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# Urban climates and heat islands: albedo, evapotranspiration, and anthropogenic heat

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**Abstract**

As an introduction to this special issue on urban heat islands and cool communities, this paper reviews some of the characteristics of urban climates and the causes and effects of urban heat islands. In particular, the impacts of surface albedo, evapotranspiration, and anthropogenic heating on the near-surface climate are discussed. Numerical simulations and field measurements indicate that increasing albedo and vegetation cover can be effective in reducing the surface and air temperatures near the ground.

*Keywords:* Albedo; Anthropogenic heating; Meteorology; Micrometeorology; Numerical modeling; Simulation; Urban climate; Vegetation

## 1. Introduction

It is known that urban climates differ from those of rural areas and that the magnitudes of the differences can be quite large at times depending on weather conditions, urban thermophysical and geometrical characteristics, and anthropogenic moisture and heat sources present in the area. For example, northern hemisphere urban areas annually have an average of 12% less solar radiation, 8% more clouds, 14% more rainfall, 10% more snowfall, and 15% more thunderstorms than their rural counterparts [1,2]. Urban pollutant concentrations can be 10 times higher than those of the 'clean' atmosphere and air temperatures can be on the average 2°C higher. The fluxes of heat, moisture, and momentum are significantly altered by the urban landscape and the contrast between the urban and 'undisturbed' climates is further enhanced by the input of anthropogenic heat, moisture, and pollutants into the atmosphere.

In most mid- and high latitude cities, urban air temperatures are generally higher than their corresponding rural values. This phenomenon, the urban heat island, has been recognized since the turn of this century and has been well documented [3–6]. A heat island can occur at a range of scales; it can manifest itself around a single building [7], a small vegetative canopy [8,9], or a large portion of a city. Depending on geographic location and prevailing weather conditions, heat islands may be beneficial or detrimental to the urban dweller and energy user. Generally speaking, low and mid-latitude heat islands are unwanted because they contribute to cooling loads, thermal discomfort, and air pollution whereas high latitude heat islands are less of a problem because they can reduce heating energy requirements. This is a generalization however; the actual impacts of urban climates and heat islands depend on the characteristics of local climates. One way of indirectly characterizing these impacts is to examine heating and cooling degree-days data. Table 1 shows that with respect to rural surrounds, urban areas have fewer heating degree-days (HDD) but more cooling degree-days (CDD). Taha et al. [10] show that the net effect of HDD and CDD modification by urban areas is an increase in cooling loads.

The causes and effects of urban climates and heat islands are diverse and their interactions complex. In this paper, three parameters of direct relevance to heat islands are examined in very simple terms. These are the surface albedo, evapo-

Table 1
Reduction in heating degree-days and increase in cooling degree-days (base 18.3°C) due to urbanization and heat island effects. Averages for selected locations for the period 1941–1970. Modified after Landsberg [44]

| Location | Heating degree-days | | | Cooling degree-days | | |
|---|---|---|---|---|---|---|
| | Urban | Airport | Δ% | Urban | Airport | Δ% |
| Los Angeles | 384 | 562 | −32 | 368 | 191 | +92 |
| Washington DC | 1300 | 1370 | −6 | 440 | 361 | +21 |
| St. Louis | 1384 | 1466 | −6 | 510 | 459 | +11 |
| New York | 1496 | 1600 | −7 | 333 | 268 | +24 |
| Baltimore | 1266 | 1459 | −14 | 464 | 344 | +35 |
| Seattle | 2493 | 2881 | −13 | 111 | 72 | +54 |
| Detroit | 3460 | 3556 | −3 | 416 | 366 | +14 |
| Chicago | 3371 | 3609 | −7 | 463 | 372 | +24 |
| Denver | 3058 | 3342 | −8 | 416 | 350 | +19 |

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transpiration from vegetation, and anthropogenic heating from mobile and stationary sources. They are discussed as factors leading to differences between urban and rural climates. The sensitivity of the temperature field to changes in these parameters is also discussed based on results from numerical simulations. For this purpose, a version of the URBMET model [11–13] was used to simulate typical summer days in a mid latitude location with warm climate. The Colorado State University Mesoscale Model (CSUMM), originally developed by Pielke [29], was also used to simulate the effects of large scale increases in surface albedo and vegetation cover on regional meteorology.

