h_i = \frac{\Delta H_{se} \Delta x_u}{U_m}. \tag{10}
\]

Here, \( U_m \) and \( \theta_m \) are the mixed-layer-averaged wind speed and potential temperature, respectively, and \( H_{se} \) is the heat flux from the surface, including the anthropogenic heat. The mixed-layer-averaged vertical advection is assumed to be zero for simplicity, although the advection could retrain an increase of the heat island intensity in the numerical model. The left-hand side of Eq. (10) represents the heat required to increase potential temperature by \( \Delta \theta \) in an atmospheric column of height \( h \). The right-hand side represents the additional heat that the column absorbs while advecting over the urban area; \( \Delta H_{se} \) is the difference between sensible heat flux from the urban and rural surfaces if the amount of the entrainment over urban areas is equal to that of rural areas. Thus,

\[
\Delta \theta_m = \frac{\Delta H_{se} \Delta x_u}{\rho c_p h_i U_m} - 0.01 \Delta H_{se}, \tag{11}
\]

where \( h_u = 1200 \text{ m}, \ U_m = 1 \text{ m} \text{s}^{-1}, \ x_u = 15 \text{ km}, \) and \( \rho c_p = 1200 \text{ J K}^{-1} \text{ m}^{-3} \). At 0300 LST, the strong stable layer is found over the rural site (Fig. 6). The figure shows that the heat required to increase surface air temperature by \( \Delta \theta_m \) is

\[
\Delta Q = \frac{1}{2} \rho c_p \Delta \theta_m \left( \frac{\Delta \theta_m}{1} \right) = \frac{1}{2} \rho c_p \Delta \theta_m h_i. \tag{12}
\]

Thus,

\[
\Delta \theta_m = \frac{2 \Delta H_{se} \Delta x_u}{\rho c_p h_i U_m} - 0.08 \Delta H_{se}, \tag{13}
\]

where \( h_u = 300 \text{ m}, \ U_m = 1 \text{ m} \text{s}^{-1}, \ x_u = 15 \text{ km}, \) and \( \rho c_p = 1200 \text{ J K}^{-1} \text{ m}^{-3} \). These results indicate that the impact of the anthropogenic heat during the nighttime is much larger than that of the daytime. Cases 1 and 2 are consistent with these results. The impact of the anthropogenic heat during the nighttime is about 3 times that of the daytime when the heat of 12.5 \text{ W m}^{-2} is constantly given in the model (figure is omitted).

Second, we investigate the impact of a larger volumetric heat capacity due to the existence of walls, that is, the vertical urban surface. This factor could be important for the diurnal range of the surface air temperature. The heat flux into the model system is calculated as the difference between the net radiation and the sensible and latent heat fluxes at the roof level \( z_r \). Neglecting horizontal heat advection and the sensible heat content of the air volume within the canopy layer, the heat flux into the model system is

\[
G = G_o + 2 \int_{z_r}^{z} \left[ \frac{\partial (\rho c_p T_z)}{\partial t} \right] \, dz. \tag{14}
\]

Here, \( G_o \) is the surface heat flux into the ground per unit area, including the roof and road, and \( \rho c_p \), \( T_z \), and \( T_y \) are the density, specific heat, and temperature of the building, respectively. The increased heat content due to the walls roughly equals the wall area ratio to the unit area. Hence, the effective heat capacity of case 4 could be set to be 60\% (\( \sim 1/1.7 \)) of case 2. The results from case 4 are also shown in Figs. 10 and 11. Figures 10 and 11 show that the smaller heat capacity increases the amplitude by 0.85\(^{\circ} \text{C} \) and delays the phase by 1 h. This indicates that this factor prevents temperature warming during the daytime and temperature cooling during the nighttime. The enhanced heat capacity dissipates the effect of the anthropogenic heat release for the daytime temperature and increases the nocturnal temperature by 0.39\(^{\circ} \text{C} \) at 0600 LST.

