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Heather Glickman-Eliezer

Graduate Center, City University of New York

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AEROSOL AND URBAN HEAT ISLAND EFFECTS ON WARM CLOUD-TOP SHORTWAVE INFRARED REFLECTANCE AND VISIBLE ALBEDO

by

HEATHER GLICKMAN-ELIEZER

A dissertation submitted to the Graduate Faculty in Earth and Environmental Science in partial fulfillment of the requirements for the degree of Doctor of Philosophy, The City University of New York

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This manuscript has been read and accepted for the Graduate Faculty in Earth and Environmental Science in satisfaction of the dissertation requirement for the degree of Doctor of Philosophy.

Dr. Reza Khanbilvardi

Date

Chair of Examining Committee

Dr. Cindi Katz

Date

Executive Officer

Dr. Reza Khanbilvardi  (City College)

Dr. Robert Rabin  (NOAA/National Severe Storms Laboratory)

Dr. Juliana Maantay  (Lehman College)

Dr. Nir Krakauer  (City College)  
Supervisory Committee

THE CITY UNIVERSITY OF NEW YORK
ABSTRACT

AEROSOL AND URBAN HEAT ISLAND INTERACTIONS WITH WARM CLOUD-Top
SHORTWAVE INFRARED REFLECTANCE AND VISIBLE ALBEDO

By
Heather Glickman-Eliezer

Adviser: Professor Reza Khanbilvardi

This study uses ten years of Geostationary Operational Environmental Satellite (GOES) data (1999-2009) to assess urban effects of anthropogenic aerosols and Urban Heat Island (UHI) on cloud-top radiative properties of optically thick clouds. GOES images at noon local time of channels 1, 2, and 4 are used to calculate visible albedo, and shortwave infrared reflectance at 3.9-µm. Cloud-top particle radius is inversely related with SWIRR. Albedo and cloud fraction are measured for cold and warm clouds, and SWIRR is used as a particle-radius proxy for warm clouds only. AERONET fine-mode particle retrievals (fAOD) from the CCNY station are used for the aerosol measure from 2002-2008. A daytime UHI is calculated using ground weather stations in and around New York City. Days are divided into high, medium, and low Aerosol, and UHI. The urban location at Central Park, in NYC is compared with two rural control locations 130 km to the north and south. Climatologies one hour downwind of these three starting points are created using ground station wind data for areas of 5, 10, 20, and 35-km radius.

In overall aerosol results, warm cloud SWIRR is variable but decreases with increasing aerosol. The decrease is significantly greater for areas downwind of the rural locations compared with urban (NYC). This urban-rural difference increases significantly with
increasing fAOD for all seasons but winter. Cloud fraction for warm clouds varies inversely with SWIRR for seasons other than winter. Urban-rural differences in cloud fraction are increasingly negative with increasing urban fAOD, suggesting an urban aerosol effect of cloud dissipation. This could be explained by cloud sedimentation and evaporation-entrainment effects, or a semi-direct effect. In a visual inspection of a large number of mapped climatologies, SWIRR values of pixels are inversely correlated with albedo.

Summer season warm clouds have significant positive urban-rural SWIRR differences at all aerosol levels, and differences increase with higher fAOD and at smaller downwind-area radii. Spring, summer and fall SWIRR differences increase with fAOD, and have an inverse relationship with cloud fraction urban-rural differences, with the only exception of the spring high aerosol category. Cold-cloud albedo in the summer season increases with increasing CCNY aerosol level, with the rural values becoming increasingly higher than urban.

In UHI outcomes, lower urban-rural albedo and cloud fraction with higher UHI is expected, and is found in larger downwind areas. Urban cloud enhancement may be present in cases of deviations, or in trends toward a deviation from this pattern in data drawn more exclusively from the urban and rural starting points. High UHI urban-rural differences are higher than expected at the smaller downwind areas for summer and fall albedo, and for winter and fall cloud fraction. Higher UHI days have higher warm-cloud SWIRR in both urban and rural areas in spring and fall. SWIRR urban-rural differences are high for all of
the summer season, and increase with UHI in spring and fall. An urban effect may be at play in winter cloud fraction outcomes, in which the urban-rural difference is consistently positive. Urban cloud fraction is generally lower downwind of the NYC location for other seasons.

Low, optically thick clouds are fewer, and/or are fewer downwind of more narrowly defined study areas in urban compared with rural locations across both datasets for seasons other than winter. These clouds have a negative radiative forcing; that is, their net effect is cooling, with implications for the global radiative balance. Short-term warming can be felt directly by urban and suburban residents. Refinements in predicted cloud cover is important to urban planners both regarding peak energy loads and water resources. Comparisons of these outcomes with climate prediction models could be useful in their incorporation of urban effects.
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<td>The Aerosol Robotic Network</td>
</tr>
<tr>
<td>BC</td>
<td>Black Carbon</td>
</tr>
<tr>
<td>CCN</td>
<td>Cloud Condensation Nuclei</td>
</tr>
<tr>
<td>ETA</td>
<td>Eta Data Assimilation</td>
</tr>
<tr>
<td>fAOD</td>
<td>Fine-mode Aerosol Optical Depth using level 1.5 AERONET data at 500-nm, also denoted as $\tau_a$</td>
</tr>
<tr>
<td>GSOD</td>
<td>Global Summary of Day</td>
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<td>IN</td>
<td>Ice Nuclei</td>
</tr>
<tr>
<td>NCDC</td>
<td>National Climate Data Center</td>
</tr>
<tr>
<td>NDVI</td>
<td>Normalized difference vegetation index</td>
</tr>
<tr>
<td>NESDIS</td>
<td>United States National Environmental Satellite, Data, and Information Service</td>
</tr>
<tr>
<td>NOAA</td>
<td>National Oceanic and Atmospheric Administration</td>
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NOHRC  NOAA’s National Operational Hydrologic Remote Sensing Center

PBL    Planetary Boundary Layer

SWIR   Shortwave infrared, in this case at the 3.9-µm central wavelength

SWIRR  Shortwave infrared reflectance, in this case at the 3.9-µm central wavelength

TIR    Thermal Infrared

UCL    Urban Canopy Layer

UHI    Urban Heat Island

UTC    Coordinated Universal Time
1 INTRODUCTION

The Urban Heat Island (UHI) refers to elevated temperatures found in urbanized landscapes. Urban areas are less vegetated, and undergo less evaporative cooling than their suburban and rural counterparts. Urban materials of asphalt and concrete absorb sensible heat during the day, and release it later in the day and night. Urban regions are also a source of anthropogenic heat. Urban heat island affects the climate by modifying boundary layer processes (Shepherd, 2005).

Cities are both a source and recipient of aerosols both natural and anthropogenic, and their effects on clouds and climate interact in complex ways with aerosols. The anthropogenic fine-mode aerosols chosen for this study in some cases enhance, and in others dissipate clouds. The distribution and the net radiative impact of aerosols on clouds is not well understood, and remains one of the biggest sources of uncertainty in global climate models.

Satellite imagery of the Northeast coast centered around New York City over a ten-year period provides a robust climatological dataset of cloud properties. This study aims to improve our understanding of the effects of two urban variables affecting clouds: those of aerosols and UHI. The development, prevalence, and movement of clouds are affected by both aerosols and UHI. Cloud properties are measured with satellite images one hour downwind of urban and rural starting points. Residents of the metropolitan region directly experience effects on solar radiation on the ground. These outcomes also provide
empirical data relevant to climate forecasting.

Surface temperature, moisture, winds, as well as aerosols play a part in the thermodynamic conditions of the planetary-boundary layer (PBL). Land-surface coupling with weather conditions in the well-mixed PBL, the lowest layer of the atmosphere, can influence the development or dissipation of low clouds. These low clouds, with warm cloud-top temperatures are the focus of this study.

Cloud albedo is a measure of the percent of reflectance of visible-spectrum radiation back into space. This radiative property influences surface temperatures; high albedo is felt on the ground as shading of sunlight. Shortwave infrared reflectance (SWIRR) is used as a proxy for cloud particle radius. This cloud microphysical property has implications for the cloud’s future development, and can also alter the amount of radiation blocked by clouds of equivalent water content. Liquid particle radius can also affect the likelihood of precipitation (Rosenfeld and Gutman 1994). Last, cloud fraction is a measure of effects on percent cover of low, thick cloud cover downwind of the study areas. Aerosol effects on cloud droplet-size can be assessed through the SWIRR measure, and can interact with and influence cloud fraction and albedo. Boundary-layer UHI levels can be explained by cloud fraction and albedo properties of clouds, but can also influence or initiate weather processes that feed back into changes in those same measures.
Objectives

In all cases cloud properties one hour downwind are explored for short-term urban effects. The cloud properties studies are cloud albedo, fraction, and SWIRR, and their urban-rural differences.

The objectives of this work are to explore the following questions:

- What cloud-property differences are evident under different levels of fine-mode aerosol in urban and rural areas? Which theories of aerosol-cloud interactions best explain New York City metropolitan region cloud-aerosol and urban-rural outcomes? How are the outcomes seasonal?

- What is the relationship between the NYC daytime UHI magnitude and cloud properties, and their urban-rural differences? Is there evidence for urban heating weather effects, such as of convergence or urban-enhanced thermal convective cells? If so, what are their seasonal variations?

- While urban effects on precipitation has received much attention in prior studies, effects on cloudiness has not. What can urban-rural differences tell us about the balance of urban effects under different conditions on cloud fraction, albedo and SWIRR?
**Motivation**

Increased urbanization is a continuing global trend, and a source of uncertainty in climate prediction. Urban and regional residents themselves are affected both by cloud changes and by our understanding of urban weather. One half of the world’s population lives in a city; one fourth in a mega-city (United Nations Department of Economic and Social Affairs & Population Division, 2014). Urbanization is accompanied by changes in land use and UHI, affecting cloud prevalence and dynamics over and downwind of urban areas.

Anthropogenic air pollutants have impacts on microphysical and radiative properties of clouds. The role of these urban impacts on climate, however, is not well understood or adequately represented in models. Clouds and aerosols also contribute the largest uncertainty to estimates and interpretations of the Earth’s changing energy budget (Boucher et al, 2013). Climatological studies can help to resolve gaps in our understanding of urban climate, and contribute to improved predictions and risk assessments which form the basis of planning and policy.

Regionally, clouds affect freshwater resources, agriculture, and energy use. In their interactions with urban heating, clouds also affect human health through changes in temperatures and pollution levels. The short-term time scale of effects in this study often take place within the metropolitan region. Changes in clouds and solar flux thus affect quality of life for residents of large metropolitan regions in coastal temperate zones, and are also important to urban planners.
This study will explore one region over the course of ten years, creating a robust data set
with which to explore the relationships between UHI, aerosols, cloud albedo, cloud fraction,
and shortwave-infrared reflectance as a proxy for cloud particle radius. Observational
studies have found urban heating and aerosol effects to manifest themselves very
differently in different regions. As such, there is also a need for regional analyses like this
one, which focuses on the NYC-metropolitan region and two less built-up control areas.

The planetary boundary layer (PBL) is lowest portion of the atmosphere (from surface to
about 1 to 2 km high), and is where most atmospheric dynamics take place. The PBL is the
main source of heat, water vapor, and aerosols in the upper atmosphere, and the main
atmospheric sink of momentum and kinetic energy. Boundary-layer clouds cover 23-30%
of earth’s surface, with a lower fraction over land than water. Land use-land cover changes
affect the Earth's physical surface properties, and impose a radiative forcing on the climate
system (Sagan et al. 1979; Hansen et al., 1998; Intergovernmental Panel on Climate Change,
2007). The emphasis of this study will be on warm PBL clouds composed primarily of
liquid droplets, though cold clouds are also included. Improving our understanding of
changes in low-cloud cover and radiative effects is crucial to global climate models.

A secondary component of this study is the inclusion of albedo and cloud fraction for cold
clouds, where microphysical processes are very poorly understood (Boucher 2013). This
receives less attention in part because SWIRR is not a good proxy for ice particle radius,
and thus is not used for cold clouds.
This climatological study provides a statistical analysis of cloud properties based on satellite and ground based aerosol and urban heating data. Its outcomes can be compared against modeled predictions, and thus contribute our understanding of these processes.
2 BACKGROUND AND LITERATURE REVIEW

2.1 Aerosols and cloud interactions

Aerosols are very small suspended particles in the atmosphere such as sulfate, mineral dust, carbonaceous particles, other organic aerosols and sea salt. Natural aerosols include those from sea spray, volcanoes (sulphates), dust storms, forest and grassland fires, and living vegetation. Anthropogenic aerosols include those resulting from the burning of fossil fuels and the alteration of natural surface cover. Anthropogenic aerosols are typically in the submicrometer- to micrometer-size range and are composed of numerous inorganic and organic species (Haywood et al, 2000; Penner et al., 2001). Aerosols fall under four broad categories: sulfates, carbonaceous aerosols, which include black carbon (BC) and organic carbon (OC), dust, and sea salt. For more about aerosol sources and measurements see Appendix B.

The AOD is the vertical integral of the aerosol concentration weighted with the effective cross-sectional area of the particles intercepting (by scattering and absorption) the solar radiation. The globally and annually averaged value of AOD (at 0.55 µm) is about 0.12 (±0.04). Most aerosols form a thin haze in the lower atmosphere (troposphere), and aerosol lifetimes are only a week or less (Kaufman et al., 1997). This is in contrast with long-lived greenhouse gases (GHGs), which are distributed uniformly over the globe. Tropospheric
aerosols are thus relatively localized in their distribution, though they can be transported many miles with a moving airmass, and if they reach the stratosphere. Long-range atmospheric transport of elevated aerosols transforms haze into a regional-scale layer (Penner et al., 2001). The effect of stratospheric aerosols is also dependent on prevailing wind patterns but their effects are far broader geographically. Aerosols thus have substantial spatial and temporal variations, with peak concentrations near the source. Nucleation in liquid cloud drops incorporates about 90% of the initial aerosol particle mass (Flossman, 2010). Ice particles, by comparison, take up very little aerosol.

