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Global effects of albedo change due to urbanization

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Abstract

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Human population on Earth is increasing and is projected to continue increasing for many years to come. Most of the increase will probably occur in cities and the extent of urban areas will increase with it. As cities grow larger and new cities are built, the many and diverse climatic effects of urbanization will become more evident.

A specific climatic effect of urbanization which has not received as much attention as it might deserve is the change in the reflective properties of the surface. This report aims to investigate the climatic effects of urbanization, especially albedo (the ratio of incoming to outgoing solar radiation for a given surface), by reviewing relevant literature. Through this it aims to answer what the effects of albedo change due to urbanization are in a global perspective.

Urbanization has many effects on surface albedo. It reduces it through the introduction of dark surfaces, through changing the surface geometry to better capture solar radiation, and through reducing snow cover in winter. It can also increase it by increasing the cloud cover and by emitting air pollutants and aerosols. Some effects, such as a modified surface moisture regime or climate changes downwind of cities, can act to both reduce and increase it, with the net effect very dependant on local factors.

Studies report that cities usually have lower albedo values than rural surfaces, commonly 2-5 % lower than crop-lands at the same latitude for example. The radiative forcing of this albedo reduction is very small. This is mainly because the fraction of Earth's land surface covered by urban areas is only 0.44 % according to the best estimate. While the climatic effects of albedo change due to urbanization are small on a global scale, they are quite distinct on regional to local scales. The small extent of urban areas today is still more than twice as large as the estimate cited in the last assessment report by the IPCC (AR4). With the projected increase in urban populations, the small global effects will become progressively more important and will need to be taken into account in future studies of climate and land use change.

Keywords: Geography, Physical Geography, Albedo, Urbanization, the Energy Balance, solar radiation, land use, land cover.

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Sammanfattning (Swedish abstract)

Mattias Spångmyr

Kan städernas mörka sidor förvärra klimatuppvärmningen?

Global effects of albedo change due to urbanization

Hur många gånger har du tagit på dig en svart tröja och sedan upptäckt hur varmt det blir när solen tittar fram? Med hjälp av den erfarenheten är det säkert tydligt att mörka material värms upp mer av solens strålar än ljusa material. Man säger att mörka material ”tar till sig”, eller *absorberar,* en större andel av den strålning som solen sänder ut i form av ljus. Det som inte absorberas kan istället ”studsa vidare”, eller *reflekteras*, och andelen strålning som reflekteras av den totala mängden inkommande strålning kallas för *albedo*.

Precis som svarta tröjor har även andra mörka ytor ett lågt albedo. Asfalt, mörka tak och väggar med mera är exempel på ytor som ofta hittas där människor bor och jobbar. I städer och tätbebyggelse utgör dessa ytor en stor del av markytan. Därför bidrar de till att sänka tätbebyggelsens albedo och orsaka en uppvärmning. Den här rapporten syftar till att, genom att granska relevant litteratur, undersöka hur städernas utbredning, eller *urbaniseringen*, påverkar det globala klimatet med fokus på dess påverkan på markytans albedo.

Mörka ytor utgör bara en av tätbebyggelsens effekter på markytans albedo. Vissa effekter, såsom ökat molntäcke och utsläpp av luftföroreningar, kan höja tätbebyggelsens albedo (sett från utanför atmosfären) medan andra både kan öka och minska albedo. Mörka ytor, ett minskat snötäcke i tätbebyggelse på vintern och att solens strålar kan reflekteras flera gånger mellan byggnader så att mer absorberas, gör att tätbebyggelse oftast har ett lägre albedo än dess omgivning.

Effekten av tätbebyggelsens förändring av albedo blir mycket liten på global nivå, eftersom tätbebyggelsen täcker så liten yta när man ser till hela jorden. Ungefär 0.44 % av den totala landytan beräknas vara täckt av tätbebyggelse. Även om ytan är liten är den förmodligen minst dubbelt så stor som t.ex. den internationella klimatpanelen IPCC citerade i sin senaste rapport och även om effekterna är små ur ett globalt perspektiv påverkar urbaniseringen klimatet högst märkbart på regional till lokal nivå. Då allt fler människor kommer att bo i städer och städerna kommer att växa och bli fler, kommer städernas effekter på klimatet att bli allt viktigare i framtiden. Därför måste forskningen på området fortsätta och urbaniseringen bör tas i beaktande i framtida studier av klimat- och markyteförändringar.

Nyckelord: Geografi, naturgeografi, albedo, urbanisering, energibalansen, solinstrålning, markanvändning, marktäcke.

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1 Introduction

For the past few decades, the issue of global warming has increased public awareness of the magnitude of our effect on this planet. At the prospect of increased global air and ocean temperatures and other, perhaps drastic, changes to our climate and environment, the international community is beginning to take measures to reduce the negative impacts.

At the forefront of attention is the International Panel on Climate Change (IPCC) who since its founding has monitored and reviewed research in these matters and who released its Fourth Assessment Report (AR4; Solomon et al., 2007) in 2007. Along with describing the issues and research, it also presents estimations of the current levels of scientific understanding pertaining to the phenomena discussed. While the effects of long-lived green-house gas emissions are well known and labelled with a *high* level of scientific understanding by the IPCC in the AR4, other factors such as changes in surface albedo due to land cover change (*medium to low* level of scientific understanding) still require more research before their contribution can be confidently quantified.

The assessment of the contribution of land cover change in the AR4 focuses mainly on deforestation since 1750 and suggests a net cooling effect since crop-lands, for example, usually have a higher albedo than forest. That is, they reflect a higher portion of the incoming solar radiation. A specific land cover change that could reduce this cooling effect is urbanization. Conversion of high albedo surfaces such as some crop-lands into urban surfaces like asphalt or dark walls can lower the albedo considerably (see Table 2.1). While the AR4 recognizes that urbanization may alter local and regional climates in several ways, potential effects on the global scale are largely overlooked.

Viewing the effects of urbanization as negligible on a global scale might be tempting. Urban areas cover only 0.046 % of the earth's surface according to an estimation given in the AR4. This estimation, however, originates from global land cover data (Loveland et al., 2000) which for the urban areas used a data layer from the US Defense Mapping Agency's Digital Chart of the World (DCW; Danko, 1992). The data layer, in turn, is drawn from map sources made between 1950 and 1979 as also pointed out by Potere et al. (2009). Needless to say, the extent of urban areas has increased dramatically over the last 60 years, rendering the global effects of urbanization progressively less negligible.

In fact, according to a publication by the United Nations Population Fund (UNFPA, 2007) the size of the world's urban population will have grown to 4.9 billion people by 2030, with most of the growth occurring in the developing world. Feddema et al. (2005) and Pielke (2005) among others agree that land cover changes need to be included when analysing global climate change. To fully appreciate the effects of land cover change, urbanization can not be ignored.

1.1 Objectives

The aim of this report is to investigate the climatic effects of urbanization with special focus on albedo, to review the nature and magnitude of urbanization, and to use this knowledge to answer the following question:

- What are the effects on global climate from the change in surface albedo due to urbanization?

2 Background & theory

In order to understand Earth's climate system, conceptual frameworks are commonly developed, (e.g. Solomon et al., 2007; NRC, 2005) separating anthropogenic, natural and feedback processes which affect the climate through direct and indirect climate forcing. To be able to fully grasp the complexity of the climate, however, one need to understand the basic energy and water balances as well as each individual aspect in great detail.

While the study of most of these factors affecting the climate is outside the scope of this report, the focus on albedo and urban environments requires thorough explanations of the Earth's energy balance, the definition of albedo itself and of specific urban climate effects.

2.1 The energy balance

Earth's primary source of energy is the electromagnetic radiation emitted by the sun. All objects emit electromagnetic radiation. The intensity depends on the object's temperature and the wavelength, as defined by Planck's law. The total intensity, integrated over all wavelengths, is the area under the curves (see Figure 2.1) and can be calculated with the Stefan-Boltzmann law:

$$E = \sigma T^4 \qquad \text{Equation 2.1}$$

where E is the maximum radiation rate in Watts per square meter (W/m^2), σ is the Stefan-Boltzmann constant, $5.67 \cdot 10^{-8}$ $W \cdot m^{-2} \cdot K^{-4}$, and T is the object's temperature in degrees Kelvin (Stull, 2000).

Equation 2.1 is valid only for objects that emit (and absorb) at their maximum rates. These are called blackbodies and the Earth and sun practically act as such. Most objects, however, do not. To get the radiation rate of non-blackbody objects Equation 2.1 need to be multiplied with the object's emissivity (ε), that is, the ratio of actual to maximum (blackbody) radiation for that temperature (Oke, 1987).