## 2. The surface energy balance equation

The surface energy balance equation explains quite simply the roles of surface properties and anthropogenic heating in near-surface climates. The general form of this equation for a unit surface area which is flat, horizontal, and homogeneous, is given by:

$$\alpha_s Q + \alpha_s q + \alpha_L L + Q_f = \epsilon\sigma T_o^4 + h_c(T_o - T_a) + k\frac{dT_s}{dz}\Big|_{Z=0} + \lambda E \qquad (1)$$

where $\alpha_s$ and $\alpha_L$ are absorptivities for short and long wave radiation, respectively, $Q$ and $q$ are the direct and diffuse short wave radiative fluxes, respectively, $L$ is the incoming long wave radiative flux, $Q_f$ is the anthropogenic (man-made) heat flux, $\epsilon$ is the emissivity of the surface, $\sigma$ is the Stefan–Boltzmann constant, $T_o$ and $T_a$ are, respectively, the surface and air temperatures, $h_c$ is a convective heat transfer coefficient, $k$ is the thermal conductivity of the ground, $T_s$ is the surface temperature, $\lambda$ is the latent heat of vaporization, and $E$ is the evaporation rate.

The first two terms on the left-hand side represent the absorbed solar radiation, the third term is the absorbed long wave radiation, and the fourth term is anthropogenic heat flux at the surface. The first and second terms on the right-hand side represent the radiative and sensible heat fluxes, the third term represents heat conduction into or from the submedium through the surface, and the fourth term is the latent heat flux. Eq. (1) can be recast as:

$$(1-a)I + L^* + Q_f = H + \lambda E + G \qquad (2)$$

where $a$ is the solar albedo, $I$ is incoming solar radiation, $L^*$ is net long wave radiation at the surface, $Q_f$ is anthropogenic heat, and $H$, $\lambda E$, and $G$ are the sensible, latent, and ground heat fluxes, respectively.

In the following sections, three of the terms in Eq. (2) are discussed, namely: surface albedo, $a$, latent heat flux/evapotranspiration, $\lambda E$, and anthropogenic heating, $Q_f$. Their impacts on the temperature field, which in turn affects the amount of cooling energy needed, are also described.

### 2.1. Albedo (a)

The albedo of a surface is defined as its hemispherically- and wavelength-integrated reflectivity. This definition applies to simple uniform surfaces as well as to heterogeneous and complex ones. Typically, urban albedos are in the range 0.10 to 0.20 but in some cities these values can be exceeded. North African towns are good examples of high albedo urbanized areas (albedos of 0.30 to 0.45) whereas most US and European cities have lower albedos (0.15 to 0.20). For example, Taha [14] performed low altitude flights over the Los Angeles Basin and found that the highest albedo in that area was 0.20 (near Downtown Los Angeles). He also found that much of the urbanized basin had an albedo between 0.12 and 0.16 and that city-core albedo was higher than that of its surrounds partly because of more extensive vegetation in the latter.

Table 2 lists values of snow-free urban albedos for several cities. Where available, the difference between urban and rural albedos is given in the third column. The example of Lagos is interesting because the city is built on an intricate

Table 2
Selected urban albedo values

| Urban area | Albedo | Δ (urban–rural) | Author |
|---|---|---|---|
| Los Angeles (city core) | 0.20 | 0.09 | Taha [a] [14] |
| Madison, WI (urban) | 0.15–0.18 | 0.02 | Kung et al. [33] |
| St. Louis, MI (urban) | 0.12–0.14 | | Dabberdt and Davis [34] |
| St. Louis, MI (center) | 0.19–0.16 | 0.03 | Vukovich [b] [35] |
| Hartford, CT (urban) | 0.09–0.14 | | Brest [36] |
| Adelaide, AUS (commercial) | 0.27 (mean) | 0.09 | Coppin et al. [37] |
| Hamilton, Ontario | 0.12–0.13 | | Rouse and Bello [38] |
| Munich, West Germany | 0.16 (mean) | −0.08 | Mayer and Noack [39] |
| Vancouver, BC | 0.13–0.15 | | Steyn and Oke [40] |
| Tokyo | 0.10 (mean) | −0.02 | Aida [41] |
| Ibadan, Nigeria | 0.12 (mean) | 0.03 | Oguntoyinbo [42] |
| Lagos, Nigeria | 0.45 | 0.25 | Oguntoyinbo [43] |

[a] Measured from low altitude aircraft flights ( < 200m above ground level) in summer 1993.
[b] Limited shortwave sensitivity of sensors.

mesh of creeks and lagoons that cause albedo to decrease as one gets closer to the suburbs (the albedo of water is lower than that of the light-colored buildings).