Anthropogenic aerosols are generally smaller in size and more absorbing than the natural aerosol (Myhre, 2009; Loeb and Su, 2010). Models estimate an anthropogenic AOD at 550 nm of 0.03 ± 0.01 relative to 1850, which represents 24 ± 6% of the total AOD (Myhre et al., 2013), though some satellite-based studies suggest much higher values, especially over land (Loeb and Su, 2010). The IPCC estimates with medium confidence that between 20 and 40% of the global mean AOD is of anthropogenic origin (Boucher et al., 2013). It also states with low confidence that the anthropogenic fraction of CCN is between one fourth and two thirds in the global mean.

Aerosol-cloud interactions affect the climate system through various effects on energy fluxes in earth’s atmosphere. The subsections below describe aerosol primary and secondary effects on clouds and atmospheric structure. These effects can vary with initial cloud and atmospheric conditions, as well as with aerosol characteristics.
2.1.1 Direct radiative forcing, and resulting “radiative” effects

Changes arising from the aerosol scattering and absorption of radiation are referred to as direct radiative forcing. Both human-made and naturally occurring aerosols enhance scattering and absorption of solar radiation. Carbonaceous aerosols tend to absorb solar radiation within the atmosphere. Previous research found that the reduction of surface insolation could be up to 20-40 Wm$^{-2}$ for a typical polluted sky for New York City (Jin 2005).

When particles absorb and scatter radiation aloft they can cause secondary changes to atmospheric temperature and humidity profiles, and influence dynamical processes. This is known as a ‘radiative effect’ (Koren et al, 2008; Altaratz, 2014), and is also called the ‘semi-direct’ effect (see also section 2.1.4). Solar energy that stays in the atmosphere instead of reaching the ground can result in the suppression of rainfall, and less efficient removal of pollutants. Radiative effects can also modify microphysical processes such as cloud particle condensation and evaporation rates.

Absorbing aerosols cause local heating in the atmosphere (Hansen et al., 1997; Koren et al., 2004, 2008; Davidi et al., 2009) and can reduce the relative humidity in that layer, while reducing warming at the surface. Heating of the lowest layer of the atmosphere has been found to inhibit the formation of shallow clouds (see also section 2.1.1). Cloud suppression has been found with absorbing aerosols in low clouds over the Amazon (Koren et al., 2004; Feingold et al., 2005; Koren, 2008). Absorbing aerosols can alternatively reduce cloud
cover by mixing with and warm shallow broken clouds in the same layer (Ackerman et al., 2000).

When aerosols reside above clouds, aerosol absorption can be greatly amplified by multiple scattering between aerosol and underlying cloud (Yu and Zhang, 2013). Smoke over marine stratocumulus clouds can warm the free-troposphere and create a stronger capping inversion, resulting in a thicker and lower cloud layer below (Wilcox, 2010). This process stabilizes the temperature profile below the cloud layer and reduces the surface heat and moisture fluxes. Associations have been found between absorbing aerosols and increased stability in low-level layers, and inhibited development of clouds (Yu et al., 2002; Feingold et al., 2005).

Clouds also affect aerosol properties. It was reported that the cloud diurnal cycle affects aerosol forcing in the Indian Ocean Experiment by up to 1–2 Wm$^{-2}$. Aerosol loading, for example, is greatly reduced as particles are scavenged by rain drops. Aerosol size distribution can also be changed by aerosol-cloud interactions (Remer and Kaufman, 1998).

2.1.2 The cloud albedo effect, or the first indirect radiative forcing

Aerosols increase the reflection of solar radiation to space through a variety of complex radiative and microphysical processes (Penner et al., 2001). This can result in negative
Radiative forcing as clouds become more efficient reflectors of sunlight, while they reemit longwave radiation back to the surface (Ramanathan et al., 2001).

Aerosols act as cloud condensation and ice nuclei in the atmosphere. More polluted clouds have an increased droplet concentration and a smaller cloud droplet effective radius than their un-polluted counterparts with an equal amount of total water (Haywood, 2000; Penner, 2001; Ramaswamy, 2001). This increased droplet number concentration leads to an increased optical thickness of the cloud, resulting in greater reflection to space of solar radiation from clouds, leading to more climate cooling (Twomey and Cocks, 1977; King et al., 1993, Haywood et al., 2000). This is called the cloud albedo effect, and has also been known as the first indirect radiative forcing, or the Twomey effect.

Other studies have supported the physical basis of this theory (IPCC, 2013). Facchini (1999) found that organic aerosols acted as surfactants, lowering cloud droplet surface tension and critical supersaturation value, thereby allowing more particles to become activated cloud droplets. In situ aircraft observations of cloud droplet concentrations have been in agreement with predictions based on the aerosol properties (Fountoukis et al., 2007). Model predictions including entrainment effects have also been in agreement with cloud vertical profiles of droplet effective radius (Lu et al., 2008).
2.1.3 The cloud lifetime, or “second indirect” effect

More, and smaller droplets can increase the number of clouds and also lead to an increase in cloud lifetime, and alter their likelihood and timing of precipitation. The increased cloud lifetime is also known as the Albrecht effect. This microphysical effect leads to suppression of precipitation in polluted clouds. A greater number of smaller water droplets for an equivalent level of water in polluted clouds results in droplets that never reach the 14 micron threshold needed to fall out (Penner et al., 2001; Haywood et al, 2000, Ramaswamy et al., 2001). Aerosols can reduce rainfall due to the reduced source of heat at earth’s surface, and through reduced droplet size (Rosenfeld, 2000). Aerosols, including those of urban and industrial sources, have been shown to weaken the hydrological cycle by suppressing rain and snow (Givati and Rosenfeld, 2004; Ramathan et al, 2002). A longer cloud lifetime can also result in overall cooling as brighter, longer-lived clouds lead to reductions in the amount of solar irradiance reaching Earth's surface. Aerosol first and second indirect effects have been found to reduce global solar flux by 21.55 to 24.36 W m\(^{-2}\) in simulations (Menon et al., 2002). Some models find a linear effect in of cloud thickness on lifetime. Others find a U-shaped curve, with effects diminished as AOD increases.

2.1.4 The “semi-direct” effect

Carbonaceous aerosols absorb solar radiation within the atmosphere (Hansen et al, 1997, Ramaswamy et al., 2001). Through this absorption black carbon (BC) aerosols in a cloud layer can heat the air, resulting in cloud evaporation and reduction. This is known as the
'semi-direct’ effect of absorbing aerosols (Hansen et al., 1997).

A study by Koren et al (2008) details the conceptual model for effects of absorbing aerosol cloud reduction, including the semi-direct effect. This study also provides observational data to show that at a lower initial cloud fraction in particular, increasing aerosol optical depth is accompanied by a decline in cloud fraction. The non-linear relationship seen in Figure 2.1 shows 2006 observational data over the Brazilian Amazon. Microphysical effects account for the initial rise, while absorption effects, including the semi-direct effect, explain the latter reduction. The semi-direct effect is a particular kind of radiative effect, and can also refer to cloud reduction through atmospheric stabilization, described in section 2.1.1 (Feingold et al., 2005).

![Figure 2.1](image)

Figure 2.1: Cloud fraction versus aerosol optical depth (τ) for a dataset restricted to a cloud fraction less than 0.5.
2.1.5 Other cloud dissipation effects

Reduced droplet size has been associated with reduced droplet sedimentation, and higher rates of evaporation and entrainment. These effects can result in accelerated cloud evaporation and dissipation.

Smaller droplet sizes at the tops of polluted clouds have lower fall speeds and sedimentation rates, and maintain a higher initial cloud droplet density available for evaporation. In a modeling study higher entrainment rates are attributed to more evaporative cooling and outgoing longwave radiation due to the presence of more available liquid water (Bretherton et al., 2007). More polluted clouds experience greater heat loss, and more sinking of cold air in the entrainment zone, resulting in thinner clouds. This is referred to as the “sedimentation” effect. These clouds have been found to have reduced sedimentation rates and an ultimately reduced liquid water path (Ackerman, 2004; Bretherton et al., 2007).

In the “evaporation-entrainment” effect the shorter evaporation time scale of smaller cloud-top droplets is the driver of the thinning process. The idea, illustrated below in Figure 2.2, was raised by Jiang et al. (2006), and has been identified in other studies (Hill et al., 2009; Small, 2009). The figure compares clean with polluted clouds. Polluted clouds experience higher rates of evaporation at cloud edges, producing horizontal negative buoyancy gradients, or cool downward flows. This results in stronger vortical circulation, and in turn higher entrainment rates with drier air from the free atmosphere above. This feeds back into increasing evaporation, resulting in clouds with reduced lifetime, liquid
water path, and reduced cloud frequency.

Figure 2.2: Schematic of the evaporation-entrainment feedback mechanism (Small, 2009).

2.1.6 Cloud and rainfall invigoration

Rainfall suppression can ultimately result in greater downwind rainfall as clouds are allowed to take on more moisture before precipitating. Increased availability of CCN initially results in smaller cloud droplets which can delay the onset of precipitation. In convective clouds this may ultimately result in deeper, more vigorous convection (Khain, 2005; Boucher et al., 2013; Stevens and Seifert, 2008). In convective clouds, smaller cloud droplets freeze at higher altitudes above the cloud base, and can create invigorating
updrafts through the release of latent heat higher up in the atmosphere (Devasthale et al., 2005; Koren et al., 2005). This “thermodynamic effect” may lead to higher cloud-top heights.

This process illustrated below in Figure 2.3. In warm clouds less precipitation means more liquid is lofted to the cloud-top region, where it evaporates. The associated cooling destabilizes the environment, making it conducive to the growth of deeper clouds, which in turn can produce more rain. The greater delayed rain production can more than compensate for its initial suppression.

![Figure 2.3: The deepening effect. The local inhibition of precipitation helps precondition the environment for deeper convection, which then rains more (from Stevens, 2009)](image)

Koren's 2012 study showed rainfall intensification was statistically correlated with relatively high aerosol levels within a dataset restricted to low aerosol days. The greater availability of cloud condensation nuclei ultimately result in deeper clouds, countering the
effect of cloud suppression. Figure 2.4 and Figure 2.5 below show these outcomes, in which relatively low aerosol levels have the greatest impact on precipitation. The upper cut-off for AOD in Koren’s 2012 study was 0.3. Figure 2.4 (a) shows the rain rate for 13:30 local time during June, July and August 2007. The study area is in the Atlantic off the east coast of the US. In (b) rain rate, $R$, increases as a function of AOD; vertical bars represent the standard error. Note that in (c) the rain rate intensifies with AOD in all cases.

![Figure 2.4](image)

**Figure 2.4:** (a) Map of the study area in the red box, showing average rain rate ($R$). (b) Average $R$ values for six aerosol-loading sets (blue, including zero $R$ grid squares; red, without zero $R$ grid squares). (c), $R$ histograms for three aerosol-loading sets (average AOD: blue 0.07, red 0.12, black 0.22). (adapted from Koren, 2012).
The IPPC summarizes our evolving understanding of expected aerosol effects as follows: in less polluted clouds with a greater precipitation potential increased aerosol is more likely to result in increased cloud cover by inhibiting precipitation. In non-precipitating conditions, clouds tend to thin in response to increasing aerosol through a combination of droplet sedimentation (Bretherton et al., 2007) and evaporation–entrainment adjustments (Hill et al., 2009, Boucher et al., 2013).

2.1.7 Cold clouds

Cirrus clouds have an annual global average frequency of occurrence of about 30% (Wylie and Menzel 1999; Wang et al., 1996; Rossow and Schiffer, 1999). These clouds modify the global radiative balance by scattering short-wave radiation. In the upper troposphere they are an important absorber of and emitter of long-wave infrared radiation (Liou 1986; Ramanathan and Collins 1991).
Ice formation in cirrus clouds may result from both homogeneous freezing of solution droplets formed on soluble cloud condensation nuclei (CCN) at very cold temperatures (below -37°C), and heterogeneous freezing. For sulfate haze droplets at upper-tropospheric temperatures homogeneous nucleation requires supersaturations with respect to ice in excess of 40% (Koop et al. 1998; Bertram et al. 2000). Heterogeneous nucleation involving Ice nuclei (IN) requires a lower supersaturation, and is much more common. IN freeze supercooled water at higher temperatures (−7 °C to −15 °C) than in cases of homogenous freezing (Altaratz et al., 2014). Insoluble or partially insoluble aerosol particles can serve as a substrate for heterogeneous ice nucleation. These are usually solid or crystalline aerosol particles. Heterogeneous mechanisms include direct deposition from vapor to ice on a suitable nucleus (deposition nucleation) and freezing of previously condensed supercooled cloud or haze droplets. Freezing is initiated either by contact of nuclei with the cloud droplets (contact nucleation) or by nuclei immersed within the cloud or haze droplet (immersion nucleation; Liu 2007, Cotton & Yuter, 2009).