The sun emits the most radiation in wavelengths between 0.36 and 0.75 μm, which are the wavelengths we can see as visible light (Oke, 1987). At what wavelengths the most radiation will be emitted is also determined by temperature according to Wien's law:

$$\lambda_{max} = a \,/\, T \qquad \text{Equation 2.2}$$

where λ_{max} is the wavelength of peak emission (μm), a is 2897 μm·K and T is the object's temperature in degrees Kelvin (Stull, 2000). Figure 2.1 shows the emission rates of the sun and Earth for different wavelengths, notice the scale differences.

By using Equations 2.1 and 2.2 one can calculate the peak wavelength and the amount of energy emitted by the sun. With T_{sun} from Figure 2.1, λ_{max} is 0.501 μm and E is $6.33 \cdot 10^7$ W/m^2. This extreme energy output is spread over an increasingly large area as the radiation travels through space. At the top of Earth's atmosphere the amount of energy received, measured by satellite, is reduced to 1368±7 W/m^2 for surfaces perpendicular to the radiation (Stull, 2000).

This energy input of 1368±7 W/m^2, also called the solar constant, is what drives the climate system of Earth's surface and atmosphere. The annual average over the top of the atmosphere is a fourth of the solar constant (Oke, 1987). It is represented by (a) in Figure 2.2, which shows the energy balance created. Some of this energy is absorbed, increasing Earth's temperature, which also increases its emission according to Equation 2.1. As Figure 2.1 shows, λ_{max} for Earth is just above 11 µm, which is very much larger compared to the sun's 0.501 µm and the two radiation regimes nearly do not overlap at all. From this relationship, radiation with wavelengths associated with solar emission (0.15-3.0 µm) is called short-wave radiation and that associated with Earth emission (3.0-100 µm) is called long-wave radiation (Oke, 1987).

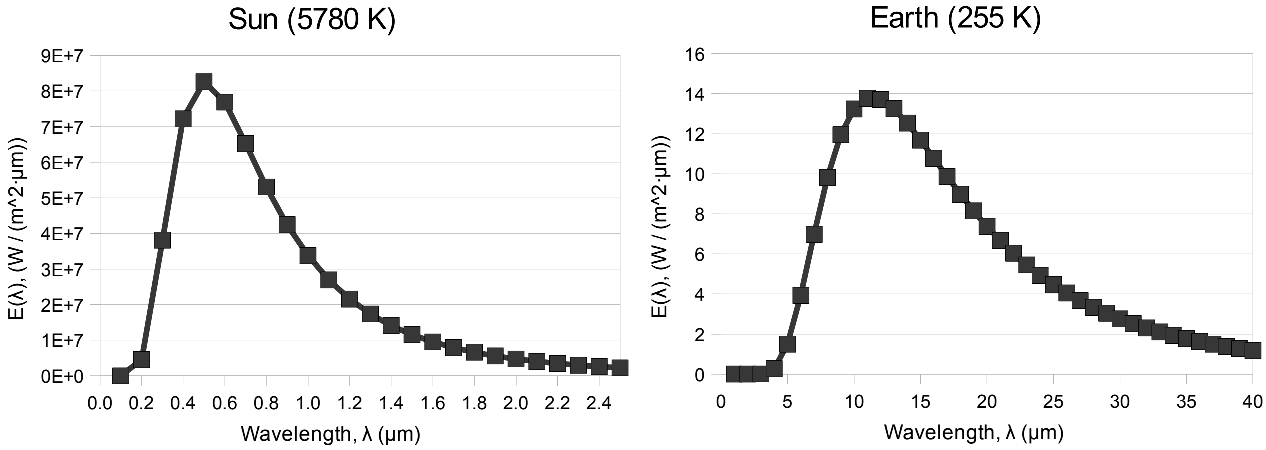

Figure 2.1: Radiation emitted by the sun and Earth as blackbodies, given temperatures of 5780 K and 255 K (for the whole Earth-atmosphere system), respectively. Computed using Planck's law. Equation and temperatures from Stull (2000).

Looking more closely at Figure 2.2, 1368±7 W/m^2 of short-wave radiation input to the top of the atmosphere is translated to 100 units of energy (a). Out of these, 6 units are scattered (reflected) back into space (b) by atmospheric constituents such as aerosols, water vapour and other gases. Another 20 are reflected by clouds (c). In addition, these same constituents absorb 19 units (d). That leaves 55 units which are transmitted through the atmosphere to the surface. 4 of the 55 units are then reflected (e) and the remaining 51 units are absorbed (f), increasing the temperature of the surface (Ahrens, 2009).

The surface temperature causes it to emit 117 units of energy (g). 6 are transmitted directly into space (h) and the rest are absorbed by the atmosphere (i). The temperature increase do not only fuel long-wave radiation; some of it is conducted as sensible heat from the molecules of the top layer of the surface to the air molecules closest to the ground. Conduction increases the air temperature near the ground which causes volumes of air to rise. This is called convection, and the conduction and convection together transfer 7 units of energy (j) from the surface higher into the atmosphere. 23 additional units are transferred from the surface into the atmosphere (k) in the form of latent heat (Ahrens, 2009).

Latent heat is the energy stored in or released from a substance as it changes phase (Ahrens, 2009). In this case, water is the transferring substance. All liquid water on Earth's surface, that is, oceans, lakes and rivers etc. is evaporating at a rate depending on its temperature. Evaporation (changing into the gas phase – water vapour) requires energy, and thus cools the surface. At the same time, all vegetation is releasing water vapour in a process called transpiration, through openings in their leaves called stomata (Chapin et al., 2002). Together, evaporation and transpiration are called evapotranspiration. Water vapour being transported by winds and convection etc. then rise higher into the atmosphere where it condensates (changes back into liquid form), releasing the same amount of energy that was withdrawn from the surface (Oke, 1987).

Many of the atmospheric constituents are so called selective absorbers (Ahrens, 2009). This means they absorb more radiation in some wavelengths than in others, specifically absorbing more long-wave than short-wave radiation. This effect, called the green-house effect, is responsible for Earth's comfortable surface temperatures. In other words; the capacity of Earth's surface to radiate much more long-wave radiation than it absorbs short-wave (compare g with f).

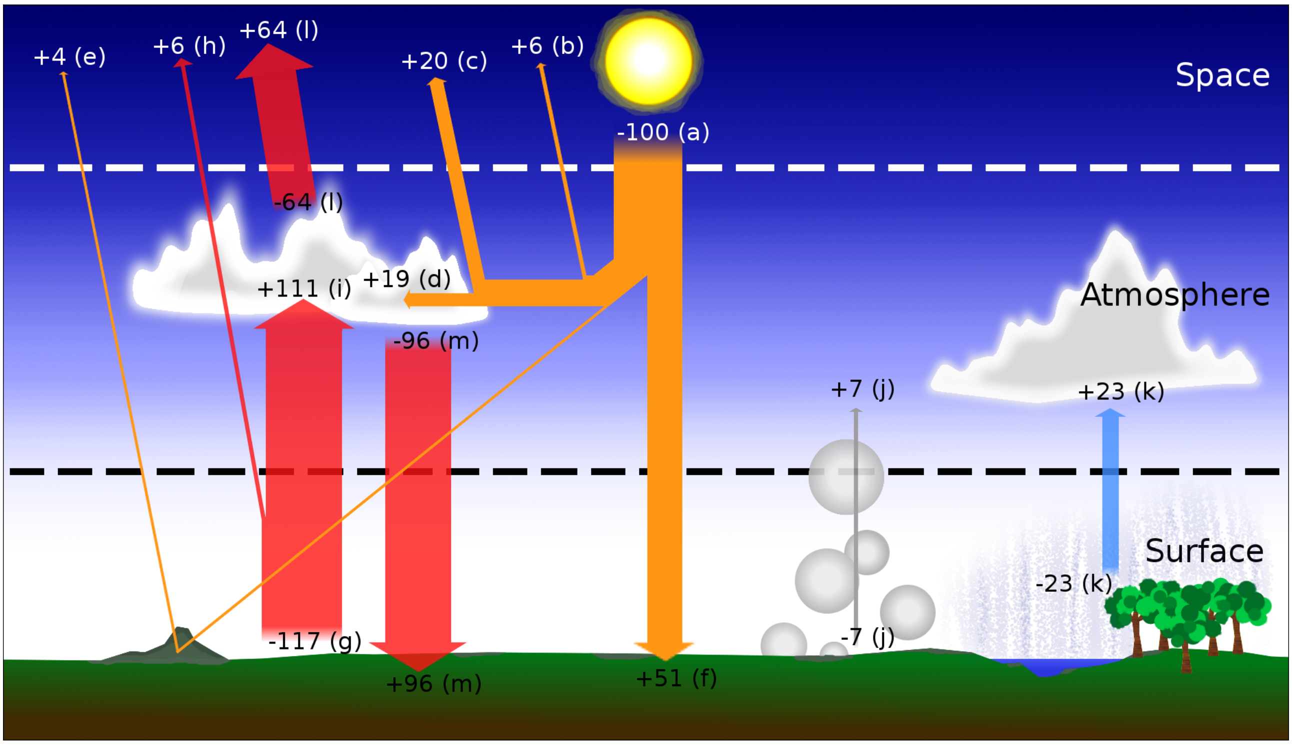

Figure 2.2: The annual energy balance of Earth's surface and atmosphere, showing equal input and output of energy in each zone. Number approximations from Ahrens (2009).

a) Incoming (short-wave) solar radiation at the top of Earth's atmosphere.
b) Short-wave radiation scattered back into space by the atmosphere.
c) Short-wave radiation reflected back into space by clouds.
d) Short-wave radiation absorbed by clouds and atmosphere.
e) Short-wave radiation reflected back into space by the surface.
f) Short-wave radiation absorbed by the surface.
g) Long-wave radiation emitted by the surface.
h) Long-wave radiation emitted by the surface which passes through the atmosphere out into space.
i) Long-wave radiation emitted by the surface and absorbed by the atmosphere.
j) Conductive and convective heat transfer from the surface to the atmosphere.
k) Latent heat transfer from the surface to the atmosphere by evapotranspiration and condensation of water.
l) Long-wave radiation from the atmosphere emitted out into space.
m) Long-wave radiation from the atmosphere emitted back to the surface.