Third, to determine the role of the wall for the radiation, we imagine an urban canopy layer where the building walls are similar to air, but have the thermal inertia of concrete and do not absorb any radiation: \( S_w = L_w = 0 \) for case 5. Thus, the surface energy budget on the wall is

\[
0 = H_w + G_w. \tag{15}
\]

The road receives all downward short- and longwave radiation; thus, the surface energy budget is

\[
S_w(1 - \alpha) + S_g(1 - \alpha) + e_c(L_\downarrow - \sigma T_b^4) = H_o + G_g. \tag{16}
\]

Figures 10 and 11 illustrate the results from case 5. The diurnal variation of the temperature is significantly different from case 2, particularly for the cooling rate during the several hours after sunset. The cooling rate is the largest of all cases. This indicates that the modification of the radiation due to the walls results in uniform cooling at a constant rate, which is one of the essential features of the urban climate. As a result, the temperature difference between cases 2 and 5 reaches 0.52\(^{\circ} \text{C} \) around midnight.

Last, we discuss the impact of albedo on the temperature. In case 6, we ignore the reflected radiations in the canyon \( S_{w,2} \) and \( S_{g,2} \), as well as the anthropogenic heating. The difference in the temperature between cases 2 and 6 is less than 0.15\(^{\circ} \text{C} \) (figure is omitted). The downward solar radiation is 854 \text{ W m}^{-2}, and the net solar radiation of cases 2 and 6 are, respectively, 712 and 682 \text{ W m}^{-2} at 1300 LST. Enhanced net radiation of 30 \text{ W m}^{-2} is partitioned into upward longwave radiation, sensible heat flux, and heat flux into the ground. At most, half of the energy could be used as sensible heat flux. Albedo is smaller around sunset and before sunrise when compared with that around noon. However, the impact around sunset and sunrise is smaller than that around noon primarily because of a smaller...
downward solar radiation. The enhanced daytime and daily mean net solar radiations are 18 and 9.3 W m$^{-2}$, respectively.

5. Conclusions and remarks

An urban canopy-layer model was incorporated into a two-dimensional mesoscale model in order to examine the thermal effects of anthropogenic heating and canyon structure on the nocturnal heat island for an idealized case.

We first confirmed the following basic features of the urban heat island on a clear, calm summer day from the control run.

The model simulates a heat island intensity of 1.4°C at 1500 LST and 2.4°C at 0300 LST when the urban area has an artificial surface coverage ratio of 80%, a building coverage ratio of 30%, a building height of 6 m, and anthropogenic heat. The daily mean of the anthropogenic heating is assumed to be 25 W m$^{-2}$, and it is allowed to vary diurnally. The maximum anthropogenic heat is 39% greater than the mean, while the minimum is 43% less than the mean.

The model produces a linear decrease in the surface layer air temperature at the urban site after sunset, and the difference of the cooling rate between the urban area and its surroundings produces a nocturnal heat island.

Latent heat flux from the urban site is smaller than that of the rural site.

Larger heat storage during the daytime is simulated at the urban site.

Upward sensible heat flux after sunset is simulated at the urban site, and downward sensible heat flux is simulated at the rural site. The upward sensible heat flux from the urban surface produces a shallow mixed layer of 300 m at 0300 LST.

Second, sensitivity experiments were conducted in order to compare the impacts on the surface layer air temperature in the urban area of the following: anthropogenic heating, a larger heat capacity due to walls, a smaller sky-view factor due to walls, and a smaller albedo due to walls. We summarize as follows:

Anthropogenic heating increases the surface layer temperature through the day. However, the heating has a greater effect on the nocturnal temperature. The time of the maximum impact is 0500 LST, and its value is 0.6°C. The impact of the anthropogenic heating on the nocturnal heat island is found to be the largest of all of the factors.

The larger heat capacity due to the walls of the urban area decreases the diurnal range of the temperature by 0.85°C and delays the phase by 1 h. A maximum increase of 0.39°C occurs at 0600 LST because of this effect.

The smaller sky-view factor due to the walls increases the temperature during the first several hours after sunset. The maximum increase of 0.52°C occurs around midnight. This factor results in a uniform cooling at a constant rate, which is one of the essential features of urban climate.

The impact of albedo is not as large as the other factors, and its value is less than 0.15°C.

The present results are from an idealized simulation. However, they will prove useful for urban planning, for solving uncomfortable thermal environmental problems within a metropolitan area, and in numerical modeling of urban surface processes. The procedure will be repeated for the various more realistic land cover and atmospheric conditions, and the results will be presented in a future study.