Ice nuclei have an important influence on mixed-phase clouds, in which the formation of ice depends on heterogeneous freezing initiated by IN. A very small fraction of aerosols are able to act as ice nuclei, such that much higher saturation levels are found over ice than over water in the atmosphere (Gettelman et al., 2006). Additional available IN could enhance heterogeneous nucleation, and significantly lower the threshold supersaturation required for ice nucleation, with significant impacts on the microphysical and macrophysical evolution of clouds (Kärcher et al., 2007).
In observational studies of convective clouds a negative correlation has been found between aerosol pollution and ice crystal effective radii (Jiang et al, 2009, Sherwood, 2002a,b). Some modeling studies have found a “cloud glaciation effect,” wherein increased IN enhances glaciation and ultimately reduces cloud lifetime (Lohmann and Feichter; 2005, Hoose; 2008, DeMott 2010). Increased ice nucleation increases the albedo of these clouds, but the increased IN numbers enhance Bergeron-Findeisen (BF) ice-crystal growth, making precipitation more likely. Storelvmo et al. (2011) found reduced cloud lifetime to counter-balance the radiative effects of increased albedo in a modeling study. The BF process also results in reduced total particle concentration. Verheggen et al. (2007) found the activated fraction of aerosols to remain constant with increasing aerosol in mixed-phase clouds. The exception was in the cleanest clouds, in which aerosol particles were incorporated as cloud nuclei at a higher rate. Black carbon (BC), a product of oil and motor vehicle combustion, and can serve as an ice nucleus (IN) in the glaciation effect. BC has increased since pre-industrial times may have caused changes to the lifetime of mixed-phase clouds and thus to radiative forcing (Lohmann, 2002). Lohmann (2002) found that anthropogenic aerosol, including BC resulted in decreased cloud lifetime due to the same processes.

2.1.8 Aerosols and cloud Interactions in New York City

This study uses a single measure for all fine-mode aerosol. Aerosols smaller than 2.5-μm in New York City are composed, in descending order of volume, of secondary sulfate, road dust, sea salt, secondary nitrates, and organic and black carbon (Li et al., 2004). Its composition varies, with organic aerosol and sulfate on average comprising 54% and 24% respectively of total PM1 mass in a different study (Sun et al., 2011). While cloud effects
are influenced by aerosol composition, particle size appears to be more important than composition in aerosol-cloud interactions. More information on aerosol distribution in NYC can be found in Appendix B5.

Regional distribution of aerosol for the northeastern US can be seen in Figure 2.6, which uses a combination of MODIS (Moderate Resolution Imaging Spectroradiometer) and MISR (Multiangle Imaging Spectroradiometer) satellite instruments to calculate AOD (Donkelaar et al., 2010). Both instruments are aboard the sun-synchronous National Aeronautics and Space Administration’s (NASA) Terra satellite. Terra’s orbit encircles the earth approximately 15 times each day; MODIS provides wider spatial coverage, while MISR provides ongoing high spatial detail. This figure is important in the context of the aerosol measure in this study, which is only taken at the urban location. While variation in aerosol level is only measured in one place, this figure demonstrates that fAOD in the NYC-metropolitan region is on average significantly elevated over that of its rural counterparts.
Figure 2.6: Mean AOD for 2001–2006 of the MODIS and MISR satellite instruments for the Northeast United States (adapted from Donkelaar 2010).

Figure 2.7: shows local spatial variability of modeled PM2.5 (particulate matter smaller than 2.5 µm) concentrations based on 1999–2007 data for the New York City area for August 2006 (Yanosky et al., 2014). This image highlights the transportation contribution to air emissions, and their further concentration directly surrounding the urban center of Manhattan. Local industry, including industrial areas in the boroughs near Manhattan, contributes additionally to local air emissions. Adjacent to Manhattan in northern New Jersey there are major population centers, heavy industrial zoning as well as Newark Airport.
Figure 2.7: Predicted PM2.5 concentrations (from the 1999–2007 model) in of New York City and environs, for August 2006 showing local spatial variability (5th to 95th percentiles shown). (Yanosky 2014)

When demand is moderate about half of the energy consumed in NYC is produced by the “in-city fleet” of 24 power plants within or directly connected to New York City. Almost two-thirds of this fleet is over 40 years old, and has lower efficiency and higher air emissions than modern plants (Bloomberg, 2013). Figure 2.8 shows some local power-generating stations, using natural gas and some coal. During peak load production shifts to meet higher demand, and at those times more than 80 percent of New York City’s energy is produced by local power-plants. The highest peaks are in summer, but on cold winter days the city’s natural gas demand exceeds pipeline capacity, and local generating plants switch from natural gas to liquid fuel. These factors compound the peak amounts of both anthropogenic heat and aerosol pollution produced in summer and winter.
Submicron particles constitute almost all CCN, and anthropogenic pollutants tend to be in this size range, thus the selection of fine-mode particles for this study. Soluble fraction can be an important parameter in predictions of aerosol-cloud interactions, which varies over a limited range of about 40–90%. However particles size, again, is the most important factor in the ability of aerosols to act as CCN (Andreae and Rosenfeld, 2008). This is because soluble mass changes with the third power of particle diameter, but only linearly with soluble fraction (Dusek et al., 2006). This means that a very large change in composition such as the reduction of soluble fraction from 100% to 10% has about the same effect as a reduction in fully soluble particle size by 50%. Most particles after some atmospheric aging are to some degree internally mixed. And almost all particles contain some
deliquescent component that will aid the initial water uptake and growth even of particles dominated by very low-solubility compounds (Andreae and Rosenfeld, 2008).

In a study of New York City, Hosannah (2014) found that particle size distribution with a high volume of fine mode particles in urban areas can suppress precipitation. Particle size distribution with a high volume of coarse mode particles (with giant cloud condensation nuclei) enhanced precipitation totals. Urban factors also changed spatial precipitation patterns. Jin et al (2005) found weekly signals of aerosol optical thickness for summer months in AERONET measurements, and in MODIS cloud effective radius over New York City. Aerosols were also elevated seasonally, with peaks in summer. This study also found that cloud optical thickness decreased with increased AOD, possibly because aerosols also affected the liquid content of clouds, which may decrease due to evaporation. Cerveney and Balling (1998) also found rainfall over New York City and the surrounding region to be increased on weekends compared with weekdays. Other studies have pointed to an equalization of initially suppressed precipitation when simulations are run for longer time-periods (Flossman and Wobrock, 2010). Therefore a finding of suppressed or enhanced precipitation may represent a delay rather than a net effect.
2.2 Urban Heat Island (UHI) and cloud interactions

2.2.1 Definition of an Urban Heat Island

The "urban heat island" effect takes place when the air in a city is 2-8°F hotter than the surrounding countryside (U.S. EPA, 1992). The UHI phenomenon is created in part by differences between thermal properties (e.g., heat capacity and thermal inertia) of artificial urban surfaces and natural land surfaces. Urban landscapes have reduced vegetative cover. In less developed areas trees and other vegetation cool the air by evapotranspiration, the evaporation of water from the surfaces of leaves and the soil. Because water has a high specific heat, when it undergoes a phase change from liquid to vapor it absorbs a great deal of heat. The loss of latent heat results in a cooling effect on the surface from which it evaporates. Vegetation also cools by shading buildings and blocking solar radiation. Urban surfaces, by contrast, such as roof and paving materials with low reflectivity absorb more solar radiation. These materials store this energy and convert it to sensible heat. Other factors contributing to the onset of UHI include differences in surface albedo, and anthropogenic heat release in the urban area.

The idealized urban heat island effect is illustrated by Figure 2.9, illustrating temperature variations across different land surfaces. Another idealized image is Figure 2.10, showing diurnal fluctuations in UHI intensity. The magnitude of the UHI effect is defined as the urban-rural temperature difference, written as $\Delta T_{U-R}$. Part (d) in this figure shows how the loss of heat is much greater in the evening in rural compared with urbanized regions,
when urban structures release sensible heat, and (e) shows that the strongest UHI is from about 7pm to 7am.

Figure 2.9: Idealized urban heat island  (http://eetd.lbl.gov/HeatIsland/HighTemps/)
Figure 2.10: Idealized urban/rural differences under UHI conditions. (a) shows urban and rural temperatures on a diurnal period from noon to noon; (b) shows the rates of change of those temperatures, and (c) shows the difference between urban and rural. (Cox, 2004)

The UHI effect is most evident on a clear and windless night with peaks in the late evening to early morning hours (Kim and Baik 2002). UHI magnitude is generally proportional to city size (Oke 1981).
2.2.2 Urban Heat Island and land cover-climate interactions

Urban heat island interacts with and affects weather patterns in various ways. Greater heating over urban regions compared with rural counterparts result in horizontal gradients and an upward or downward flux in a thermally forced system.

Urban heat island affects the climate by modifying boundary layer processes (Shepherd, 2005). Figure 2.11 below shows various scales linking urban environments to the environmental system (Oke, 1987). Micro and local scale effects of the urban canopy layer may affect the movement of a storm front; at a larger scale the urban “plume” will carry warmer and more polluted air, affecting weather downwind.

Figure 2.11: The urban canopy layer (UCL) and urban boundary layer (UBL) (modified after Oke 1987).
Equation 1 below shows the surface heat budget equation. The terms are QSW (net shortwave irradiance), QLW (net longwave irradiance), QSH (surface sensible heat flux), QLE (latent turbulent heat flux), QA (anthropogenic heat input), and QG (ground heat conduction). Differential heating from horizontal gradients in one or more terms of equation (1) can result in thermally forced systems with an upward or downward flux of heat, which can in turn influence mesoscale circulations.

\[ Q_{SW} + Q_{LW} + Q_{SH} + Q_{LE} + Q_{G} + Q_{A} = 0. \]  

Equation 2.1

Figure 2.12 shows how energy from the sun is gained in both urban and rural systems, including the lower albedo of urban systems shown in the difference in QR. Calculated values of the corresponding increases in storage heat, and in release of sensible heat are also illustrated in Figure 2.12, as well as the reduced latent heat release in urban areas.

Figure 2.12: Typical rural and urban surface energy balance. The values are in units of kW h m² day⁻¹ (R. Sass, 2015).
Information about the Urban Heat Island social and economic costs can be found in Appendix A.

### 2.2.3 Effects on clouds and precipitation

As noted earlier, while this study does not separate rain clouds from others, precipitation enhancement, if it occurs, would be consistent in this study of short-term effects with higher cloud fraction and albedo, and lower SWIRR measures. The same processes that lead to precipitation, such as of a UHI-enhanced convective cell, may induce or enhance the formation of non-precipitating clouds.

Urban areas have been shown to enhance the intensity of storms and increase downwind rainfall (Huff and Vogel, 1978; Changnon, 1978; Shepherd et al. 2002). Urban land cover can affect clouds in a number of ways. Possible mechanisms for urban impacts on convection include one or a combination of the following: 1) enhanced convergence due to increased surface roughness in the urban environment (e.g., Changnon et al. 1981, Bornstein and Lin 2000, Thielen et al. 2000); 2) destabilization due to UHI-thermal perturbation of the boundary layer and resulting downstream translation of the UHI circulation or UHI-generated convective clouds (e.g., Shepherd et al. 2002, Shepherd and Burian 2003); 3) enhanced aerosols in the urban environment acting as additional cloud condensation nuclei (CCN) (e.g., Diem and Brown 2003; Molders and Olson 2004); or 4) bifurcating or diverting of precipitating systems by the urban canopy or related processes.
(e.g., Bornstein and Lin 2000, Loose and Bornstein 1977). As clouds form or are enhanced in the lead-up to these precipitation processes, greater cloudiness can be expected, whether or not precipitation results.

Landscape heterogeneities have been found in atmospheric models to trigger the formation of mesoscale circulations, and as shown further below, empirical studies have attributed the same mechanisms to their findings. In a model study by Avissar and Liu (1996) the circulation is initiated in the originally dry part of the domain, but a thermal cell creates an upward motion that eventually homogenizes land water content. The result is shallow convective clouds as warm, moist air is transported to higher elevations. Days with high UHI tend to be clear, and other studies find elevated humidity to be an important factor in initiation of convection. Cloud formation was also found by a large-eddy simulation model using a varied Bowen ration between two patches. In a study using two-dimensional numerical model simulations UHI heating initiated moist convection and resulted in surface precipitation in the downstream region (Baik et al., 2001). Other work has shown that the urban circulation is primarily enhanced by the related factors of increased sensible heat fluxes and surface roughness of the urban area (Huff and Vogel, 1978).

Urban areas can provide a low-level moisture source that favors UHI-induced precipitation needed for convective development. This can take place even in temperate regions, where artificial irrigation sources are not the principle cause (Dixon and Mote, 2003). Though more paved surfaces mean that urban landscapes produce less water vapor, greater heating can lead to higher incidence of rainfall (Burian and Shepherd, 2005), as in the
destabilization effect. Vukovich and Dunn (1978) show that heat island intensity and boundary layer stability have dominant roles in the development of heat island circulations.

A number of UHI circulation studies focus on particular seasons, and many on summer in particular. Urban effects lead to increased precipitation during the summer months (Changnon et al. 1977; Huff 1986). Increased precipitation is typically observed within and 50–75 km downwind of the city reflecting increases of 5%–25% over background values (Huff and Vogel 1978; Changnon 1979; Changnon et al. 1981; Braham et al. 1981; Changnon et al. 1991). A warm season study of rainfall patterns around various US cities found precipitation increases of up to 51% in urbanized compared with surrounding areas (Shepherd et al., 2002). In an observational study based on twelve years of lightning data in Houston UHI-induced convergence was a central factor higher lightning rates (Orville et al., 2001). This study found the highest rates in summer and winter, and increases were also attributed to anthropogenic aerosols. An urban-boundary layer study using observations and a simulation found summer daytime small-scale urban convective rolls and UHI-modified local circulations (Miao et al., 2009). Another observational and simulation study found urban strengthening of land-lake breeze circulations and strengthened upward wind over the urban area (Zhang et al., 2011). In coastal or lake-side cities a morning sea breeze is another daytime weather effect that will interact with urban heating cloud processes.
Even though UHI is strongest late evenings to early mornings, UHI circulations are more likely to take place in the daytime because of the urban-rural pressure gradient and vertical mixing during daytime hours (Shreffler, 1978; Fujibe and Asai, 1980).