The total energy absorbed by the atmosphere, in the form of short-wave (d) and long-wave (i) radiation and conductive, convective (j) and latent (k) heat transfer, can be summed up to 160 units of energy. This energy, like with the surface, increases the temperature, causing it to emit long-wave radiation. 64 units are lost to space (l) and 96 units (m) are radiated back to the surface (causing the green-house effect). Thus, the atmosphere emits as much energy as it absorbs, just like the surface and the Earth-atmosphere system as a whole, creating the balance of energy which keeps our global climate from changing too drastically on a short time-scale (Ahrens, 2009). This is illustrated in Figure 2.2 where the positive and negative values exactly balance each other out in each zone.

2.2 Albedo

Nearly everyone have probably experienced the warming effect of wearing dark clothes in sunlight. This effect is due to the fact that less visible light is reflected off of dark substances. What we perceive as colours are the specific combinations of wavelengths of the radiation reflected off a surface (Ahrens, 2009).

The definition of albedo according to Stull (2000, p. 35) is: "the ratio of total reflected to total incoming solar radiation (i.e., averaged over all solar wavelengths)". As can be seen in Figure 2.2, about 30 % of the annually received solar radiation is reflected or scattered (b,c and e), which gives Earth an effective albedo of approximately 0.3 (Ahrens, 2009). The average surface albedo is around half of that (Kukla and Robinson, 1980) but bear in mind that most of Earth's surface is covered by water, which has a very low albedo.

With the definition, the relationship between the colour of different objects and its albedo is clear; darker objects usually having a lower albedo than light coloured objects. The reason why substances have different characteristics in reflectivity, absorptivity and emissivity has to do with the nature of the substance. The particle structure, size and molecular composition etc., give each substance a unique spectral signature (i.e. absorption as a function of wavelength). In the case of scattering and reflection by water droplets and aerosols in the atmosphere, the size of the particles strongly influences its effect on radiation (Stull, 2000). The green colour of vegetation, as another example, is a result of the spectral signature of the chlorophyll molecule inside leaves and other parts (Chapin et al., 2002).

Table 2.1 shows albedo values for different surfaces. While the albedo of vegetation materials themselves such as individual leaves (usually around 0.3), may not be extremely low, the collective albedo of a forest can be as low as 0.05 (Oke, 1987). The many layers and surfaces enable multiple reflections within the forest canopy, each surface that is hit by the radiation absorbing a part of it. In this way, the geometry of Earth's surface affect the albedo of the surface system as a whole, in addition to the specific surface properties.

Table 2.1: Albedo of surface examples. Source: Oke, 1987.

Non-urban		Urban		
Surface	**Albedo**	**Surface**	**Material**	**Albedo**
Soils	0.05-0.40	Roads	Asphalt	0.05-0.20
Desert	0.20-0.45	Walls	Concrete	0.10-0.35
Grass	0.16-0.26		Brick	0.20-0.40
Agricultural crops & tundra	0.18-0.25	Roofs	Tar and gravel	0.08-0.18
Deciduous forest	0.15-0.20		Tile	0.10-0.35
Coniferous forest	0.05-0.15	Windows	Glass, zenith angle <40°	0.08
Snow	0.40-0.95		Glass, zenith angle 40-80°	0.09-0.52
Water, small zenith angle	0.03-0.10	Paints	Whitewash	0.50-0.90
Water, large zenith angle	0.10-1.00		Black	0.02-0.15
Mid-latitude cities in snow-free conditions	0.10-0.27	(mean: 0.15)		

Albedo not only varies between substances and surfaces, it also varies for the same surface over the coarse of a day or a year. Diurnally, the albedo can double from early morning to midday and then decrease again toward the evening as a result of the changing angle of incoming solar radiation (Chapin et al., 2002). Kukla and Robinson (1980) showed effectively how albedo varies annually and how the variability is different depending on latitude. Figure 2.3, which depicts the annual range of albedo as a function of latitude, show the effect of seasonal surface alteration on albedo. In high latitudes where snow and ice cover is common in winter, the albedo can increase by more than 0.4 compared to summer albedo.

Albedo can be separated into two extreme cases, so called white-sky and black-sky albedo (e.g. Jin et al., 2005). White-sky albedo is the ratio of reflected to total diffuse radiation. That is, the portion reflected when there is no direct radiation reaching a surface, only that received from previous scatter. Conversely, black-sky albedo is the ratio of reflected to total direct radiation, in the absence of diffuse radiation.

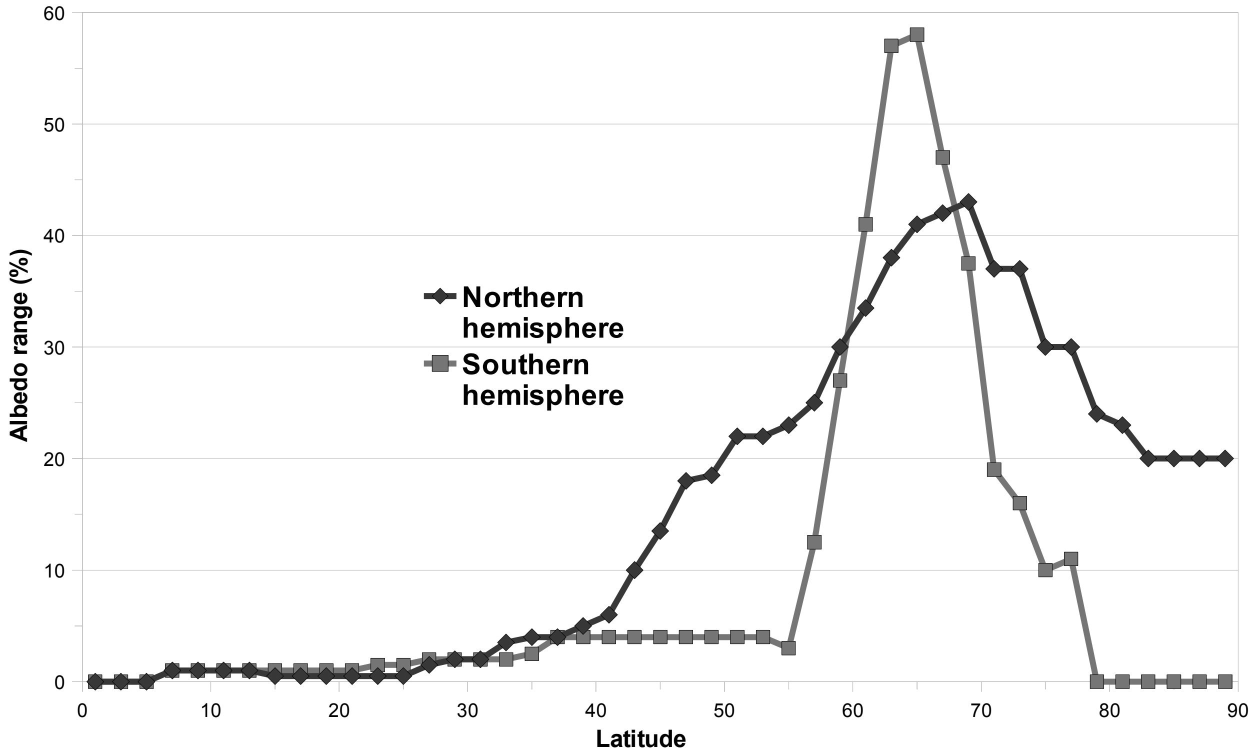

Figure 2.3: Annual range of zonal monthly surface albedo estimates by 2° latitudinal belts. Adapted from Kukla and Robinson (1980). While many assumptions underlie the estimates, the pattern is clear; with higher albedo ranges in latitudes with regular snow and ice cover.