Using high albedo materials reduces the amount of solar radiation absorbed through building envelopes and urban structures and keeps their surfaces cooler. Taha et al. [15] measured the albedo and surface temperatures of a variety of surfaces in the field. They found, for example, that white elastomeric coatings (with an albedo of 0.72) were 45°C cooler than black coatings (with an albedo of 0.08) in the early afternoon of a clear day in summer. A white surface with an albedo of 0.61 was only 5°C warmer than ambient air whereas conventional gravel with an albedo of 0.09 was 30°C warmer than air.

The reduction in surface temperature also reduces the intensity of long wave radiation. Local and downwind ambient air temperatures would be lower because of smaller convective heat fluxes from cooler surfaces. Such temperature reductions can have significant impacts on cooling energy consumption in urban areas, a fact of particular importance in hot climate cities. Thus, the sensitivity of air temperature to albedo change has been of interest to those researching the impacts of urban heat islands.

One-dimensional meteorological simulations by Taha et al. [16] showed that localized afternoon air temperatures on summer days can be lowered by as much as 4°C by changing the surface albedo from 0.25 to 0.40 in a typical mid-latitude warm climate. Three-dimensional mesoscale simulations of the effects of large scale albedo increases in the Los Angeles basin showed similar magnitudes of impacts on air temperature in summer; an average decrease of 2°C and up to 4°C may be possible by increasing the albedo by 0.13 in urbanized areas of the basin [17,18]. Temperature decreases of this magnitude could reduce the electricity load from air conditioning by 10% [19] and smog (ozone concentrations) by up to 20% during a hot summer day [18].

### 2.2. *Latent heat flux ($\lambda E$)*

Evapotranspiration (evaporation and transpiration) from soil–vegetation systems is another effective moderator of near-surface climates, particularly in the warm and dry mid- and low latitudes. Given the right conditions, evapotranspiration can create 'oases' that are 2–8°C cooler than their surroundings [4,8,9]. In extreme oasis conditions, the latent heat flux ($\lambda E$) can be so large that the sensible heat flux ($H$) becomes negative, meaning that the air above vegetation and over the dry surroundings must supply sensible heat to the vegetated area and the Bowen ratio (ratio of sensible to latent heat fluxes) becomes negative. For example, Flohn [20] reported that in the Tunisian deserts, oases could develop with Bowen ratios ($\beta$) of $-0.26$.

In more average oasis conditions, Bowen ratios in vegetative canopies are within 0.5–2. For example, measured noontime sensible and latent heat flux densities in a pine forest in England in July were 400 and 200 W m$^{-2}$ ($\beta=2$), respectively [21]. In a Douglas fir stand in British Columbia, the corresponding values were 200 and 300 W m$^{-2}$ ($\beta=0.66$) [22]. In comparison, $\beta$ in urban areas is typically around 5, in a desert it is in the neighborhood of 10, and over tropical oceans, it is about 0.1.

Urban areas, with extensive impervious surfaces, have generally more runoff than their rural counterparts. The runoff water drains quickly and, in the long run, less surface water remains available for evapotranspiration, thus affecting the urban surface energy balance [23]. The lower evapotranspiration rate in urban areas is a major factor in increasing daytime temperatures. In Tokyo for example, vegetated zones in summer are on the average 1.6°C cooler than non-vegetated spots [24,25], and in Montreal, urban parks can be 2.5°C cooler than surrounding built areas [26]. In a vegetative canopy (orchard) in Davis, CA, Taha et al. [8,9] measured typical summer daytime oases 2°C cooler than surrounding open fields of bare soil. On one hot and clear day, the oasis grew as large as 6°C. Robinette [27] reported summer afternoon temperatures 8°C lower at 50 cm height over grass than over bare soil.