2.2.4 UHI, weather, and climate of New York City

In a study of New York City’s UHI, Gedzelman et al. (2003) found that UHI is most pronounced on calm, dry, clear nights, and in high-pressure anticyclonic conditions. In this situation heat stays in the city while the countryside experiences a nocturnal inversion as the ground radiates heat rapidly to space. The smallest UHI was seen in anomalous “backdoor cold fronts” in late winter and spring, in which cold air approaches from the northeast. Sea breezes are a more common occurrence, producing low, or inverted UHI conditions. These are found most often on warm spring days when the largest land-sea temperature contrasts are found, as well as in summer, and to a lesser extent in spring.

The ‘urban canopy’ of New York City has also been found to interact with mesoscale weather systems. Loose and Bornstein (1977) found that under high UHI conditions synoptic-scale fronts were accelerated by 25% after they pass over New York City. This study also found that under non-UHI conditions frontal movement was retarded by 50% over the city. This was attributed to the frictional drag on the front from surface friction of the urban canopy. Another observational study found that under non-frontal conditions winds decelerate under non-UHI conditions, and accelerate during them (Bornstein and Johnson, 1977). This study found that at wind speeds below 8 miles per hour (3.6 m/s) air
was accelerated as it passed through the city, and above that speed it was decelerated. Analysis by Bornstein and LeRoy (1990) found that New York City affects both summer daytime thunderstorm formation and movement. They found radar echo maxima on the lateral edges and downwind of the city. A later modeled study using data collected from the NYC-metropolitan region also found ‘barrier effects’ of the urban canopy which slowed wind speeds during an urban heat island (Bornstein et al., 1993). The METROMEX study, cited above for its findings on summertime precipitation, also found a 10% increase in summertime cloudiness in St. Louis due to urban effects.

Childs and Raman (2005) found enhanced convergence and upwind vertical velocities over NYC at 18 UTC. This was under sea breeze conditions in a study that modeled several case studies in combination with a cluster of high resolution meteorological sensors. Enhanced 100 m level convergence was simulated over the same region, with maximum upward vertical motion exceeding 0.6 ms⁻¹. This study also found a maximum of turbulent kinetic energy over Manhattan between 17 and 18 UTC, and a minimum nearby over the Hudson River. During this time the boundary layer grew from 400 to 800 meters, most likely due to heating from urbanized land use. In a numerical study of shallow convective clouds induced by land surface forcing, clouds develop between 12:00 and 16:00 local time (Avissar and Liu 1996). In this "active phase" horizontal thermal and pressure gradients provided enough energy to create and sustain mesoscale circulations.

A modeled anthropogenic heat flux study found New York to have the highest individual grid cell heat flux, at 577 W m⁻² (Allen et al., 2011). It had the second-highest highest
average heat emissions, after Tokyo, which produces 60.8 W m². In Allen’s study this heating has a strong diurnal pattern, increasing seven-fold between 3am and 1pm.

Figure 2.13 below shows a synthesis based on a literature review of urban effects on precipitation in the context of a modeling study in NYC, showing high UHI days as associated with more and deeper clouds (Hosannah, 2013). In this paradigm weather ‘dynamics’ of wind, UHI, and convergence or divergence first create conditions for rainfall; precipitation efficiency is modulated by aerosols. This study ran NYC land-cover simulations which included observational data including aerosols. In simulations with and without urbanization, the urbanized run had added convection and a higher cloud base height starting at noon local time. It also found that land cover had more of an impact than aerosol particle-size distribution.

Figure 2.13: Synthesis of urban effects on precipitation. Based on the ideas of Robert Bornstein (2011; personal communication in Hosannah, 2013), Grimmond (2011) and Shepherd (2011) (From Hosannah, 2013)
2.2.5 Study area

The scale of urbanization in the large metropolitan region of NYC is ideal for the study of UHI since its effect is proportional to city size (Oke, 1981). This high-density region, including nearby parts of New Jersey across the Hudson, has undergone greater development since 1980, but due in part to zoning restrictions in outlying areas, it remains a very centralized urban region, especially compared with other US cities like Los Angeles or Atlanta.

In order to understand the effects of the NYC-region’s a broader study area is needed. This allows for downwind effects to be captured within the study area where necessary. It also allows for the inclusion of less urbanized, more vegetated control regions to be included for comparison. The study area below (Figure 2.14) is a 10° x 10° degree grid centered on New York City.
Figure 2.14: Study domain, NYC metropolitan region is indicated by the red circle.
3 DATA USED AND SOURCES

3.1 Ground weather station sources for UHI calculations

3.1.1 Data sources for daily UHI calculations

The National Climatic Data Center is the world’s largest active archive of weather data. The Global Summary of the Day (GSOD) dataset is one of many station datasets available through this service. Measurements provided at these stations include temperature, dew point, pressure, wind speed, and precipitation. NCDC (National Climatic Data Center) also provides these measurements from different sources, as well as hourly datasets used in this study. GSOD was the source for the six-station rural UHI measure described in section 4.1.1.

3.1.2 Wind sources for downwind data extraction

Hourly wind speed and direction data was obtained from the NOAA National Climatic Data Center. This is part of the World Meteorological Association’s Global Surface Hourly database. Hourly ground weather station data was available for a 1998-2000 time period, though these sources have some data gaps within those years. Reanalysis and model output data at the NOAA National Climatic Data Center and Earth System Research
Laboratory were explored extensively, but the resolution of these datasets were all unsuitable. Either spatial or temporal resolution was too large, or the time range of data availability was too limited.

3.2 GOES satellite data for clouds

The Geostationary Operational Environmental Satellite (GOES) system is operated by the United States National Environmental Satellite, Data, and Information Service (NESDIS). It supports weather forecasting, severe storm tracking, and meteorology research. Spacecraft and ground-based elements of the system work together to provide a continuous stream of environmental data. The GOES system uses geosynchronous satellites for research, as well as weather monitoring and forecasting. The GOES-East satellite is located at 75 degrees West, over the equator.

All data was downloaded at a 4km resolution. GOES geostationary satellite (GOES-East) channels 1, 2, and 4 were used in this investigation. GOES visible channel 1 was used to derive visible albedo, an important radiative measure both itself, and in relation to changes in cloud microphysics. Shortwave infrared reflectance (SWIR), calculated using GOES channel 2, which has its central wavelength at 3.90-μm is used here as a proxy for cloud particle radius. The modeled inverse exponential relationship between cloud droplet size and SWIR reflectance is shown in Figure 3.1. The relationship is presented with effective radius on a logarithmic scale. Effective radius increases exponentially with decreasing SWIR reflectance. Brightness temperature from GOES channel 4, thermal infrared (TIR), at
10.7-µm is used to remove emission in the 3.9-µm reflectance (SWIRR) calculation, and to filter cloud types based on cloud-top temperature.

![Figure 3.1: Modeled inverse exponential relationship between cloud liquid particle radius and shortwave-infrared reflectance. (Cattani, 2007)](image)

GOES data was collected for a period of ten years between 1999-2009. The hour selected was 17 UTC, or noon or 1pm local time. The time of day is discussed in section 5.2.

More information about GOES data, including on calibration is in Appendix C.
3.3 AERONET aerosol data

Daily averaged Aerosol Optical Depth (AOD, also called AOT, aerosol optical thickness) at 500-nm ($\tau_a$) is a dimensionless measure of the total column amount of aerosol. It represents the attenuation of sunlight by a column of aerosol at a particular wavelength. AOD is the key parameter for modeling the radiative effects of column aerosol, as in studies using the MODIS remote sensing algorithm (Kaufman et al., 1997; Chu et al., 2002, 2003). The 500 or 550-nm AOD channel is commonly used for anthropogenic AOD measurements and in studies exploring aerosol-cloud interactions (Jin 2005, Devara, 2013; Manoj et al, 2011), and is closest to the wavelength MODIS uses as its aerosol measure of 0.56-µm (Kim 2004).

Aerosol optical depth data is available in global datasets from the Aerosol Robotic Network (AERONET), a ground-based aerosol/radiation source. This source has the advantage of being a high-quality aerosol data available at the center of the study area. This study used data from the City College station, located in New York City, in upper Manhattan. This AOD dataset is available from the year 2000 to the present, and was collected for the period of 2000 through 2008. Very little data is available prior to 2002 or for 2009, thus the 2002-2008 study period. The City College station is a far more consistent AERONET source than most in the region, but even so its temporal availability is irregular. Another drawback of AERONET, as with other aerosol sources, is that it reports data only under cloud-free conditions. As such, the aerosol climatology represents only a subset of relatively clear synoptic conditions.
Level 1.5 cloud-screened fine mode data is used, referred to here as fAOD. Fine mode data refers to particles measuring up to about 0.1-1.2 µm in diameter, or a cut-off between fine and coarse mode in the range of 0.44-0.99-µm. (AERONET Inversion Products). Level 1.5 data has the advantage of many more data points than the level 2 data. This data has not had the post-deployment calibration applied to level 2 data. Due to the very limited availability of level 2 data, aerosol assimilation scientists generally make use of Level 1.5. More information about AERONET AOD measurement, level 1.5 criteria, and urban aerosols can be found in Appendix B1.

An example of one day's available data, showing both 500 and other wavelengths, used towards a daily average is below in Figure 3.2.
4 METHODOLOGY

This study will assess relationships between Urban Heat Island and cloud properties, as well as between aerosols and cloud properties. Satellite and ground-based station data will be used in a complimentary fashion. Aerosol measurements will be taken from the ground-based AERONET (Aerosol Robotic Network). Aerosols can affect cloud optical properties through changes in cloud particle size. UHI is measured using ground-based (NCDC) Cooperative Observer Program weather stations.

Derived from GOES satellite channel 1 data, the albedo measure is of interest as a radiative property, and also in its interaction with cloud microphysical properties. Albedo is a measure of direct effects of clouds on the local radiative balance. Cloud albedo can also be altered by cloud microphysical properties, and these possible effects are also discussed. From GOES channels 2 and 4, 3.9-µm reflectivity is used as a proxy for relative cloud-particle size. Reflectance calculations at 3.9-µm are taken from Lindsey et al. (2006).

Clouds in the well-mixed lowest layer of the atmosphere are most likely to be affected by urban land use and pollution. Cloud processes for these ‘warm’ liquid-particle clouds are closely coupled with land surface heat and moisture fluxes, wind, and thermodynamic conditions. This study focuses on effects on low clouds, but also includes high clouds for total column aerosol effects, to include convective clouds, and because cold cloud processes remain an important area of uncertainty for modelers.
For the ten year study period of 2000-2009, data is first divided into seasons of DJF, MAM, JJA, and SON (December, January, February, and so on for three-month seasons). Days for the study period are partitioned into high, medium and low categories of Aerosol, and UHI, respectively. The kind of data extracted from GOES images for each seasonal dataset is shown in Table 1. The data set extracted is of downwind values for the urban location, and for two control locations. Climatologies are produced for each dataset, showing average values for each pixel for that season.

Table 1: datasets created for UHI and Aerosol seasonal runs.

<table>
<thead>
<tr>
<th>Cloud type</th>
<th>Data-type extracted</th>
<th>Numeric plots</th>
<th>Mapped image plots</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>Averaged data</td>
<td>Urban-rural difference</td>
</tr>
<tr>
<td>Warm cloud</td>
<td>3.9-µm reflectivity</td>
<td>✔</td>
<td>✔</td>
</tr>
<tr>
<td>Vis. Albedo</td>
<td></td>
<td>✔</td>
<td>✔</td>
</tr>
<tr>
<td>Cloud frct.</td>
<td></td>
<td>✔</td>
<td>✔</td>
</tr>
<tr>
<td>Cold cloud</td>
<td>Vis. Albedo</td>
<td>✔</td>
<td>✔</td>
</tr>
<tr>
<td></td>
<td>Cloud frct.</td>
<td>✔</td>
<td>✔</td>
</tr>
</tbody>
</table>

**4.1 UHI measure**

UHI is measured as a differencing between urban and “rural,” or outlying non-urban surface temperatures, usually using ground stations. UHI measurements vary widely depending on climate zone, development patterns surrounding the city, topography, time
of day, and season. In some instances a reverse-UHI pattern will appear, in which the urban region is cooler than outlying areas. An example of this phenomenon in New York City is a ‘back-door cold-front’ (Gedzelman, 2003). An urban area in the desert in which the city has more vegetation and irrigation than outlying regions can also be cooler, but this study region is temperate.

A set of ground weather stations were selected to create a working definition of a daytime UHI for New York City. New York’s Central Park was selected as the urban station, though LaGuardia Airport data was used when Central Park data was not available. Locations in outlying non-urban locations were chosen for comparison. These locations were in less densely populated and less and built-up areas with higher vegetative cover. Since the NYC metropolitan region also encompasses smaller cities and extensive suburban land-cover, control locations outside of these areas were selected for downwind data extraction.

There is no general rule for the scale of Urban Heat Islands, as they depend on the size of the urbanized region, and of land surface properties in the non-urban region. Prior research on the NYC UHI has used stations in the 50-150 km (31-93 miles; Gaffin et al., 2008). This measure relies mainly on a UHI calculated using non-urban stations with an average of 105 km (65 miles) from the city center.
4.1.1 Six NCDC ‘rural’ stations used for UHI in climatologies.