2.3 Urban climate

Urban areas are among the most altered environments on Earth. From cities to suburbs, the materials composing the surface are determined by their use to humans. The surface geometry is related to human concepts such as construction cost, design and tradition. What species of plants and animals live here are accepted by human society, or deemed too expensive to remove. Changes affect quantifiable climatic variables including temperature, wind patterns, precipitation, evapotranspiration and even incoming solar radiation (Oke, 1987; Collier, 2006).

A well known effect experienced in the urban environments is the urban heat island effect (UHI). This effect is the heightened air and surface temperatures (Oke, 1987; Jin et al., 2005) in cities which led the temperature record to be questioned in the global warming context (Solomon et al., 2007). While the issue of the UHI's effect on global temperature records have been resolved, the net effect on global temperatures from urbanization is still not fully understood.

Comparing the global energy balance (Figure 2.2) with that of urban environments, some additional inputs and transfer mechanisms of energy need to be included. Most notable is anthropogenic heat release, which can even be a larger energy source than net radiation in some cities (Oke, 1987). For a given urban building-air volume, the following energy balance may be formulated:

$$Q* + Q_F = Q_H + Q_E + \Delta Q_S + \Delta Q_A \qquad \text{Equation 2.3}$$

where Q* is net radiation, Q_F is anthropogenic heat release, Q_H and Q_E are sensible and latent heat transfer and ΔQ_S is the change in stored energy, seen as changes in the temperature of the air and surfaces within the volume. Lastly, ΔQ_A is the net effect of advection, or horizontal winds, transferring e.g. latent heat in and out through the sides of the volume (Oke, 1987).

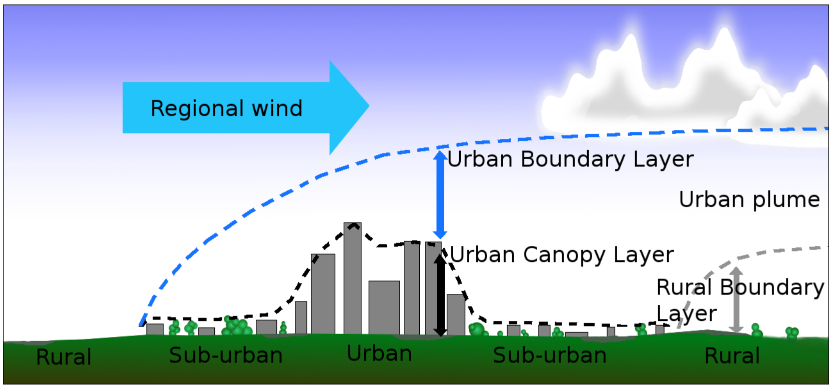

Figure 2.4: Schematic illustration of a two-layer classification of urban atmospheric modifications. After Oke (1987).

The UHI effect is the result of a combination of modifications to the energy balance. Firstly, Q* is modified in several ways; urban areas often have a lowered average albedo and emissivity, altered cloud cover and air pollution. The low albedo and emissivity stems from the introduction of urban materials (Table 2.1) and from the surface geometry. Multiple reflections within spaces between buildings (see Figure 3.2), called urban canyons, are important for the urban energy balance (Harman et al., 2004; Oke, 1987) in the same way as for forest canopies. They trap both incoming short-wave and outgoing long-wave radiation.

Many studies have shown the potential of urban areas to increase rainfall in and downwind of them (e.g. Shepherd et al., 2002; Jauregui and Romales, 1996), but there is still need for more research on the matter, evident by the criticism delivered by Diem et al. (2004) and subsequent replies. Shepherd et al. (2002) outline four primary ways in which urban areas change natural precipitation patterns, namely by:

1. creating, enhancing or modifying <u>meso-scale circulations</u> and thus destabilizing the atmosphere,
2. promoting <u>convergence of air</u> near the surface through high surface roughness,
3. increasing the amount of <u>condensation nuclei</u>, for example from pollution,
4. and by <u>adding moisture</u> to the air from industrial sources.

In their study, rainfall increased on average 5.8 % over the cities examined and 28.4 % over downwind areas, compared to upwind control areas. In another study (Burian and Shepherd, 2005), average warm season rainfall in Houston, Texas, USA, was increased by 25 % between a pre-urban and post-urban period (1940-1958 and 1984-1999). Such an increase in rainfall may correspond to an increased cloud cover, in which case the reflection of incoming short-wave radiation is increased while more long-wave radiation is absorbed and re-radiated to the surface. As mentioned, aerosols and other atmospheric constituents act to scatter and absorb radiation, so the addition of pollution to the atmosphere reduce the short-wave radiation reaching the surface (Solomon et al., 2007).

As for air pollution, in addition to providing condensation nuclei for cloud formation, Arnfield (2003) reports that recent studies largely confirm early assessments, showing up to 10 % reduction in incoming short-wave radiation.

Secondly, Equation 2.3 is modified by the addition of Q_F, which represent all energy inputs to the system from human sources such as the heating of buildings, combustion in car engines or industry. Oke (1987) report annual average magnitudes of Q_F for selected cities between 3 (Singapore in 1972) and 117 W/m^2 (Manhattan in 1967). Although these values are certainly outdated, the point that Q_F varies greatly between cities and has the potential to be a large energy input is still valid. Along with increasing urbanization, there has also been a steady increase in total energy consumption in all sectors for both OECD and non-OECD countries, between 1971 and 2005 (Costantini and Martini, 2010). This suggests that the importance of Q_F is greater today and will continue to increase.

Thirdly, Q_H and Q_E are modified. Urban areas generally have higher ratios of sensible to latent heat transfer (Bowen ratio) due to reduced vegetation and the high fraction of impervious surfaces and drains, which lead to reduced evapotranspiration and increased sensible heat transfer (Jin et al., 2005; Taha, 1997). In spite of potentially increasing rainfall, the net effect in mid-latitude cities seem to be drier air by day and more moist conditions at night (Oke, 1987), but Mayer et al. (2003) report on diverging results.

The wind patterns and turbulence is also altered, affecting ΔQ_A. Generally, the urban canopy increases surface roughness and exerts a frictional drag, reducing wind speeds (Oke, 1987). Under weak synoptic (large-scale) wind conditions, UHI:s tend to generate closed circulations, with surface winds converging toward the city centre where air rises and diverges (Collier, 2006). How changed wind patterns affect energy transfer by advection depends on the conditions at each city.

Lastly, when the previously discussed alterations to Equation 2.3 combine to produce a surplus of energy, due to the thermal properties of building materials this surplus can lead to increased storage of energy as raised temperatures (Oke, 1987).

Jin et al. (2005) found surface skin temperatures, $T_{skin,}$, to be on average 1-5 °C higher in urban areas than in adjacent crop-lands. This analysis was based on all cities, globally, defined by the MODIS (Moderate-Resolution Imaging Spectroradiometer) version 4 land cover data (Schneider et al., 2003). Jin et al. (2005) also showed albedo and the UHI effect to vary with latitude. Trusilova et al. (2008) modelled the effect of urban areas on climate in Europe with a resulting average reduction in diurnal temperature range by 1.26 ± 0.71 °C and 0.73 ± 0.54 °C in summer and winter, respectively. The reduction of diurnal temperature range by urban areas was also observed in the continental U.S. by Kalnay and Cai (2003).

Additional effects experienced in urban environments include a reduced snow cover in winter in high latitudes, a reduced effect of snow cover on albedo and as mentioned for precipitation, urban areas can have considerable effects on downwind areas. The atmospheric properties may be carried along in the urban "plume" (Figure 2.4) bringing thermal and movement characteristics as well as pollution and moisture to areas hundreds of kilometres downwind (Oke, 1987; Shepherd et al., 2002; Collier, 2006)

There are efforts being made to mitigate climatic effects of urbanization. Green roofs (Levallius, 2005), the use of more reflective surfaces (Taha et al., 1988), the planting of vegetation for shading and urban reforestation (Akbari and Konopacki, 2005) are some techniques being researched and applied at different scales.

3 Effects on albedo from urbanization

In a global perspective, urban albedo change needs to be considered in relation to the albedo of the replaced surface. As a rough example; paving a desert road with an initial albedo of, say, 0.40 with a resulting albedo of 0.10 is a considerable reduction. Paving a road through a coniferous forest after clearing a path might, on the other hand, double the albedo from an already low 0.05. In the same way, albedo changes due to urbanization need to consider previous land cover. Since potential natural vegetation as well as agricultural practises (such as choice of crops) vary with climate, the latitude of a city influences its impact on albedo.