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APPENDIX

Description of the Single-Layer Urban Canopy Model

a. Shortwave radiation flux

The effects of the shadows on the Lambertian surface are included in the model. The normalized shadow on the road is defined as

$$l_{\text{shadow}} = \begin{cases} h_0 \tan \theta_s \sin \theta_s & (l_{\text{shadow}} < l_{\text{road}}) \\ l_{\text{road}} & (l_{\text{shadow}} > l_{\text{road}}) \end{cases}.$$  (A1)

Here, $\theta_s$ is the solar zenith angle. The net amount that is absorbed by the roof, wall, and road is calculated from (cf. Fig. 2)

$$S_R = S_D(1 - \alpha_R) + S_Q(1 - \alpha_R),$$  (A2)

$$S_{W:1} = S_D \frac{l_{\text{shadow}}}{2h_0} (1 - \alpha_W) + S_Q F_{W-a}(1 - \alpha_W),$$  (A3)

$$S_{W:2} = S_D \frac{l_{\text{road}} - l_{\text{shadow}}}{l_{\text{road}}} \alpha_G F_{W-a}(1 - \alpha_W) + S_Q F_{W-a}(1 - \alpha_W).$$
+ S_p \frac{I_{\text{shadow}}}{2h_c} \alpha_w F_{W-w}(1 - \alpha_w) \\
+ S_p F_{W-w} \alpha_w F_{W-w}(1 - \alpha_w),

S_{G,1} = S_p \frac{I_{\text{road}} - I_{\text{shadow}}}{I_{\text{road}}} (1 - \alpha_G) + S_p F_{G-G} (1 - \alpha_G),

\text{(A4)}

\text{and}

S_{G,2} = S_p \frac{I_{\text{shadow}}}{2h_c} \alpha_w F_{G-G}(1 - \alpha_G) \\
+ S_p F_{G-G} \alpha_w F_{G-G}(1 - \alpha_G).

\text{(A5)}

The solar radiation is positive when directed toward the surface. Respectively, \(S_p\) and \(S_p\) are the direct solar radiation and the diffuse solar radiation received by a horizontal surface; the subscripts 1 and 2 refer to the absorption of the direct and reflected radiation; and the subscripts \(W, G,\) and \(S\) denote wall, ground (road), and sky, respectively. The view factors \(F\) are also computed in the same way as in Kusaka et al. (2001). The total direct solar radiation to a grid element is evaluated by the weighted average according to the relative area of different canyons. Diffuse solar and downward longwave radiation are assumed to be emitted from the entire sky; that is, it is assumed to be isotropic.

d. Longwave radiation flux

The net longwave fluxes absorbed by the roof, wall, and road surfaces are calculated from

\[ L_R = \varepsilon_p (L^a + \sigma T_R^4), \]

\[ L_{W,1} = \varepsilon_w (L^a F_{W-w} + \sigma T_{W-w}^4 F_{W-w} + \varepsilon_w \sigma T_{W-w}^4 F_{W-w}), \]

\[ L_{W,2} = \varepsilon_w \left[ (1 - \varepsilon_w) L^a F_{W-w} - \sigma T_{W-w}^4 F_{W-w} \right] \]

\[ + (1 - \varepsilon_w) L^a F_{W-w} + (1 - \varepsilon_w) \varepsilon_w \sigma T_{W-w}^4 F_{W-w} \]

\[ + (1 - \varepsilon_w) \varepsilon_w \sigma T_{W-w}^4 F_{W-w} + \varepsilon_w (1 - \varepsilon_w) \sigma T_{W-w}^4 F_{W-w} \]

\[ - \sigma T_{W-w}^4 \], \quad \text{(A8)}

\[ L_{G,1} = \varepsilon_G (L^a F_{G-G} + \varepsilon_G \sigma T_{G-G}^4 F_{G-G} - \sigma T_{G-G}^4), \quad \text{(A9)} \]

\[ L_{G,2} = \varepsilon_G \left[ (1 - \varepsilon_w) L^a F_{G-w} + (1 - \varepsilon_w) \varepsilon_G \sigma T_{G-w}^4 F_{G-w} \right] \]

\[ + \varepsilon_G (1 - \varepsilon_w) \sigma T_{G-w}^4 F_{G-w} \]. \quad \text{(A10)}

Here \(L^a\) is the downward atmospheric longwave radiation; respectively, \(T_R, T_W,\) and \(T_G\) are surface temperatures of the roof, wall, and road; and subscripts 1 and 2 refer to the absorption of the direct and reflected radiation.