For daily UHI calculations a set of 6 “rural” stations was chosen, which are available over a long time-period (1980–2010). Daily measures were used, and UHI was calculated based on the differences of maximum temperature. This study used a daytime UHI measure, in contrast with the more common night-time measure. UHI peaks at night, but UHI-generated circulations are more likely to take place in the daytime because of the urban-rural pressure gradient and vertical mixing during daytime hours (Shreffler 1978; Fujibe and Asai 1980). Anthropogenic aerosol production is highest in daytime hours, as is anthropogenic heat flux, which peaks at 1pm (Allen, 2011). This data complimented the mid-day time for the satellite image reading, which was chosen to minimize reflectivity errors due to higher solar zenith angles.

Station locations are pictured below in Figure 4.1, and their locations and distances are detailed in Table 2 below. As seen in Table 2, the average distance from NYC to these stations is 65.3 miles.
Table 2: Stations used for the 6-station UHI calculation, including average latitude, longitude, and distance to New York City’s Central Park.

<table>
<thead>
<tr>
<th>Stat. Code</th>
<th>Station. Name</th>
<th>State</th>
<th>Distance to NYC</th>
<th>Lat</th>
<th>Long</th>
<th>Elev</th>
</tr>
</thead>
<tbody>
<tr>
<td>KNYC</td>
<td>New York Central Park</td>
<td>NY</td>
<td>Mi Km</td>
<td>40.8</td>
<td>-74.0</td>
<td>47.5</td>
</tr>
<tr>
<td>KFOK</td>
<td>Westhampton Beach</td>
<td>NY</td>
<td>70.0 112.7</td>
<td>40.8</td>
<td>-72.6</td>
<td>20.4</td>
</tr>
<tr>
<td>KPOU</td>
<td>Dutchess Co</td>
<td>NY</td>
<td>56.6 91.1</td>
<td>41.6</td>
<td>-73.9</td>
<td>49.4</td>
</tr>
<tr>
<td>KSWF</td>
<td>Stewart Intnt</td>
<td>NY</td>
<td>50.1 80.6</td>
<td>41.5</td>
<td>-74.1</td>
<td>150.0</td>
</tr>
<tr>
<td>KABE</td>
<td>Allentown Bethlehem-Easton</td>
<td>PA</td>
<td>78.2 125.9</td>
<td>40.7</td>
<td>-75.4</td>
<td>117.3</td>
</tr>
<tr>
<td>KNXX</td>
<td>Willow Grove</td>
<td>PA</td>
<td>74.2 119.4</td>
<td>40.2</td>
<td>-75.2</td>
<td>110.3</td>
</tr>
<tr>
<td>KWRI</td>
<td>McGuire AFB</td>
<td>NJ</td>
<td>62.5 100.6</td>
<td>40.0</td>
<td>-74.6</td>
<td>40.5</td>
</tr>
<tr>
<td>Rural Averages</td>
<td></td>
<td></td>
<td>65.3 105.1</td>
<td>40.8</td>
<td>-74.3</td>
<td>81.3</td>
</tr>
</tbody>
</table>

Figure 4.1 Station map for six-station UHI. New York City (in red) and “rural” stations used for six-station UHI in blue.
A two-sample t-test was run to verify a statistically significant difference between rural and urban values using Matlab’s “ttest2” function. This test was run for all of 1983’s data. The result of this test verifies that for that year of data at a 5% significance level the two populations had unequal means.

4.2 GOES satellite data and cloud properties

Visible albedo is an important radiative property in itself. Channel 2 SWIRR at 3.9-µm was used as a proxy for cloud particle size, or for relative particle size, as presented in Section 3.2. Channel 4 brightness temperature was used for cloud-top temperatures to differentiate warmer from colder clouds. Warmer clouds have a greater fraction of water to ice particles. It was also used to remove the emission portion from the channel 2 SWIRR measure.

The hour selected for GOES images was 17 UTC, or noon or 1pm local time. For mid-day images errors in reflectivity values due to a high solar zenith angle are minimized. While this angle changes with seasons, data is separated by and compared within seasons. Though the visible band is available at higher resolution, 4km resolution was selected for all data so that thresholds in one channel could easily be applied to others. A summary of the derivation of cold and warm-cloud visible albedo, and warm-cloud SWIR is shown below in Figure 4.2.
4.2.1 Thresholds

Various thresholds were applied to identify cloudy pixels, separate warm clouds from cold, and for data quality. These were determined in part through a comparison with Daniel Lindsey’s effective radius model, and in consultation with him. More about the model is found in section 4.2.4.1 below. A summary of initial thresholds is in Table 3 below, and albedo and warm cloud threshold information is presented in sections 4.2.2 and 4.2.5 respectively. For baseline thresholds excluding spurious data, see appendix C2. The initial
thresholds below screen for cloudy pixels, and are initial thresholds for warm, liquid-particle-top clouds. An additional set of conditions are included for warm cloud designation, described in section 4.2.3. The all-cloud plots are the sum of warm and cold cloud outcomes and are only discussed in the context of the other findings.

Table 3: Initial warm cloud thresholds applied to GOES channels.

<table>
<thead>
<tr>
<th>GOES Channel</th>
<th>Central Wavelength (µm)</th>
<th>Lower Limit</th>
<th>Upper Limit</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.65 µm</td>
<td>50% albedo</td>
<td>100%</td>
</tr>
<tr>
<td>2</td>
<td>3.9 µm</td>
<td>283 °K (-10°C)</td>
<td>330° K (57°C)</td>
</tr>
<tr>
<td>4</td>
<td>10.7 µm</td>
<td>256 °K (-17°C)</td>
<td>330° K (57°C)</td>
</tr>
</tbody>
</table>

4.2.2 GOES visible channel albedo

Cloud albedo for warm, liquid-particle clouds is studied in order to assess the data for evidence of a cloud albedo or other radiative forcing from anthropogenic aerosols. GOES visible channel 1, with a central wavelength of 0.65 µm, is used for this measurement. Visible albedo is an important radiative property in itself, as it measures visible reflectance back into space. Albedo also corresponds with cloud optical depth (Park 1974, Twomey 1977, Nakajima and King 1990; Platnick et al. 2001; King et al. 2004), the attenuation of the light passing through an atmospheric layer. Optical thickness is relevant to the amount of radiation reaching the atmospheric layers below, and the ground. Visible reflectance varies mostly with optical thickness (Wetzel et al. 1996). The physical basis of the cloud albedo
and cloud optical thickness retrieval problem is described in the Algorithm Theoretical Basis Document “Cloud Albedo and Cloud Optical Thickness” (Fischer et al., 1998).

Albedo was averaged over seasonal data sets and climatologies for warm and cold cloud pixels with a minimum 50% albedo. This same albedo threshold is applied to all clouds so that all measures describe a consistent set of optically-thick clouds. These images are intended to capture optically thick and convective clouds associated with urban heating, as well as possible aerosol effects on albedo. Climatology methods are described in greater detail in section 4.3.

4.2.3 GOES visible channel Cloud screening

As seen in Table 3 above the central wavelength of GOES visible channel 1 is 0.65 µm, and the 4km resolution was used in this study. Equations used for the calibration of channel 1 for visible albedo are found in Appendix C1. Albedo from the visible channel was used as a screen for pixels with sub-pixel clouds or no clouds.

A 50% minimum albedo threshold was applied for all clouds, below which the pixel was excluded. This is a higher threshold than necessary for the detection of clouds, and excludes thinner clouds. The higher threshold was chosen because of evidence of contamination to the reflectance values at cloud edges of warm clouds in an inspection of case study days. A lower threshold is also more likely to include partly cloudy pixels. A scene with a thin layer of cold cirrus clouds could include transmission of shortwave IR
radiation from below. The inclusion of reflectance from the ground below would result in erroneously warm 3.9-µm temperatures. Cloud systems with a 40% threshold still showed some edge effects in channel 2 radiance that could not by explained by physical reasons, so were likely an effect of partly cloudy readings. At 50% albedo pixels with these effects were excluded.

Albedo cloud thresholds with a minimum of 40% to designate optically thick cloudy pixels are commonly applied (Ba and Gruber 2001, Rosenfeld and Gutman, 1994). A 40% limit is used by Rosenfeld (1994) to prevent errors in partly cloudy scenes. In this work clouds of infinite optical thickness are assumed in the use of 3.7 µm reflectance for inferences about cloud particle size.

Figure 4.3 shows part of the visible image of a sample study day; Figure 4.4 shows the same date and time for the image with a 50% albedo threshold. Note the exclusion of thinner clouds off the Long-Island Sound. Figure 4.5 is the Lindsey particle radius (described in section 4.2.4.1) image for the same scene. Pixels showing cloud edge effects in this image are excluded from the output seen in Figure 4.4.
Figure 4.3: High-resolution GOES visible image for Jan 11, 2004 17:01 UTC
Figure 4.4: GOES 3.9-\(\mu\)m reflectivity (shown in scale on right) for Jan 11, 2004 17:01 UTC with a 50\% minimum albedo threshold. Latitude and longitude are shown on the left and bottom axes.
Figure 4.5: Liquid particle radius in microns. Output Image of Daniel Lindsey’s particle radius model for Jan 11, 2004 17:01. Latitude and longitude are shown on the left and bottom axes.

4.2.4 GOES channels 3.9-μm reflectance for Cloud Top Particle Information

The combined characteristics of 3.9-μm and 10.7-μm brightness temperatures are used in various satellite product applications. An example of such a product can be found in the referenced NASA link (NASA GSFC). See Table 4 for summaries of these channels’ characteristics. Well-known cloud reflectance properties at 3.9-μm, shortwave infrared (SWIR), allow a user to make inferences about cloud particle size. Radiation at 3.9 μm is
composed of both emitted thermal and reflected solar radiation (UCAR, NASA GSFC). The second channel used for the reflectivity calculation is Channel 4, at 10.7-µm, which is a measure of emissivity. Since channel 2 radiance includes emissivity, the 10.7-µm brightness temperature can be used as a correction for reflectivity at 3.9 µm. The reflectivity equations are found in section 4.2.4.4. Table 4 below shows a summary of Channel 2 and 4 properties.

Table 4: Summary of Channels 2 and 4 properties

<table>
<thead>
<tr>
<th>Channel No.</th>
<th>Central Wavelength (µm)</th>
<th>Sample sub-point (E/W x N/S) resolution (km)</th>
<th>Common Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>2</td>
<td>3.9</td>
<td>2.3 x 4.00</td>
<td>shortwave infrared, shortwave IR</td>
</tr>
<tr>
<td>4</td>
<td>10.7</td>
<td>2.3 x 4.00</td>
<td>window, longwave IR</td>
</tr>
</tbody>
</table>

**4.2.4.1 Cloud-top particle radius model**

A cloud-top particle effective radius models was run for selected study days for comparison with the 3.9 reflectivity images. This model was developed at NOAA’s Cooperative Institute for Research in the Atmosphere by research meteorologists Daniel Lindsey and Louie Grasso (Lindsey & Grasso, 2008). The technique is based on the fact that the reflection function in the near-infrared at 3.9 microns is primarily a function of particle size, as discussed below in section 4.2.4.2. Though the model has not been fully validated, the ice-particle radius portion of it has. The cloud liquid particle size retrieval is similar to an ice
particle size retrieval except that it assumes Mie scattering, (Lindsey 2009, personal communication) and a much simpler calculation.

The calculation to compute GOES channel 2 reflectivity uses the TIR, channel 4, to remove emissivity from the Near-IR, channel 2. It is derived from Stevak and Doswell (1991), and can be found in Appendix C5. The techniques for deriving radii for liquid water droplets utilize Mie scattering calculations of spherical particles. It also makes use of lookup tables based on prior work, including that of Nakajima and King (1990) which combines a theoretical approach with remote sensing observations. This model utilizes a variety of radiative transfer calculations for selected wavelengths in the visible and near-infrared to assess the sensitivity of the reflection function to cloud optical thickness and cloud radius.

Output from this model is used to verify in sample days the SWIR-particle-size relationship, to compare with 3.9-µm reflectivity calculations for warm cloud designation, and for optically thick cloud designation.

### 4.2.4.2 Channel 3.9: shortwave infrared

As the size of water droplets becomes smaller they become more reflective at 3.9-µm, and scattering increases (Arking and Childs, 1985, King, 1983, NASA GSFC, UCAR). Turk (1998) also demonstrated with a microphysical model that reflectance at 3.9-µm is highly dependent on particle size and phase. Durkee (1989) demonstrates with in situ measurements that at 3.9-µm there is a good relationship between reflection and effective
particle radius. Figure 4.6 illustrates with modeled data that 3.9-µm reflectance (SWIRR) is strongly dependent on cloud drop radii, and varies weakly with viewing geometry (Rosenfeld and Gutman, 1994). This relationship is also shown in Figure 3.1 in Section 3.2. Infinite optical thickness is assumed for the figure's contour map. The 50% minimum albedo threshold applied to SWIRR climatologies in this work should exclude measurements with ground reflectance. Extinction in clouds in the SWIR range takes place at a shorter distance than in the visible range, so clouds can more easily be considered infinitely optically thick (Rosenfeld and Gutman, 1994). For clouds with a high optical thickness shortwave IR alone can be used to determine effective radius (Wetzel et al., 1996). Figure 4.7 below illustrates the relationship between reflectance and particle radius for water clouds in green at the top left of the chart.