Oke (1988) presents results from many studies which report the albedo of urban, suburban and rural areas. From these early (1958-1987) results, based on snow-free tower-, aircraft- or satellite-mounted sensor images, it is reported that urban development consistently reduces the albedo. It is further proposed that 0.14 and 0.15 may be representative average albedo values for urban and sub-urban areas, respectively. As explained, what surfaces urban areas replace is important, but as many urban areas extend into agricultural areas, the global comparison performed by Jin et al. (2005) becomes interesting. According to this comparison of zonal average white-sky broadband albedo between cities and adjacent crop-lands, derived from MODIS satellite images, cities have 2-5 % lower albedo values.

Jin et al. (2005) also note that desert areas near 30 °N have the highest urban albedo values. However, even such areas have lower albedo values in urban than in rural areas, as shown by Frey et al. (2007). In the study by Frey et al. (2007) four satellite images from 2000-2003, from NASA's Advanced Spaceborne Thermal Emission and Reflection radiometer (ASTER), are used to analyse albedo values for Abu Dhabi and Dubai, United Arab Emirates. The two cities are located in desert areas around 25 °N, and both show the albedo of urban land use classes to be 7-8 % lower than rural classes.

3.1 Dark surfaces

The introduction of dark impervious surfaces is a common and important feature of urbanization. Many construction materials have lower albedo than non-urban surfaces (Table 2.1). Offerle et al. (2003) reported for example that roofs and asphalt paving covered 49 % at a site in Chicago and 15 % in a Los Angeles residential area, which has large effects on urban temperatures. Wu and Murray (2003) found impervious surfaces to cover 75 % of a central business district site in Columbus, Ohio, of which a large portion consist of low-albedo surfaces such as asphalt.

In fact, Taha et al. (1988) simulated local effects in Sacramento, California, increasing the surface albedo from 0.25 to 0.40, and showed that the high-albedo case was 2-4 °C cooler in summer and 0.5-1.5 °C cooler in winter. Simply painting a house white (to albedo 0.90) was also found to reduce cooling energy use by 18.9 %, while if at the same time, the albedo of the house's surroundings could be increased to 0.4, the reduction in energy use was as large as 62 % (Taha et al., 1988).

The albedo of surface materials vary with time, as previously discussed regarding diurnal change due to solar angle and seasonal change due to e.g. shedding of leaves in deciduous forest canopies. Many anthropogenic surfaces have an albedo varying on longer time scales as well. Fresh asphalt, with an albedo of 0.05-0.10 slowly increases its albedo with time as the dark binder wears away, revealing lighter rock aggregates (Pomerantz et al., 1997). After years of use, worn asphalt surfaces have an albedo of 0.15-0.20, two to three times higher. Figure 3.1 illustrates differences in the visible-spectrum albedo of three asphalt surfaces of different age. An example of the opposite effect is cement concrete, which after darkening by dirt tends to have an albedo of 0.25-0.30, lowered from 0.35-0.40. The age factor may affect simulation results adversely in cases where albedo parametrization assume young-surface values (e.g. Offerle et al., 2003).

Generally, with dark anthropogenic surfaces often covering large portions of the urban surface and often having lower albedo than the replaced surfaces, as with the case of crop-lands, the effect of dark urban surfaces is to reduce the albedo.

Figure 3.1: Asphalt pavement of different ages in Lund, Sweden. © Mattias Spångmyr

3.2 Surface geometry

As mentioned before, the geometry of the urban environment affects albedo in the same way vegetation canopies do; by enabling multiple reflections of the incoming short-wave radiation. In modelling contexts the concept of urban canyons are frequently used to describe a generalized situation (Arnfield, 2003; Oke, 1987). Figure 3.2 gives a simplified idea of the concept.

Considering the energy balance, the geometry also increases the total surface area, thereby, along with temperature, increasing cities' long-wave radiation (Oke, 1988) and heat storage capability (Offerle et al., 2003). Just like multiple short-wave reflections, inter-element long-wave emission help making urban areas "more effective in capturing energy than horizontal surfaces made of the same materials" (Offerle et al., 2003, p. 1157).

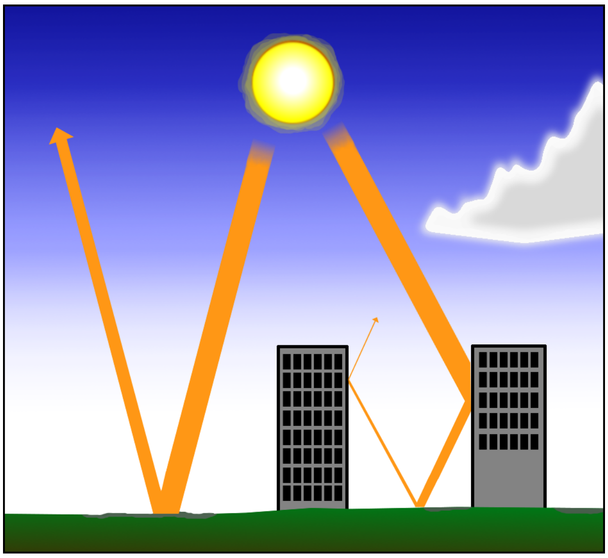

Figure 3.2: Simplified illustration of the multiple reflection effect, showing how less short-wave radiation escapes from urban areas due to the surface geometry.

Generally, as reported by Arnfield (2003), the energy balance of urban canyons is thought to be very dependant on the ratio of the height of walls to the distance between buildings (i.e. the *aspect ratio*, an indicator of urbanization density). Also, generally, deeper canyons (higher buildings) mean lower albedo (Sailor and Fan, 2002). Sailor and Fan (2002) also show the geometry to be important in determining the diurnal variability of urban albedo and that heterogeneity in building heights seem to increase the albedo. With an aspect ratio of 1, Aida and Gotoh (1982) simulated the effect of building geometry two-dimensionally, and found a reduction of the urban albedo by 2-15 % when other variables where held constant and the ratio of building width to the distance between buildings was 0.5-2.

3.3 Increased cloud cover

From studying satellite Advanced Very High Resolution Radiometer (AVHRR) images, Inoue and Kimura (2004) saw a clear increase in low level cloud cover over urban areas of Tokyo, Japan. The processes responsible were hypothesized to be the higher sensible heat flux leading to an increased height of the mixing layer, so that clouds can form at the top of the enlarged thermals (warm rising air volumes), and also convergence and uplift of air due to local circulation induced by the urban to rural temperature contrast.

Romanov (1999) used AVHRR cloud cover data spanning from 1993 to 1996 to infer how the city of Moscow, Russia, affects regional cloud cover. The results showed an increase in the average fractional cloud cover of 4-13 % between March and September and maximum reduction of ca 15 % in December. Average clear sky frequency was reduced by 2.5-15 %, except in December. The reversed effects in winter was attributed to the UHI warming of the atmosphere. As large-scale atmospheric processes determine the height of the top of stratiform clouds the warmer atmosphere over the city may reduce cloud thickness or the fractional cloud cover.

Care should be taken, however, when comparing urban and rural cloud cover differences. Simply looking at the difference as an urban increase from a normal, rural, state may lead to erroneous conclusions since the common local urban circulation may act to suppress rural cloud formation by wide weak down drafts (Inoue and Kimura, 2007).

The effect of clouds on albedo, if one considers the whole Earth-atmosphere system, is to increase it. The albedo of clouds themselves can vary with height and thickness. Thin clouds typically have an albedo of 0.20-0.65, but thick clouds can have as high as 0.70-0.95 (Stull, 2000). Actually, looking at Figure 2.2, (c) shows that 20 % of all short-wave radiation is reflected back into space by clouds. Therefore, urban-induced increases in cloud cover directly increases (c) and increases albedo, albeit not the surface albedo.

The amount of radiation reaching the surface may be reduced by 90 % when under complete overcast of thick, low-level clouds (Oke, 1987). In such cases, the impact of varying surface albedo is similarly reduced. While increased cloud absorption (d in Figure 2.2) does not directly increase the albedo of the Earth-atmosphere system, it further reduces surface albedo impact and modifies the energy balance.

3.4 Air pollution & aerosols

Air pollution practically has the same effects on urban albedo as increased cloud cover. Through increased scattering and absorption in the atmosphere, less short-wave radiation reaches the surface, reducing the impact of urban surface albedo. Alpert and Kishcha (2008), for example, show that urban areas (defined as areas with a population density of ≥400 people per km^2) received ca 8 % less solar radiation than rural areas between 1964-1989. Also, a little more radiation may be reflected back, slightly increasing the albedo of the Earth-atmosphere system as a whole.

3.5 Decreased snow cover in winter in high latitudes

The effect of urban heat islands on the winter snow cover has long been recognized (e.g. Otterman, 1977). The higher UHI temperature and road salting increase melting and much snow is removed by ploughs (Oke, 1987). The remaining snow is often soiled by pollution. Additionally, because the large vertical areas (building walls etc.) are never covered, the albedo stays quite low, much like forest canopies.