c. Heat and momentum fluxes

Sensible heat flux from the building roof, building wall, and road is estimated at each surface individually. The sensible heat fluxes from the wall and road are calculated using the Jurges formula:

\[ H_w = C_w (T_w - T_s) \quad \text{and} \]

\[ H_G = C_G (T_G - T_s), \]

\[ \text{(A12)} \]

\[ \text{where} \]

\[ C_w = C_G = \begin{cases} 7.51 U_s^{0.78} & (U_s \geq 5 \text{ m s}^{-1}) \\ 6.15 + 4.18 U_s & (U_s < 5 \text{ m s}^{-1}). \end{cases} \]

\[ \text{(A14)} \]

The momentum, sensible heat, and latent heat exchanges between the canyon space and the overlying atmosphere are the flux through the canyon top, using the Monin–Obukhov similarity theory. In summary,

\[ \tau = -\rho \frac{k^2}{\Psi_m(\zeta)^2}, \]

\[ H = -\rho_c c_p (\theta - \theta_s) \frac{k u_o}{\Psi_h(\zeta)}, \quad \text{and} \]

\[ \bar{I}E = -\rho_v (q_v - q_{v,w}) \frac{k u_o}{\Psi_h(\zeta)}. \]

\[ \text{(A15)} \]

\[ \text{(A16)} \]

\[ \text{(A17)} \]

Here \(q_v\) is the specific humidity at \(z_T + d\), and \(\Psi_m\) and \(\Psi_h\) are the integrated nondimensional shear function:

\[ \Psi_m = \int \frac{\phi_m}{\xi} d\xi \quad \text{and} \]

\[ \Psi_h = \int \frac{\phi_h}{\xi} d\xi, \]

\[ \text{(A18)} \]

\[ \text{(A19)} \]

Here, \(\xi_0 = z_a/L, \xi_T = z_T/L,\) and \(\zeta = (z_s - d)/L,\) and \(L\) is the Monin–Obukhov stability length. The nondimensional shear functions of Dyer and Hicks (1970) are used for unstable conditions and those of Kondo et al. (1978) are used for stable conditions. Note that the upper limit of the integration for nondimensional shear functions is \(\zeta = (z_a - d)/L\) and the surface temperature for this scheme is the canyon air defined at the height of \(z_T + d\). This scheme is also applied for the fluxes from the roof.

The air within the urban canopy layer has a negligible heat capacity, and so sensible heat flux from the building wall \(H_w\) and from the road \(H_G\) must be balanced by the sensible heat flux to the atmosphere from the canyon space; that is,

\[ l_{\text{road}} H_u = 2h_c H_w + l_{\text{road}} H_G. \]

\[ \text{(A20)} \]

The total heat flux to the atmosphere from the urban area is the area-weighted average from the roof and that through the canyon top. If the ground surface in a grid area is classified into artificial and natural surfaces with area ratios \(A_A\) and \(A_N\), respectively, the total heat fluxes
at the grid points are the averaged heat fluxes on the artificial and vegetated surfaces weighted by their area (Kimura 1989).

\[ H = A_H H_A + A_N H_N. \]  
(A21)

Here, \( H_A \) and \( H_N \) are sensible heat fluxes from artificial and natural surfaces; \( H_N \) can be estimated from a land surface model. (The standard slab grassland model is used in the present study.)

d. Wind speed within the canopy

The mean wind speed in the canopy layer is used as a reference to calculate \( H_C \) and \( H_W \). Within the canopy layer, buildings are usually diffusely distributed and act as a continuumlike sink for momentum. Under such conditions the mean wind speed and shearing stress should decay with depth below the canopy top. In the present model, the stress divergence in the air is assumed to balance the drag per unit volume of air,

\[ -\frac{\partial}{\partial z} \left( \gamma_r K_m \frac{\partial U}{\partial z} \right) - \gamma_c c_d a_z U = 0, \]  
(A22)

where eddy viscosity model is used. Here, \( a \) is a function of height and is equal to the surface area of the buildings per unit volume of air, and \( c_d \) is a drag coefficient; \( \gamma_r \) is the effective fluid volume. The present study uses the assumptions of a constant mixing length \( l \) and drag coefficient in the canopy layer by Inoue (1963). The result is close to an exponential wind profile given by

\[ U = U_o \exp \left[ -n \left( 1 - \frac{z}{z_r} \right) \right]. \]  
(A23)

Here, \( n = z_r (c_d a_z / 2 l)^{1/3} \). In fact, wind profiles within an urban canopy from filed observations are close to the exponential (e.g., Depaul and Sheih 1986; Rotach 1995).

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