![Figure 4.6](image)

Figure 4.6: Reflectance at 3.9 µm from an infinitely thick water cloud, as a function of the effective radius of the droplets, and of the sun-target-sensor angle (Rosenfeld and Gutman, 1994)
Emissivity from clouds is measured with the GOES thermal infrared (TIR) Channel 4, at 10.7 μm. Thermal, or long-wave IR also provides a cloud-top temperature allowing for filtering out of higher and colder ice clouds. Initial threshold designations for warm clouds are shown in Table 3, and the full set of conditions is in section 4.2.5. Hunt (1973) showed that at 10.7 microns clouds radiate close to the blackbody temperature. Again, since the 3.9-μm radiances reading includes emission in addition to reflection, reflectivity is calculated by removing the emissivity portion using the TIR channel. Hunt found that the 3.9-μm emission is slightly less than that at 10.7, and the thermal component of 3.9-μm reflectance can be approximated in this way (Setvák and Doswell 1991, Lindsey et al 2006).
Stephens (1978) found the emissivity at 10.7-µm for an optically thick cloud is a reasonable approximation to unity.

The relationships assumed in this calculation are as follows, as in Lindsey (2006):

$$R_{3.9} = R_{r,3.9} + \varepsilon_{3.9} R_{e,3.9}(T)$$  
Equation 4.1

Where $R_{3.9}$ is the total radiance at 3.9-µm, $R_{r,3.9}$ is the solar-reflected component at 3.9-µm, $\varepsilon_{3.9}$ is the emissivity of the scene at 3.9-µm, and $R_{e,3.9}(T)$ is the blackbody radiance at 3.9-µm at temperature $T$.

### 4.2.4.4 Calculations for 3.9-µm reflectivity

Equations for reflectivity at 3.9-µm are taken from Lindsey (2006), which uses a method described by Setvák and Doswell (1991). This measure is used as a proxy for relative particle size. As shown in section 4.2.5, the calculated reflectivity matches with model results. As in the Rosenfeld Lensky Technique (RLT; Lensky and Rosenfeld, 2008) there is an assumption that cloud-top particle radius is similar to the radius of particles within the cloud. For the RLT this assumption was verified with in situ aircraft measurements (Rosenfeld and Lensky, 1998; Freud et al., 2005). In equation 4.1 the use of 10.7-µm radiance for blackbody at 3.9-µm is the central assumption. Taken from Lindsey (2006), the second term can be written as:
\[ R_{r3.9} = \alpha_{3.9} \left[ Re_{3.9}(T_{sun}) \left( \frac{A}{B} \right)^2 \cos(\phi) \right] \]  

Equation 4.2

where \( \alpha_{3.9} \) is the 3.9-\( \mu \)m reflectivity, \( Re_{3.9}(T_{sun}) \) is the blackbody radiance of the sun (\( T_{sun} \) is taken to be 5800 K), \( A \) is the radius of the sun, \( B \) is the radius of earth’s orbit, and \( \phi \) is the solar zenith angle. The solar zenith angle calculation is described in section C3. The \( B \) term is variable, and an eccentricity term accounts for this variability, taken from Partridge (1976). The equations used for this are in Appendix C4, Earth's Eccentricity. Test cases for solar zenith angle and Earth's eccentricity were run and compared with online calculator results.

As in Lindsey (2006) only optically thick clouds are chosen, thus transmissivity is assumed to be zero, allowing for equation 4.3:

\[ \varepsilon_{3.9} + \alpha_{3.9} = 1 \]  

Equation 4.3

In Equation 4.4 below, \( S \) is the solar flux at the top of the atmosphere, the equation in brackets in Equation 4.2. The \( \alpha_{3.9} \) term below is the 3.9-\( \mu \)m reflectivity shown in climatology images. The \( R_{3.9} \) term is the total radiance at 3.9-\( \mu \)m as measured by the satellite. Radiance at 10.7-\( \mu \)m, channel 4, is used for the \( Re_{3.9}(T) \) term, the blackbody radiance 3.9-\( \mu \)m at temperature \( T \).
4.2.5 Thresholds for warm clouds

The thresholds in Table 5 below screen for warm, liquid-particle-top clouds. Colder higher clouds are less likely to be affected by local land use or pollution factors taking place near the ground. Because this study is intended to detect a mesoscale effect from urban aerosols, and effects of land surface properties, low clouds are an appropriate target. Higher clouds such as cirrus are more likely to have advected from more distant locations. The presence of ice in cloud-tops would reduce the accuracy of a 3.9-µm reflectance measurement used as a proxy for liquid particle size. These conditions should exclude most cloud-tops with ice. Visible images do not suffer from these distortions, and an additional data set is run for albedo which includes higher clouds.

As seen below in Table 5, when either of two sets of conditions are met the pixel is designated as warm cloud. Apart from these conditions only pixels with a 50% albedo or greater are accepted. The first set of conditions depends more on a relatively warm cloud-top temperature at 10.7-µm of -17°C (256 °K). It also uses a relatively bright 3.9-µm reflectance, setting it at -10°C; the baseline temperature, shown in Appendix C3, is -110°C (163° K). The presence of ice particles greatly reduces reflectance at 3.9-µm. This
reflectance property is utilized in the second set of limits. Here colder clouds as measured at 10.7-μm are accepted if they are very reflective at 3.9-μm. Clouds brighter than 15 °C (288 °K) are unlikely to contain ice, so if that condition is met clouds as cold as -30 °C (243 °K) at 10.7-μm are accepted. These designations were based in part on a comparison of brightness temperature difference images with output from Daniel Lindsey’s particle radius model, and also in consultation with him.

Table 5: Warm cloud thresholds: pixels with either condition 1 or condition 2 applied to GOES channels are accepted.

**Condition 1**

<table>
<thead>
<tr>
<th>GOES Channel</th>
<th>Central Wavelength</th>
<th>Lower Limit</th>
<th>Upper Limit</th>
</tr>
</thead>
<tbody>
<tr>
<td>2</td>
<td>3.9 μm</td>
<td>283 °K (-10 °C)</td>
<td>330 °K (57 °C)</td>
</tr>
<tr>
<td>4</td>
<td>10.7 μm</td>
<td>256 °K (-17 °C)</td>
<td>330 °K (57 °C)</td>
</tr>
</tbody>
</table>

**Condition 2**

<table>
<thead>
<tr>
<th>GOES Channel</th>
<th>Central Wavelength</th>
<th>Lower Limit</th>
<th>Upper Limit</th>
</tr>
</thead>
<tbody>
<tr>
<td>2</td>
<td>3.9 μm</td>
<td>288 °K (15 °C)</td>
<td>330 °K (57 °C)</td>
</tr>
<tr>
<td>4</td>
<td>10.7 μm</td>
<td>243 °K (-30 °C)</td>
<td>263 °K (10 °C)</td>
</tr>
</tbody>
</table>

The images below were used in this assessment, using the same set of test images, including January 11 2004 at 17:00 UTC. Figure 4.3 from section 4.2.2 is the visible image.
Figure 4.8 below shows the same scene as Figure 4.4 excluding all pixels colder than -5 °C at 10.7. This image shows channels 2-4 brightness temperature difference, the 3.9-µm reflectance approximation used before Section 4.2.4.4’s equations were applied. In order to eliminate clouds with any amount of ice based on Channel 4 brightness temperature alone, a threshold of -4°C would be selected. This threshold however eliminates most clouds, including liquid water top-clouds, and is too restrictive. It was compared with a particle radius run using Daniel Lindsey’s model, seen again in Figure 4.9. Figure 4.10 shows the same scene with the thresholds in Table 5 applied. The cloud-tops may have some ice particles but are mainly water clouds.

Figure 4.8: Channels 2-4 brightness temperature difference for January 11, 2004 with a maximum -5 °C channel 4 brightness temperature threshold applied.
Figure 4.9: Daniel Lindsey’s liquid particle radius model Jan 11 2004. Effective radius in microns shown at right.
Figure 4.10: 3.9-µm reflectivity using a GOES Channel 4 warm cloud threshold for January 11, 2004. Lower 10.7-µm limit of 256 °K (-17 °C), and 283 °K (-10 °C) 3.9-µm. It also includes pixels between 243 °K (-30 °C) and 263 °K (10 °C) when 3.9-µm reflectance is greater than 288 °K (15 °C).

4.2.6 Snow cover contamination thresholds

During winter months snow cover can contaminate results of cold cloud albedo since it makes the ground highly reflective in the visible channel, so may not be removed by the 50% minimum albedo threshold. Snow cover is difficult to differentiate from ice clouds using the channels in this study. Allen (1987) had good results in differentiating clouds...
from snow cover with an algorithm using the same channels. This approach was limited in differentiating snow from thin cirrus, but was successful when applied to thick clouds. The method described below was explored as a way to remove possible snow contamination. Ultimately these thresholds were not applied as it was found that the 50% minimum visible albedo threshold was sufficient to remove snowy pixels over land.

Snow cover data from NOAA’s National Operational Hydrologic Remote Sensing Center’s (NOHRC) Regional Snow Analysis for this region was used for comparison. The Regional Snow Analysis estimates percent snow cover, and provides maps of Snow Water Equivalent. Dates for the Northeast region for the dates 2003-2010 were utilized.

Taken from Allen’s ‘temperature factor,’ $F_T$, the following equation was applied:

$$F_T = \frac{T_{3.9-\mu m}}{T_{10.7-\mu m}} - 1$$  \hspace{1cm} \text{Equation 4.5}

The reciprocal of the above equation was applied, as in Allen. A pixel is considered snow, and discarded, under the following conditions:

1) Visible Albedo less than 50 (this cutoff is already applied to all data)
2) 3.9 Reflectance less than 0.057
3) temperature factor greater than or equal to 20
The above conditions would have be applied in winter months (October 10 – May 15), when snow cover has historically been present in the study region during those months. The above conditions would remove cases with snow in empirical comparisons with satellite data in Allen's study. In one of Allen's 23 comparisons of data from observed snow cover and ice clouds, an ice cloud would be erroneously designated as snow, and discarded. All cases of snow cover would successfully be removed.

The winter season of 2005 was examined using the thresholds shown above, as well as variants of conditions 2 and 3 above. Full-resolution GOES channel 1 images for this time period were downloaded to identify clouds, along with snow data from the NOHRC dataset. These datasets were compared visually, and the images with varying thresholds of 2 and 3 above showed no clear improvement as compared with the prior conditions. Visual inspection involved identifying days with significant snow-cover as well as cloud areas that crossed from snow-covered areas to those without. The above conditions also did not appear to erroneously eliminate cloudy pixels. But the result was that existing thresholds, most importantly the 50% albedo threshold alone appeared to eliminate snow pixels, such that applying the threshold did not yield an improvement. It was left within the code for possible future use with a lower albedo threshold, but not utilized.

Below is one example of a visual comparison using these data sources. This example used a temperature factor of 21, and 3.9 reflectance of 0.057, with the test output image found in Figure 4.11. Figure 4.12 farther below shows snow cover for that day, taken from the NOHRS website. The prior day showed only a fraction of the snow cover, so this was fresh
snow in the northern-most sections of the study region, in relatively undeveloped areas. The 1-km resolution McIDAS GOES channel 1 image shown in Figure 4.13 (c) shows a larger section of this image; (a) and (b) show the close-ups of the examined area from the test run, and GOES image respectively, in which fresh snow cover is present in the upper half. These snow-covered areas do not show up as cloud in Figure 4.13 (a), and also are relatively dark in Figure 4.13 (b). While snowy pixels were eliminated with Snow Factor thresholds applied, the same pixels were eliminated using only those in Table 5. Please note that the state boundaries in the test image are misaligned, so the cloud outlines of the higher-resolution clouds should guide the comparison.
Figure 4.11: Example of a snow threshold test for December 2, 2004 showing visible albedo for with a temperature factor of 21, and 3.9 reflectance of 0.057.
Figure 4.12: NOHRS Regional Snow Analysis average showing snowpack temperature and elevation from December 2, 2004.
Figure 4.13: Shows (a) a close-up of a portion of Figure 4.11’s cloud test of the visible albedo image that corresponds with (b) a close-up of the corresponding GOES channel 1 image, which was taken from (c) the same GOES visible image at a smaller-scale resolution. Note that dark non-cloud areas in (a) and (b) include areas of fresh snow shown in Figure 4.12.

In conclusion, this visual comparison of snow-cover data with the cloud thresholds applied in this study showed that winter snow cover does not distort cloud results. It also leaves open a method for snow-contamination removal for future runs with a lower albedo threshold.

### 4.3 Climatologies and downwind data extraction

In addition to the NYC starting point, two control areas were designated, as described below. Data from areas one hour downwind of various radii was extracted for the NYC and rural areas. Climatologies in this study were intended to identify cloud properties that varied with urban aerosols and UHI conditions. Wind speed and direction data of one hour prior to the images was used for a one-hour downwind location designation, as described below in sections 4.5 and 4.6. A schematic in Figure 4.14 shows overlapping circular downwind areas, and that a different number of pixels are averaged over different points,
indicated by darker pink where there is more overlap. The actual data outcomes include many more wind directions. Numeric bar-plots in results average the entire downwind-area data set such that pixels with less data have less weight. The image plots in the results section show averages for each pixel such that edge pixels usually represent fewer data-points.
Climatologies were created for UHI and aerosol levels separately; each of those were further divided by season. Seasons were divided by June-August, September-November, December-February, and March-May. Output was created for three categories of UHI based on the six-station UHI calculation, and for three fine-mode AERONET aerosol categories, as
described below. All one-hour downwind-area climatologies were run for the following cloud categories, also summarized in Table 1:

- Warm-cloud albedo and cloud fraction
- Cold-cloud albedo and cloud fraction
- Warm-cloud SWIR at 3.9-µm for cloud particle-size information
  - The same warm-cloud fraction applies
- All-cloud results are warm and cold clouds together, and are not discussed separately

**For all above:**
- Image plots created
- Numeric data plots created
- Urban-rural difference plots for every data category

The warm-cloud pixels for SWIRR are the same as those used for warm-cloud albedo, so the same warm-cloud fraction applies to both.