Brest (1987) found from satellite observations of Hartford, USA, that while the albedo of urban land cover classes only increased by a few percent from snow cover, other classes experienced much larger increase, up to 40 % for non-tree vegetation categories.

The effects of the UHI could be larger in a near-freezing situation. When rural ponds and other small water surfaces would freeze, an urban water surface might not. While ice surfaces can be covered by snow (and have a much higher albedo), open water surfaces are usually not. Considering water tends to have the lowest albedo and snow the highest, this might compensate for the small area of the surfaces in question.

Overall, the effect of urbanization on snow cover acts to "accentuate the better absorptivity of the urban area" (Oke, 1987, p. 281).

3.6 Changes in surface moisture

Of all the surface albedo values presented in Table 2.1, water has the potentially lowest. Thus it is not very hard to image, and for most people to remember from personal observations, that surfaces have a lower albedo and appear darker when moist.

Although few studies have been done on the relation between albedo and soil moisture content (Wang et al., 2005), those that do seem to agree that moisture reduces the albedo up to at most ~0.08. On a degraded grassland site and a crop-land (corn mixed with sunflower) site with sandy soils in north-eastern China, Liu et al. (2008) measured the diurnal variability of surface albedo with data from 2003-2005. They examined the relation of albedo to soil moisture and solar elevation angle, and found a negative exponential relationship. The albedo falls rapidly with the addition of soil moisture measured 5 cm below the surface, up to approximately 0.15-0.2 m^3/m^3 after which it stabilizes at an average albedo reduction of around 0.05-0.08. A case example of a real rainfall event showed a reduction of ~0.02, which may be an indication of a more usual situation.

Wang et al. (2005) also found a negative exponential relationship between surface albedo and soil moisture, with a reduction of more than 0.07 approaching stabilization. While stabilization occurred at a much higher soil moisture content, >0.6 m^3/m^3, this difference may be explained by the more shallow measurement depth of 3 cm or the different soil characteristics of the semi-desert to desert area of Gaize on the Tibetan plateau.

How moisture affects the albedo of specifically urban surfaces have not been given much attention, which is a large source of uncertainty. Presumably the effects may be as diverse as urban materials.

The prevalence of impervious surfaces and drains in urban areas, as already mentioned, serve to increase run-off and ultimately reduce the amount of water available for evapotranspiration (Haase, 2009). Even though the impervious surfaces themselves do not store any moisture and thus are moist for shorter durations, there is a possibility that the run-off would increase the moisture in adjacent permeable surfaces that do exist in the urban environment (Easton et al., 2007).

Therefore, urbanization could act to reduce the sensitivity of albedo to soil moisture on impervious surfaces, generally keeping the albedo more stable, when other areas would experience a reduction. It could also increase its sensitivity on adjacent permeable surfaces due to increased soil moisture. Adding the potential increase in precipitation over urban areas as previously discussed, the effects of urbanization on soil moisture controls become highly complex and the net effect hard to guess.

3.7 Effects from urbanization outside of urban areas

All modifications of air properties by urban areas, such as the cloud formation and precipitation tendency already discussed or air pollutants, can be transported downwind of its origin in the urban plume (Oke, 1987; see Figure 2.4). Thus, the effects of these urban modifications on albedo can also occur outside of urban areas. Pollutants, for example, which contribute to atmospheric scattering and absorption thereby reducing the incoming short-wave radiation, can act in the same way over downwind rural areas.

Changes in rainfall patterns and moisture regimes have the potential to drive alterations in land cover by changing vegetation, or even biomes (Hély et al., 2006) but albedo changes downwind of urban areas would depend on the response of the ecosystems affected. The simulations by Hély et al. (2006) over a representative transect from African equatorial to desert biomes, suggest that none of the African biomes would change type due to precipitation increasing by 5-20 %. Deciduous and semi-deciduous ecosystems were, on the other hand, shown to be sensitive to reductions in both total precipitation and the number of rainy months. In case of urban expansion generating drier climate or changed seasonality for downwind rural areas, the albedo would tend to increase with forest losses. For semi-arid or arid regions experiencing more precipitation, even though the results from Hély et al. (2006) show no change in biome types, it is likely that the albedo would tend to decrease due to an increase and densification of vegetation.

Apart from physical relationships between urban and rural areas, there may be effects stemming from human urban-rural interactions. DeFries et al. (2010), for example, showed a positive correlation between urban growth and deforestation in the humid tropics. Causal relationships were not proven, but if urbanization is a potential driver of deforestation, with deforestation acting to increase surface albedo, this effect may act to offset increases of the albedo within urban areas.

4 Urbanization

When investigating urbanization, a need of clear definitions arise. What should be considered urban? There are a myriad of characteristics that can be used, most notably the presence of buildings and structures or people, but regardless of which are chosen there are still lines to be drawn and assumptions to be made. In the United Nations publication *The State of World Population 2007* (UNFPA, 2007), the definition of urban is based on definitions by national statistical agencies. Urbanization on the other hand, is defined as an increase in the proportion of the population which live in urban settlements.

The definition of urban areas, called localities, from the Swedish national statistical agency provide an example which is representative for the Nordic countries (Statistics Sweden, 2006). A locality is, in short, a group of buildings with at least 200 inhabitants where the distance between buildings is at most 200 m. In an international context 200 people is a very small settlement, but the Nordic countries are generally sparsely populated. North America and Europe commonly use a lower limit of one or two thousand (Statistics Sweden, 2006).

While the Nordic definition is, at least in part, related to building density, the common focus on demographics undermines the validity in using these types of definitions to derive physical properties of urban areas. But as discussed by Oke (1987) the anthropogenic heat release (per area unit) is related to population density and energy use per capita. Energy use per capita, in turn, is dependent on many factors e.g. economy. Thus, demographics may provide a useful contribution.

One example of a definition based on physical properties is a threshold on the percentage of observations in which a specific pixel is lighted at night, which was used by e.g. Imhoff et al. (1997). That is, in satellite images taken at night, the pixels which were lighted by for example street lights were recorded and then all pixels which were lighted in enough of the images were classified as urban. This technique excludes ephemeral light sources which are not commonly of urban origin. The threshold, however, was set to >88 % using US census data as reference, again relying on demographics. Using stable night light sources to delineate urban areas may also be biased for several reasons, e.g. the varying degrees of electricity use between regions with different economic or cultural characteristics.

Another physically based approach to define urban areas is presented by Griffiths et al. (2010). Using a support vector machine (SVM) classifier (explained by Vapnik, 1999, compared to other machine learning methods by Pal and Mather, 2005), with multi-spectral and multi-temporal input data from Landsat TM satellite and Saturated Aperture Radar images, urban areas could be classified according to their spectral and seasonal characteristics. Even if the SVM classifier used physical characteristics to discern if a surface is urban, it was trained using reference data which was classified manually and the rules and methods for creating the reference data were not elaborated on.

Schneider et al. (2009) use a compellingly simple physical definition, where urban area is defined as an area (i.e. a pixel) which is dominated (at least 50 % covered) by human-constructed surfaces not including vegetation. With such a description, all physical alterations of the environment normally associated with urbanization can be captured. However, they also limit urban areas to "contiguous patches of built-up land greater than 1 km^2" (Schneider et al., 2009, p. 3) and while the somewhat arbitrary limit of 50 % may make sense on a global scale, it still introduces uncertainty related to image resolution and the density of urban surfaces in a specific area.

As of yet there is no definition of urban area which has acquired general acceptance in the global mapping community (Potere et al., 2009) and this may contribute greatly to discrepancies in the results from different methods of mapping urban areas.

4.1 Extent of urban areas

With the problems previously discussed in finding a suitable definition of urban areas, it is natural that the actual extent of urban areas is still very much unclear. Data from the DCW (Danko, 1992) has been widely used for a long time and still is (Potere et al., 2009). As mentioned in the introduction, this map is outdated. Table 4.1 shows samples of estimations of Earth's land surface area, including sources using the DCW, population statistics and satellite imagery. There are many other global maps showing urban areas, and on top of this, there are countless maps showing urban areas on a regional or local scale.

Table 4.1: Some estimations of the global extent of urban areas as percentage of land surface area. Source: Potere et al., 2009. Percentage based on a total land surface area of 148.94 million km ^{2} (CIA, 2010).

Source	**Input data**	**Extent**
Loveland et al., 2000	The Digital Chart of the World (DCW), which is based on maps, charts and city gazetteers from 1950-1979.	0.185%
Bartholome and Belward, 2005	Coarse (1 km) resolution satellite images from 2000 and other, e.g. “night lights”, images supporting in classification.	0.207%
Goldewijk, 2005	Population statistics until 2000 and other global land cover maps.	0.357%
Schneider et al., 2009	Moderate (500 m) resolution satellite images from 2001.	0.441%
CIESIN, 2004	“Night lights” images from 1994-1995, urban mask from DCW, and ancillary data sets such as charts.	2.371%

There are as many methods of delineating urban areas as there are maps. The census approach, using demographics as explained earlier is a common non-physical method. The physical approaches use either aerial photography (e.g. Niedzwiedz and Batie, 1984) or satellite remote sensing. With the advent of more advanced satellite instruments, data is becoming increasingly available to generate maps from surface spectral characteristics.