In numerical simulations, Taha [12] incorporated a bulk vegetative canopy parametrization in the one-dimensional URBMET PBL model [11]. The sensitivity of air temperature to vegetation cover change was simulated with this modified version of URBMET using initial conditions from micrometeorological field measurements in Davis, California [8,9]. The vegetation canopy produced daytime temperature depressions (oases) and nighttime excesses (heat islands) compared to the bare surrounds. The factors behind the first are evaporative cooling and shading of the ground, whereas the second is a result of reduced sky view factor within the canopy. The simulations indicate that a vegetative cover of 30% could produce a noontime oasis of up to 6°C, in favorable conditions, and a nighttime heat island of 2°C. In other modeling studies, Gao [28] found that green areas decrease maximum and average air temperatures by 2°C. Also, Gao [28] found that vegetation can decrease maximum air temperatures in streets by 2°C. Regional scale, three-dimensional simulations showed that large scale urban forestation in the Los Angeles basin can be as effective in reducing air temperature as the use of high albedo materials [17,30].

### 2.3. *Anthropogenic heating ($Q_f$)*

Anthropogenic heating in urban areas can affect the near-surface air temperature and potentially play a role in creating urban heat islands. Although the magnitude of $Q_f$ depends on various factors such as the intensity of energy use, power generation, and transportation systems, the largest $Q_f$ values are typically found in cold climate urban centers in winter because of the intensive heating load during that season. In most major US cities, $Q_f$ ranges between 20 and 40 W m$^{-2}$ in summer and between 70 and 210 W m$^{-2}$ in winter [31]. These values are for city centers and the last one is an uncommon extreme. By comparison, solar radiation at the surface

Table 3
Anthropogenic heating ($Q_f$) and net all wave radiation ( $Q*$ ) in selected cities. These are annual average values within the urbanized limits of the cities. The data does not include suburban or rural surrounds [3,5,20–22,31,34,39]

| City | $Q_f$ (W m$^{-2}$) | $Q*$ (W m$^{-2}$) |
|---|---|---|
| Chicago | 53 | |
| Cincinnati | 26 | |
| Los Angeles | 21 | 108 |
| Fairbanks | 19 | 18 |
| St. Louis | 16 | |
| Manhattan, NY City | 117–159 | 93 |
| Moscow | 127 | |
| Montreal | 99 | 52 |
| Budapest | 43 | 46 |
| Osaka | 26 | |
| Vancouver | 19 | |
| West Berlin | 21 | 57 |

on a clear or partly cloudy day at noon ranges from about 700 to 1000 W m$^{-2}$. In Table 3, the annual average anthropogenic heating for selected city cores is given along with a representative local annual average net all wave horizontal radiation ($Q^*$) where available. In residential and suburban areas, $Q_f$ is negligible.

Taha et al. [32] examined energy use patterns in buildings and motor vehicles and developed a diurnal anthropogenic heating profile based on estimated heat rejection from these sources. The profile was then used to parameterize time-dependent anthropogenic heating in prognostic meteorological models to estimate the impact of $Q_f$ on air temperatures. These meteorological simulations show that $Q_f$ in a large city core can create a heat island of up to 2–3°C both during the day and at night. This kind of temperature increase, however, is not expected to occur in a suburban residential neighborhood. Also, the simulations indicate that the low $Q_f$ in residential areas has a negligible impact on air temperature. Yet, combining the effects of anthropogenic heating in city centers with those of other factors, such as low vegetation cover and dark surfaces, can enhance the heat island effect.

## 3. Conclusions

Some of the aspects of the causes and effects of heat islands were examined. Field monitoring data and meteorological simulations indicate that changes in surface albedo and vegetation cover can be effective in modifying the near-surface climate. The effects of anthropogenic heating, however, seem to be relatively small. The simulations indicate that the impact of anthropogenic heating may be important in urban centers but negligible in residential and commercial areas.

Results from meteorological simulations suggest that cities can feasibly reverse heat islands and offset their impacts on energy use simply by increasing the albedo of roofing and paving materials and reforesting urban areas. The simulations suggest that reasonable increases in urban albedo can achieve a decrease of up to 2°C in air temperature. With extreme increases in albedo, localized decreases in air temperature can reach 4°C under some circumstances. Increases in vegetation in urban areas can result in some 2°C decrease in air temperatures. Under some circumstances, e.g., potentially evaporating soil–vegetation systems and favorable meteorological conditions, the localized decrease in air temperature can reach 4°C.

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