Both image and data plots were created for all categories. All data plots are presented, while selected sets of the numerous image plots are included in the results section.
4.3.1 UHI data set and climatologies

The aim of the UHI climatologies is to detect the possible initiation of UHI-generated circulations or convective clouds. This is the result of destabilization due to UHI-thermal perturbation of the boundary layer. At the time of day of these images, one model predicts a change from a “build-up phase” to an “active phase” as clouds develop and precipitation is produced (Avissar 1989). This could be detected in the form of higher than expected urban-rural albedo, or cloud fraction. Large cloud-top particle would also be expected in convective clouds.

A visual inspection of many season’s data was used to designate cut-off points for high, medium, and low categories. The thresholds were a UHI of 1° Fahrenheit or less as low; between 1 and 2 medium, and above three degrees high. For the ten year study period, UHI data is further divided by season, and then for each season, into high, medium and low UHI. Climatologies were created from this seasonal data for each cloud category. As shown below in Figure 4.15, parallel climatologies were created for the UHI and aerosol datasets.
Figure 4.15: UHI, aerosol, and seasonal partitioning of the three cloud variables.

4.3.2 Aerosol data set and climatologies

Daily averaged AOD ($\tau_a$) at 500 nm was collected from 2000 through 2008 (see section 4.3). This data was divided into high, medium and low aerosol categories. Cutoffs were based on visual inspection of daily averaged data for categories which would all contain sufficient data for analysis. The AOD cutoff for the low category was below 0.2, medium between 0.2 and 0.6, and high above 0.6. AERONET measures column-integrated aerosol,
and while most is found in the PBL, cold clouds are more likely to interact with layers lifted into the free atmosphere.

### 4.4 NYC and two rural control locations

Two locations north and south of the urban study area were designated as control locations. NYC and the two control locations were the starting points for the downwind areas. Data downwind of these three points was used for mapped climatologies and a numerical analysis. Figure 4.16 summarizes the use of one urban and two control starting point locations, and of the downwind-area datasets extracted. Data is further subdivided seasonally.
Figure 4.16: Two sets of downwind data, for UHI and aerosol categories, are extracted for one hour downwind of the urban and control locations.

The control locations were a similar distance to the coastline as Central Park. Extraction and comparison of cloud fraction, albedo and SWIR reflectance data were evaluated. The control locations were both 130km from Central Park. This is 25 km farther than the 105km distance between the NYC central point and the locations used to measure UHI. While 105 km would have been more representative of ‘rural’ locations, ground weather stations are rarely located in truly rural environments. Sometimes they are in small regional airports outside of the urban areas; sometimes they are in small cities. Those
stations were the best approximation available for the purposes of measuring UHI, and did display sufficiently cooler temperatures from NYC to be useful in a UHI measure.

The designation of control locations at a greater distance was based on the dimensions of the NYC-urban region. New York City and the surrounding metropolitan region not only constitute a mega-city, but the many nearby cities make a designation of a representative “rural” site with respect to land-cover difficult. The selected control locations are open areas or very forested suburbs. These locations are more remote from nearby smaller cities so an extraction of data around them and downwind of them will be less influenced by effects of development and urbanization.

Control locations were also selected with the possibility of climatologies with no downwind designation, but of the areas surrounding the three points. Such climatologies would ideally include the 40-75km downwind area found in which downwind clouds effects have been found. A 130km distance of NYC to control regions allows for 65km (40.4 mi) radius data extraction without overlap between the two circles. This work only utilized data one hour downwind, and sufficient data was available such that much smaller radii were employed, allowing for better restriction to study area starting points. Areas of varying radii were extracted. The smallest radii with reasonable data quantity were considered most representative of starting points with urban and rural land cover features. A 5-km radius was possible for averages over all seasons, and for most UHI categories. For Aerosol, which encompassed a smaller dataset, higher area-radii were often more robust
datasets for seasonal averages. The downwind circle areas, in combination with wind data used to predict the downwind location capture short-term downwind effects.

Table 6 below shows the locations of the control starting-points.

<table>
<thead>
<tr>
<th></th>
<th>Distance to NYC (km)</th>
<th>Distance to NYC (miles)</th>
<th>Latitude, Longitude</th>
<th>Distance to shoreline (km, mi)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Central Park, NYC</td>
<td></td>
<td></td>
<td>40.783, -73.967</td>
<td>25.5, 15.9</td>
</tr>
<tr>
<td>South Control</td>
<td>130.47</td>
<td>81.07</td>
<td>39.681, -74.490</td>
<td>25.6, 15.9</td>
</tr>
<tr>
<td>North Control</td>
<td>129.59</td>
<td>80.52</td>
<td>41.494, -72.743</td>
<td>25.6, 15.9</td>
</tr>
</tbody>
</table>

The spherical law of cosines was used to find data within the radius of each point, including New York City. Test runs were performed to ensure it worked properly, and that latitude and longitude locations came up correctly. Various programs and online tools were used to designate the control points, based on distance from the New York City, distance from the shore, and most importantly degree of urbanization of the areas in question. Refer to Appendix D for details of this designation. Figure 4.17 below shows the selected locations. Both control starting-points were in highly vegetated regions. Figure 4.18 shows a ‘greenness’ map using normalized difference vegetation index (NDVI) data, a measure of vegetative cover, for the same locations for August 23, 2006, during the middle of the study period.
Figure 4.17: NYC central point in red and rural control locations in blue using a Google Maps satellite image.
Figure 4.18: Greenness map based on NDVI data for August 23, 2006 showing vegetation cover for the urban study region in the middle, and two control locations (USDA, 2015). Note: this NDVI data is scaled 0-255, though the actual range is -1 to +1 for 0-200; 201-255 are not vegetated.
4.5 Downwind data extraction

Data was extracted from 3.9-µm reflectivity and visible images downwind of NYC and two control stations north and south. The downwind locations were designated using ground station wind speed and direction.

Wind data from 16:00 UTC, one hour prior to that of the GOES images of 17 UTC (noon local time) was used. Ground stations collect data on the hour, or one or two minutes before or after the hour. Downwind locations were calculated starting at each of the three location in dark red in Table 7 below, and in red on the map. That calculation used the average wind speed and direction of the nearby ground stations seen in black in Table 7, and in turquoise on the map, Figure 4.19. As the weather stations were not equally distant from the study location, their data was weighted according to their distance. A Matlab function calculated the downwind latitude and longitude along a great circle. A schematic of the overlapping data-points used in numeric and mapped climatologies is shown in Figure 4.14. For the purpose of clarity this image shows all downwind areas moving in one general direction. While there are prevailing wind directions, as shown in section 4.6 below, winds in reality moved in many directions.
Table 7: Ground stations used for wind direction. The center points of the areas of study are in red: Central Park, in New York City, and two control locations, South and North. Labels at the right correspond to labels in Figure 4.19 below.

<table>
<thead>
<tr>
<th>Station/ Location</th>
<th>Latitude, Longitude</th>
<th>Map label</th>
</tr>
</thead>
<tbody>
<tr>
<td>Central Park, NYC</td>
<td>40.783, -73.967</td>
<td>C</td>
</tr>
<tr>
<td>JFK Airport</td>
<td>40.639, -73.762</td>
<td>K</td>
</tr>
<tr>
<td>La Guardia Airport</td>
<td>40.779, -73.880</td>
<td>L</td>
</tr>
<tr>
<td><strong>South Control</strong></td>
<td><strong>39.681, -74.490</strong></td>
<td><strong>S</strong></td>
</tr>
<tr>
<td>Atlantic City Intl</td>
<td>39.449, -74.567</td>
<td>A</td>
</tr>
<tr>
<td>South Jersey Rgnl</td>
<td>39.950, -74.850</td>
<td>J</td>
</tr>
<tr>
<td>McGuire AFB</td>
<td>40.017, -74.583</td>
<td>G</td>
</tr>
<tr>
<td><strong>North Control</strong></td>
<td><strong>41.494, -72.743</strong></td>
<td><strong>N</strong></td>
</tr>
<tr>
<td>Waterbury Oxford</td>
<td>41.483, -73.133</td>
<td>W</td>
</tr>
<tr>
<td>Meriden</td>
<td>41.509, -72.829</td>
<td>M</td>
</tr>
</tbody>
</table>
Figure 4.19: stations used for wind direction. Central Park (C), JFK Airport (K), La Guardia Airport (L), South Control (S), Atlantic City Intl (A), South Jersey Rgnl (J), North Control (N), Waterbury Oxford (W), Meriden (M)

Wind speeds aloft vary greatly from those at ground level, so this source can only approximate cloud movement. Nine tests were run to verify the accuracy of ground station downwind location designations. Each of the model runs were initialized at thee elevations at the same starting point, as seen in Table 8. Three dates were run for each of three starting points for the dates 2005/07/01, 2008/10/02, and 2010/03/01. Runs were compared with HYSPLIT EDAS 40km forward trajectories started at 0, 500, and 1000m
meters above ground level (NOAA Air Resources Laboratory). This is the Hybrid Single-Particle Lagrangian Integrated Trajectory, using the Eta Data Assimilation. Average differences were small when compared with trajectories begun at ground level, and went up to an average of 32 km when compared with those started at 1000 meters elevation. This maximum difference is less than the 40 km resolution of the HYSPLIT model, so this error was considered acceptable.

Table 8: Average difference in the downwind point predicted between ground weather station wind speed and direction data, and HYSPLIT EDAS 40 km downwind trajectory model output.

<table>
<thead>
<tr>
<th>Starting point</th>
<th>Average Difference, km</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Hysplit elevation start</td>
</tr>
<tr>
<td>North</td>
<td></td>
</tr>
<tr>
<td>NYC</td>
<td></td>
</tr>
<tr>
<td>South</td>
<td></td>
</tr>
<tr>
<td>Total</td>
<td></td>
</tr>
</tbody>
</table>

The initial downwind area radius was of 65 and 50-km but as it became apparent that there was sufficient data in smaller areas that initial output was excluded. Smaller downwind-area climatologies more narrowly draw from urban and rural land-cover starting points. Datasets of 5, 10, 20, and 35-km downwind areas are presented, with half in the appendices.
4.6 Image climatologies, and accompanying bar plots

The averaged mapped climatologies for cloud properties of SWIRR and albedo show the average value per pixel of the downwind areas. As the downwind circle does not always fall over the same pixels, pixels toward the edges of the area represent fewer pixels, and their values are less representative of the true average. An edge affect is apparent in discontinuities along outside edges of output images. The images serve to give a geographical representation of the outcome, and edge pixels often contain additional information about the pixel’s origin. The true averages are presented in bar plots, in which each data point is weighted equally.

The minimum albedo used for all images was 50%, so climatologies are of optically thick clouds only. Cloud fraction is presented in bar plots, in which the daily fraction of pixels with data is averaged over the number of days. Cloud fraction, or prevalence, mapped images are also presented, in units of ‘frequency per pixel per days of run.’ The number of times cloud was present in each pixel is divided by the total number of days of available data. In reference to mapped plots, cloud frequency, prevalence, and fraction are used interchangeably. Pixels on edges generally have fewer days of data, and in images containing many days of data represent pixels advected from the opposite side of the downwind data area. Large displacements also indicate high wind-speeds.

Prevailing wind direction can also be useful in the interpretation of the resulting image climatologies (described in the next section). Table 9 below is based on a long-running climatic wind dataset running from 1930-1996. These can be used as additional guides in
interpretations of the mapped images.

Table 9: Seasonal prevailing winds for each study region (NCDC, 1998)

<table>
<thead>
<tr>
<th></th>
<th>Winter</th>
<th>Spring</th>
<th>Summer</th>
<th>Fall</th>
</tr>
</thead>
<tbody>
<tr>
<td>NYC</td>
<td>NW</td>
<td>W-NW</td>
<td>SW</td>
<td>W</td>
</tr>
<tr>
<td>North</td>
<td>WSW</td>
<td>NW</td>
<td>WSW</td>
<td>WSW</td>
</tr>
<tr>
<td>South</td>
<td>WNW</td>
<td>W-WNW</td>
<td>SW</td>
<td>WNW</td>
</tr>
</tbody>
</table>
5 AEROSOLS AND CLOUD-TOP PROPERTIES

In numerical outcomes, differences in cloud-top properties were evaluated over a seven-year time period. Data was not available continuously for the AERONET dataset, and larger error bars coincide with categories for which data was limited. The error bars show standard error of mean plotted in both directions from the average. Averages and SEM are calculated using pixels meeting thresholds for the particular cloud type for all days of the category in question. Data for the three cloud categories are plotted, described in Sections 4.2 and 4.3, of warm cloud SWIRR, warm cloud visible albedo, and cold cloud albedo. Data was divided into four three-month seasons: DJF, MAM, JJA, and SON.

Alongside the above dataset, a cloud fraction for the same data is plotted in the section directly below. This shows the fraction of cloudy to cloud-free pixels from those areas. Again, all clouds presented have of a minimum of 50% visible albedo. The same warm cloud pixel locations are used for SWIRR and visible albedo. In the case of cloud fraction, the n used for the SEM bars is the number of days for which cloud fraction was calculated, so the error bars are larger though the same number of pixels are used as the same radius for the same category.