A common data source for mapping urban areas comes from the Moderate-Resolution Imaging Spectroradiometer (MODIS) carried on NASA's *Terra* and *Aqua* satellites (Jin and Shepherd, 2005). Others are for example instruments on the Landsat and ERS satellites, including synthetic aperture radar (SAR), which have a spatial resolution of 30 m. These many sources of data allowed for example Griffiths et al. (2010) to use multi-temporal and multi-spectral datasets to delineate the megacity of Dhaka, capital of Bangladesh. An important point made is that there is a trade-off between spatial resolution and temporal homogeneity, that is, while higher resolutions enable more detailed mapping, high resolution tend to be accompanied by narrower sensor paths. Different parts of large cities being scanned at different times may complicate analysis efforts.

The method used by Jin et al. (2005) to identify the extent of three large cities in the U.S. show that changes in surface albedo can be a useful indicator of urbanization. In their study, an urban index (UI) is created by multiplying skin temperature, $T_{skin,}$ with albedo, α (or rather: $UI = (1 - \alpha) \cdot T_{skin}$) giving a map of continuous values of UI, to which a threshold can be applied to delineate urban land cover. This combination allowed for example Washington, New York and Beijing to be more easily delineated, but for Phoenix, chosen as a city in a hot and dry environment, lower temperatures due to more vegetation in the city made it somewhat undistinguishable.

Trusilova et al. (2008) created a new map of urban areas of Europe by combining several different techniques, including for example night light delineation, MODIS and GLCC (Loveland et al., 2000). The Coordinated Information on the European Environment from the Europe Environment Agency (EUCORINE) land cover database was used as a reference. The resulting map underestimated European urban areas by approximately 10 % (overestimation of large cities and omission of smaller towns and villages) but still showed 2.8 % of European land area as urban. For mid-latitude land in general, Jin et al. (2005) state that 2 % is covered by urban or industrial development.

In an effort to evaluate the current variety of global urban mapping projects, Potere et al. (2009) provide an accuracy assessment rating maps in terms of city omissions, city sizes and per pixel agreement between the maps and reference cities verified with Google Earth imagery. According to this assessment, MOD500 (based on MODIS Collection 5; Schneider et al., 2009) is the best map of urban areas to date, and it depicts 657 000 km^2 as urban, which is approximately 0.44 % of Earth's land surface (Table 4.1).

4.2 Location and rate of urbanization

Most cities are located between 30 and 65 °N (Jin et al., 2005), which, considering this area spans roughly from north Africa to the Arctic circle, is not surprising. Figure 4.1 illustrates where the urban areas of the world are located, by showing the stable lights sensed by satellites at night. Stable lights refer to lights which are persistent through time; excluding fires and other intermittent light sources. The data, collected from NOAA's National Geophysical Data Center, have also been rid of background noise, moonlight, aurora effects and clouds. While the image shows stable light sources at Earth's surface and not actual urban areas, it can be used as an approximation (e.g. Imhoff et al., 1997) to illustrate where most urbanization is located. Because of the image size of Figure 4.1, the contrast has been increased for illustrative purposes.

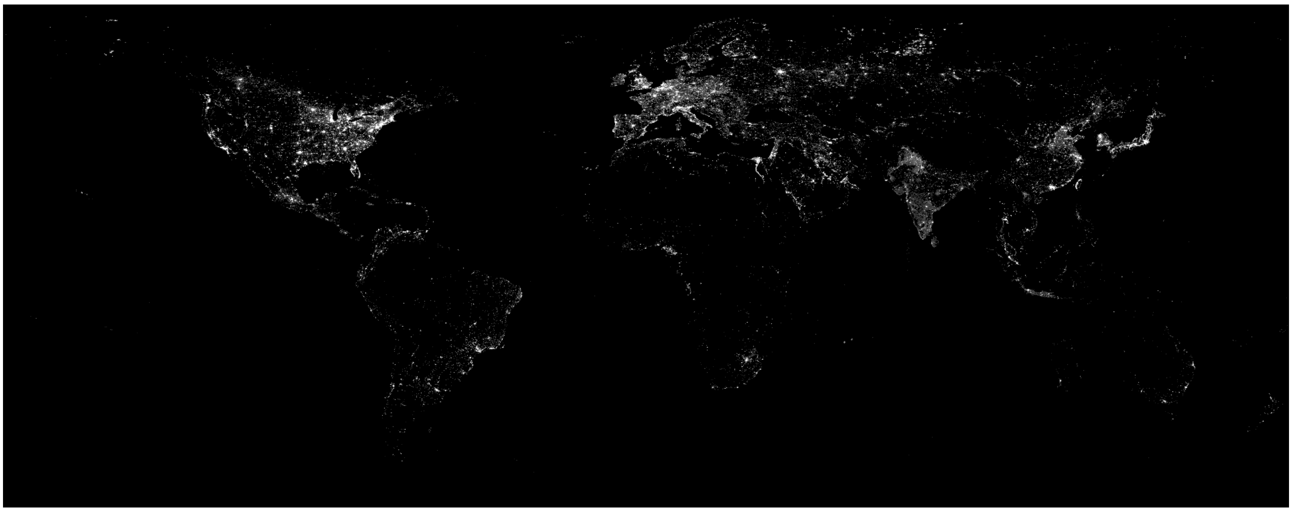

Figure 4.1: Stable night lights on Earth's surface as seen from satellite in 2009. Original image and data processing by NOAA's National Geophysical Data Center. DMSP data collected by US Air Force Weather Agency. (NOAA/NGDC, 2009)

From simple visual interpretation of Figure 4.1 it is obvious that North America, Europe, East Asia and India are the most urbanized areas today.

UNFPA (2007) describe two "waves" of demographic, economic and urban transitions. The first wave, occurring in Europe and North America from roughly 1750 to 1950, saw an increase in the regional ratio of urban to rural population from 10 to 52 %. While the developed world, most of which experienced the first wave, will only see modest urban growth the next few decades, the developing world is currently living in the second wave of urbanization. Between 1950 and 2030, their ratio of urban to rural population will see a slightly lower increase than the first wave induced for the developed countries; from 18 to 56 %. The second wave, however, is on a completely different scale. While urban populations increased by 408 million in the first wave, in the second wave they will have increased by nearly 3.6 billion. All future population growth is expected to go to cities, rural population will actually decrease some 28 million. As mentioned in the introduction, the urban population is projected to amount to 4.9 billion people, or 80 % of the total population, by 2030. In 2008 it was 3.3 billion.

Most urban population growth will occur in Asia and Africa, who will double their urban populations, and in South America and the Caribbean which will grow more slowly (UNFPA, 2007). The developed world will only show modest urban growth. While the scale of growth is very high, the rate of growth is declining and will continue to decline in all regions of the world, from ca 1.6-5.0 % per year in the 1960:s to ca 0.2-2.8 % near the end of the 2020:s (UNFPA, 2007).

There is a high degree of uncertainty in how future urban population growth will affect the expansion of urban areas. An assessment on the development of urban areas in the conterminous U.S. (excludes Hawaii and Alaska) by Brown et al. (2005) show large increases in urban area that is less densely populated. In addition to the uncertainty of the relation between population and urban area growth, the climatic effects of a certain area increase may be more than expected. Trusilova et al. (2009) reported on simulations showing that the area portion significantly affected by temperature range modification more than doubled when urban area increased by only 40 %.

5 Global impact of urban albedo change

The 4.9 billion people that will live in urban areas by 2030 (UNFPA, 2007) will clearly experience the local and regional effects outlined in the previous chapters, but how will any global effects manifest themselves?

Evident from previous chapters, many processes counteract one another and the global net result, while complex to determine, may mask large regional differences (see for example Barnes and Roy, 2008). A frequently used measurement for comparisons between anthropogenic climate drivers is the concept of radiative forcing, which describes the effects in terms of change of energy flux per surface area (W/m^2), whether ground surface or imagined surfaces such as the top of the atmosphere.

In addition to directly influencing Earth's energy balance, the urban albedo modification, as a factor in the UHI effect, contributes to increasing the consumption of energy for summer cooling and decreasing it for winter heating. This change in energy needs may be translated into equivalent CO_2 emissions from energy production, another measurement facilitating comparisons.

5.1 Radiative forcing of urban albedo change

Radiative forcing, as outlined by the National Research Council (NRC, 2005), is basically any climate forcing that affects the radiation balance. Pielke et al. (2002) hypothesize that the radiative forcing of global surface albedo changes due to land cover changes may be comparable to the forcing of solar variability, anthropogenic aerosols or to some of the green-house gases. They also suggest that more regional metrics need to be considered because of the potentially large offsetting regional changes, which would be masked by global averages.

In the latest IPCC report (AR4; Solomon et al., 2007), the radiative forcing of surface albedo change in general, was suggested to be between 0 and –0.4 W/m^2 and the level scientific understanding was raised from low to medium-low.

Myhre and Myhre (2003) estimated the global radiative forcing of land cover change by comparing current land cover with potential natural vegetation. Depending on the datasets used and different albedo values assigned to e.g. crop-lands, the radiative forcing ranged from –0.6 to 0.5 W/m^2 with the negative values being more likely. In their analysis, however, they stated that they could “at present, ignore changes in surface albedo caused by urbanization” (Myhre and Myhre, 2003, p. 1518), judging by the much smaller extent of urban areas compared to crop-lands. The dataset, on which their estimation of urban areas is based, use the Digital Chart of the World for mapping urban areas, which, as can be seen in Table 4.1, is likely to underestimate their extent severely.

Barnes and Roy (2008) use the difference between absorbed incoming short-wave radiation at the surface in 1973 and that absorbed in 2000, as the definition of radiative forcing from land use change. They estimate the radiative forcing for 43 % of the conterminous United States to be approximately 0.012 W/m^2 (a warming), but on the eco-region scale, it varied from –0.247 W/m^2 to +0.284 W/m^2. Conversion from the *agriculture* to the *developed* land use class played a major role in the warming effect.

In this analysis, the albedo, derived from MODIS satellite data, was the key parameter used to determine the radiative forcing of land use change, but mean monthly incoming surface solar radiation was also included to counter the effects of solar variability. While effects such as the introduction of dark surfaces can be captured, other urban properties such as increased cloud cover and reduced snow cover cannot, due to the use of only snow-free, non-cloudy data.

While exclusion of feedbacks was one of the reasons for raising the level of scientific understanding in the AR4, the exclusion of these effects may skew the results. In cases where snow cover is common in winter, the difference between the developed land use class and the replaced class may be underestimated, or overestimated where urban areas increase local cloud cover.

Some studies have been conducted, trying to interpret changes in surface albedo, specifically for urban areas, into radiative forcing. Akbari et al. (2009), for example, estimate a global radiative forcing of 1.27 W/m^2 per 0.01 change in surface albedo values, globally (negative forcing for increases in albedo and vice versa). While this study makes many questionable, albeit necessary, assumptions such as urban surfaces covering 1 % of the global land area, the global mean radiative forcing exerted from a 0.01 decrease in the albedo of 1 % of the land surface, would be 0.0037 W/m^2 ($0.01 \cdot 0.29 \cdot 1.27$ W/m^2, where 0.29 is the fraction of Earth's surface covered by land; CIA, 2010).

The principal result by Akbari et al. (2009) is supported by Menon et al. (2010) who report a radiative forcing of –1.63 W/m^2 per 0.01 increase in Earth's surface albedo. The difference may be explained by methodological differences such as Menon et al. (2010) only using boreal summer months for example. By using a larger estimate of urban areas (see the largest estimate in Table 4.1) their corresponding value of global mean radiative forcing exerted from a 0.01 decrease in the albedo of urban areas would be 0.011 W/m^2 ($0.02371 \cdot 0.29 \cdot 1.63$ W/m^2). While both these estimations differ considerably, they are still notably smaller than the global estimation of –0.2 ±0.2 W/m^2 given by the IPCC for land cover change in general since pre-industrial times (Solomon et al., 2007).

Both studies report that an increase of global urban albedo values by 0.01 would offset more than a year's worth of global CO_2 emissions, or 44 (Akbari et al., 2009) to 57 (Menon et al., 2010) Gt CO_2.

Davin et al. (2007) used climate models to simulate land cover changes from 1860 to 1992 to 2100 and concluded that radiative forcing of anthropogenic land cover change is a weaker control on climate than a CO_2 radiative forcing of the same magnitude. Their simulations support the net cooling from historical land cover changes and predicts even greater future cooling, but urbanization is not explicitly included. Since the importance of urbanization is predicted to increase, so will this source of error.

5.2 Impacts on anthropogenic energy consumption

In contributing to the UHI effect, urban albedo change modify human energy needs for cooling or heating buildings. That increasing the albedo of buildings can be an effective measure to reduce energy demand, as noted in an earlier chapter, is supported by Akbari and Konopacki (2005). They estimate an annual electricity savings potential of 1294 GWh in Houston, Texas, from the combined use of reflective roofs, tree shading, and urban reforestation.

As also mentioned, Taha et al. (1988) simulated a potential energy use reduction of 62 % from increasing the albedo of a prototype house in Sacramento, California, and its surroundings. The direct energy reduction in this case was 46 kWh over the course of 4 days in July.

Sims et al. (2003) present average carbon emission per kWh of generated electricity from different production sectors. With today's technological level, roughly 150 g C per kWh is released in production, which translates to 550 g of CO_2 assuming it is the only green-house gas emitted. With the Houston electricity savings potential of 1294 GWh per year, this would mean 0.711 Mt (mega tonnes) CO_2 less per year emitted from Houston only, or 1950 t CO_2 per day. For the single prototype house the reduction could be 6.3 kg CO_2 per day.

6 Discussion

Much of the research reviewed in this report has yet to mature, as it often relies on assumptions about albedo or the extent of specific land use classes, which in turn has many problems with, for example, spatial and temporal resolutions.

Albedo has spatial, temporal and spectral variability which makes it hard to integrate into a simple parameter. Because satellites are not geostationary, because they have a limited spatial and spectral resolution and because they have a limited spectral range, satellite measurements of albedo are approximate, at best. They can also not take albedo measurements from all angles.

Many studies also rely on the use of snow-free or non-cloudy data, or both (e.g. Barnes and Roy, 2008). This explicitly leaves out the potentially strong effects of urbanization-induced cloud cover and reduced snow cover on albedo.

The analysis of the extent of urban areas is hampered by the problems with defining the land use class. The diversity in the physical appearance of densely populated areas and their inherent heterogeneity makes global assessments difficult. Current approximations may not be accurate, like for example the one cited in the AR4 by the IPCC, and will very probably soon be outdated by current rates of urbanization.

While both city geometry and the UHI increase in temperatures also increase long-wave radiation because long-wave radiation is better absorbed by the atmosphere, this may not fully compensate for the absorption increase. Research supports the idea that net radiation (Q*) varies little between urban and rural areas, due to offsetting radiation effects (Arnfield, 2003). However, even if Q* varies little at the surface, a decrease in the portion of short-wave radiation reflected to space (c in Figure 2.2) inevitably cause a global, albeit small, warming effect.

Much of the urban heat island effect can be alleviated by increasing the albedo of building materials and by increasing vegetation. Green roofs are an example of a mitigation strategy with much potential, and considering the low cost of implementation in relation to the potential future costs to society of climate changes I feel that all mitigation strategies mentioned in this report deserve not only more research, but probably wide spread implementation in the near future. This is especially interesting considering the energy savings potential estimated in many studies.

Even though there is much research being conducted in the fields of urban mapping, urban climatic effects and global land cover change, I feel there is still much to be done, which is also reflected in the low level of scientific understanding stated by the IPCC. Specifically, the connection between urban climate modifications and global climate response has received little attention, which was one of the reasons I chose this subject for the report. There is no doubt that urban areas strongly affect climate on local and regional scales, and the ability of regional climate effects to propagate and affect other regions is a well known phenomenon, for example in the study of ENSO or other similar processes.

6.1 Conclusions

The effects of urbanization on Earth's albedo are as diverse and heterogeneous as the urban fabric itself. While some aspects of urbanization tend to reduce the albedo compared to non-urban areas (e.g. dark surfaces, surface geometry, reduced snow cover in winter), some may increase it (e.g. increased cloud cover and air pollution) and some are still too poorly understood to make an assumption (e.g. modified surface moisture regime and downwind effects). It was found from satellite comparison between urban and crop-land albedo, that urban surfaces have 2-5 % lower albedo.

The extent of urban areas is unclear and depend largely on the definition of urban areas, but the best estimate tend toward 0.44 % of the land surface in 2001, well over twice that cited by the IPCC in the AR4, which bases its estimation on up to 60 years old data.

Increasing the albedo of all urban areas, globally, by 0.01 is calculated to exert a negative radiative forcing of less than 0.01 W/m^2, or more than 20 times smaller than the negative radiative forcing estimated by the IPCC for all land cover change since pre-industrial times. Further expansion of urban areas into high-albedo areas such as crop-lands or deserts would result in increased warming. However small, this warming reduces the net cooling effect of land cover change.

Therefore, the final and most important conclusion is that unless urbanization is explicitly included in land use change research and acknowledged by the IPCC, and the climatic effects of urbanization are further studied, the official level of scientific understanding should be returned to low.

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