Locations one hour downwind were designated, as described in section 4.4. Data was extracted from areas of varying downwind-area radii as described in section 4.5, for NYC and rural starting points. Data from areas 65, 50, 35, 20, 10, and 5 kilometer downwind-
area radii were extracted and plotted, but only the 35, 20, 10 and 5 are shown. Half of the bar plots are in the appendix, and half are in the body of the results sections. Most 5 and 20-km results for the aerosol section can be found in Appendix E. Generalizations described here are usually apparent in the plots presented in the body, but include information plots in appendices. The AERONET data used is only available at the NYC location, though the rural locations have been shown to have lower average fAOD during the time of the study period. Days were divided into three levels of fine mode AERONET aerosol, labeled high, medium, and low, as described in Section 4.3.2.

5.1 Results: Aerosol Climatologies averaged over all seasons

Overall numerical results show greater warm-cloud SWIRR at higher fine-mode aerosol loadings, and greater urban-rural differences with higher fAOD. This can be seen in Figure 5.1. Figure 5.5 shows visible albedo results, and as with SWIRR there is an inverse relationship between albedo and fAOD. Cold clouds have a variable relationship with fAOD, but urban-rural differences have a distinct pattern. Cold clouds have higher albedo in rural locations on days with higher urban fAOD.
5.1.1 Warm Cloud Shortwave Infrared Reflectance at 3.9-µm

5.1.1.1 Numeric results (see cloud fraction for warm clouds for fraction results)

Figure 5.1 below shows shortwave infrared reflectance at 3.9-µm (SWIRR) values averaged over all seasons for all years. NYC has consistently higher reflectance than the rural areas. The differences are greatest on high fAOD days, and this difference triples as the diameter of the area with data is decreases from 50-km to 5-km. The differences for the high aerosol days are about 5% of the average urban value for the 50-km dataset, 7% for the 35-km, to 19% of the 5-km NYC value. Over this time period the lowest cloud-top SWIRR values are found when the highest fine-mode AERONET daily readings were recorded. As the downwind-area radius decreases the high-to-low difference increases for the rural areas, while it greatly decreases for NYC. Figure 5.7 shows warm cloud fraction for the 10 and 35-km downwind areas. The urban cloud fraction shows a more pronounced decrease with fAOD for smaller downwind areas. The urban area consistently has lower cloud fraction at high fAOD, and this difference increases from about 2.5% for the 50-km case (not shown) to an 87% difference in the 5-km case (in Appendix E).
Figure 5.1: Three aerosol levels and warm-cloud SWIRR (left row), and urban-rural differences (right row) from 2002-2008 over all seasons for 1-hr downwind areas of: 5-km, (a) and (b), 10-km (c) and (d), 20-km (e) and (f), and 35-km (g) and (h), radii showing standard error bars.
5.1.1.2 Mapped Image climatologies

Figure 5.2: Warm-cloud SWIRR (left row), and warm-cloud frequency per pixel per days of run (right row) for high aerosol one hour downwind of NYC and rural downwind areas from 2002-2008 for downwind areas of radii 5-km (a) and (b), 10-km (c) and (d), and 20-km and (e) and (f).
Figure 5.3: Warm-cloud SWIRR (left row), and warm-cloud frequency per pixel per days of run (right row) for medium aerosol one hour downwind of NYC and rural downwind areas from 2002-2008 for downwind areas of radii 5-km (a) and (b), 10-km (c) and (d), and 20-km and (e) and (f).
Figure 5.4: Warm-cloud SWIRR (left row), and warm-cloud frequency per pixel per days of run (right row) for low aerosol one hour downwind of NYC and rural downwind areas from 2002-2008 for downwind areas of radii 5-km (a) and (b), 10-km (c) and (d), and 20-km and (e) and (f).
Figure 5.2, Figure 5.3, and Figure 5.4 show the averaged SWIRR, and warm-cloud fraction images for the same data sets as the bar plots of corresponding downwind-area radii. Warm-cloud-frequency, SWIRR and albedo is the same cloud dataset.

From the cloud frequency images it is clear that the greater part of the data is derived from pixels over land. Averages of these values vary much more greatly with aerosol level than they do with whether they are over land or sea. The highest values in the 20-km high aerosol averaged image are in the 0.18-0.19 range, close to the average values in the low aerosol image. The high aerosol image has most of its pixels over land, and does not show a clear spatial pattern. The cloud fraction images show that pixels on edges represent a small proportion of the overall values. The medium aerosol images for higher downwind areas show little variation over the area. The corresponding low aerosol images do show a clear gradation with much higher values on the sea-side edges than those over land. Again edge pixels represent a small fraction of the overall data in that category, but show that clouds moving from the sea under low aerosol conditions have larger cloud-top droplet radii, while they are smaller in those moving from land.

The high aerosol 5-km dataset shows pixels almost exclusively over land, and is a small dataset; urban rural differences in SWIRR are apparent. While the northern control area in the 20-km plot shows higher cloud frequency, the average SWIRR values are on the same order as the south station.
Note that to allow for clearer viewing of gradations in values the color legend on these images is different for each figure. In the differenced bar plots axis values are again variable. Conversely, the non-differenced bar plots have fixed axis limits, making smaller differences sometimes harder to decipher, but allowing for easier comparison between plots.
5.1.2 Warm, cold and all-cloud visible albedo over all seasons

5.1.2.1 Numeric results

Figure 5.5: Three aerosol levels and warm, cold, and all-cloud visible albedo over all seasons from 2002-2008 for 1-hr downwind areas of (a) 10-km and (b) 35-km radii. See Appendix E for the 5 and 20-km areas.
Figure 5.5 shows values averaged over all seasons from 2002-2008 for each visible albedo cloud type, and for each AOD magnitude. Again, the warm cloud visible albedo image shows visible reflectance for the same set of pixels as the 3.9-μm data, and does not overlap with those in the cold designation, while ‘all clouds’ combine both sets. For all but the 5-km radius downwind area results warm cloud albedo, as with shortwave-infrared reflectance, has progressively higher values on days of lower fAOD. In the 5-km case the highest albedos for urban and rural are on the medium aerosol day. Urban albedo is lower in most cases than the rural, with the exception of the medium aerosol 5-km downwind area. The error bars show standard error of the mean.

For cold-cloud albedo some patterns emerge in the lower downwind-area radii results. Cold clouds in urban areas have a U-shaped response to fAOD, though at the higher downwind-radii the high aerosol days have the highest albedo. In the rural areas albedo is higher on days with higher urban aerosol, and more so in the smaller downwind radius areas. This difference is seen in the difference plots in Figure 5.6. Cold cloud urban albedo is always higher than rural on the low aerosol days, then becomes progressively much lower than rural on higher aerosol days; the 35-km high aerosol case is the only exception.

Last, as seen in the all cloud results in Figure 5.5, greater column fAOD varies inversely with visible albedo overall at the higher downwind areas. This overall result includes both the warm and cold datasets, but depending on the amount of warm or cold clouds, may represent more of one than the other.
5.1.2.2 Numeric results for urban-rural differences

Figure 5.1 and Figure 5.6 show urban-rural differences in SWIRR and visible albedo. Warm cloud-tops in the NYC region consistently reflect more at 3.9-µm than those of their rural counterparts, are of a much higher magnitude, and differences are much higher with higher CCNY fAOD. Differences also become more pronounced with narrower study areas. The warm-cloud visible albedo difference is negative in almost all cases, thought for high fAOD at 10-km the difference is positive.

For NYC cold cloud albedo the high-low aerosol difference is positive at larger downwind areas, and becomes strongly negative for smaller areas, as seen in Figure 5.5 and Figure 5.6. The sharply decreasing cold cloud urban-rural differences with increasing aerosol, and with lower-radius study areas is clearly seen in the differenced plots.
Figure 5.6: Urban-rural differences for three aerosol levels and warm, cold, and all-cloud visible albedo over all seasons from 2002-2008 for 1-hr downwind areas of (a) 10-km and (b) 35-km radii.
5.1.2.3 Numeric results for cloud fraction

![Graph showing cloud fraction for different aerosol levels and cloud types over seasons from 2002-2008 for 1-hr downwind areas of (a) 10-km (b) 35-km radii.]

Figure 5.7: Three aerosol levels and cloud fraction for warm, cold, and all-clouds over all seasons from 2002-2008 for 1-hr downwind areas of (a) 10-km (b) 35-km radii.

Warm-cloud fraction, in Figure 5.7, decreases with higher aerosol for urban downwind areas, while remaining steady in rural cases. Urban-rural differences are increasingly
negative with increasing aerosol levels, as seen in Figure 5.8. These differences also increase in magnitude with more narrowly defined study areas.

Cold cloud fraction is low in all cases, and increases slightly with higher fAOD in rural areas. Urban-rural differences, seen in Figure 5.8, follow a consistent pattern of higher urban cloud fraction for high fAOD, lower for medium, and slightly higher in all cases for low fAOD.
Figure 5.8: Urban-rural differences for three aerosol levels and cloud fraction for warm, cold, and all-clouds over all seasons from 2002-2008 for 1-hr downwind areas of (a) 10-km (b) 35-km radii.
5.1.2.5 Image climatologies

5.1.2.5.1 Warm-cloud visible albedo over all seasons

Figure 5.9: Warm-cloud visible albedo for high aerosol one hour downwind of NYC and rural downwind areas from 2002-2008 for downwind areas of radii 5-km (a), 10-km (b), and 20-km (c). Note the differences in scale at right.
Figure 5.10: Warm-cloud visible albedo for medium aerosol one hour downwind of NYC and rural downwind areas from 2002-2008 for downwind areas of radii 5-km (a), 10-km (b), and 20-km (c).
Figure 5.11: Warm-cloud visible albedo for low aerosol one hour downwind of NYC and rural downwind areas from 2002-2008 for downwind areas of radii 5-km (a), 10-km (b), and 20-km (c).

Low aerosol albedo climatologies show higher albedos on northern edges, such as in Figure 5.11, most likely from off-shore air masses. This is consistent with albedo gradations seen in other images, while the low albedo values seen on the western side of the urban area is relatively unusual. Gradations in cold-cloud albedo, seen in Figure 5.12, Figure 5.13, and Figure 5.14, are more variable. The much lower average urban albedo for high aerosol conditions seen in Figure 5.6 may be an influence of the southern control area in an image that has larger north-south differences than it does urban-rural. Visible albedo values in some images exceeding 100% are most likely due to the bidirectional reflection function angles at certain viewing/solar geometries.
5.1.2.5.2 Cold-cloud visible albedo over all seasons

Figure 5.12: Cold-cloud albedo (left row), and cold-cloud frequency per pixel per days of run (right row) for high aerosol one hour downwind of NYC and rural downwind areas from 2002-2008 for downwind areas of radii 10-km (a) and (b), and 20-km (c) and (d). (please note that the ‘Warm’ label on 20-km cloud frequency should read “Cold”)
Figure 5.13: Cold-cloud albedo (left row), and cold-cloud frequency per pixel per days of run (right row) for medium aerosol one hour downwind of NYC and rural downwind areas from 2002-2008 for downwind areas of radii 10-km (a) and (b), and 20-km (c) and (d).
Figure 5.14: Cold-cloud albedo (left row), and cold-cloud frequency per pixel per days of run (right row) for low aerosol one hour downwind of NYC and rural downwind areas from 2002-2008 for downwind areas of radii 10-km (a) and (b), and 20-km (c) and (d).

5.2 Results: Seasonal Aerosol Climatologies

5.2.1 Warm Cloud SWIRR and Cloud Fraction

5.2.1.1 Warm Cloud SWIRR

Figure 5.15 breaks down the warm cloud SWIRR and visible albedo into seasons. Despite the designation of AERONET cutoff values for high to low categories that allowed for a significant number of data points in each category, winter months tended to have much lower aerosol levels, and for the winter category there were no days with fine mode aerosol over 0.6. There are also larger error bars for some high aerosol categories, as well as for winter plots, since fewer days fell into those categories.
When comparing values within the same season in Figure 5.15 the same pattern as the seasonally averaged plot of increasing urban-rural differences with increasing fAOD is present in all seasons but winter. This is more apparent when the 5 and 20-km areas are included. There are however significant seasonal variations. Winter has the highest 3.9-μm reflectivity values, followed by fall, and summer had the lowest values. The seasonally averaged plot is most closely represented by the spring and fall seasons. The urban region usually has higher reflectance, as can be seen in the differenced plots, Figure 5.16. In the winter medium aerosol category however rural clouds have higher reflectance than the urban in all but the 10-km radius downwind areas.
Figure 5.15: Three aerosol levels and warm-cloud SWIRR from 2002-2008 for four seasons for 1-hr downwind areas of radii (a) 10-km, and (b) 35-km.

5.2.1.2 Warm Cloud SWIRR Urban-Rural Differences
Figure 5.16: Urban-rural differences for three aerosol levels and warm-cloud SWIRR from 2002-2008 for four seasons. One-hour downwind areas of radii (a) 10-km, and (b) 35-km.

5.2.2 Warm Cloud Visible Albedo and Cloud Fraction

5.2.2.1 Warm Cloud Visible Albedo

Seasonally separated warm-cloud visible albedo in Figure 5.17 is highly variable. Varying cases of lower albedo at higher aerosol levels account for that pattern in the overall averages. These figures do not show many regular patterns between seasons and aerosol levels. The urban-rural differences Figure 5.18 shows similarly variable outcomes, and also vary between downwind-area radii. A large negative urban-rural difference for medium aerosols in winter is a repeated seasonal pattern, though other seasons have large variability.
Figure 5.17: Three aerosol levels and warm-cloud visible albedo from 2002-2008 for four seasons for 1-hr downwind areas of radii (a) 10-km, and (b) 35-km.
5.2.2.2 Warm Cloud Visible Albedo Urban-Rural Differences

Figure 5.18: Urban-rural differences for three aerosol levels and warm-cloud visible albedo from 2002-2008 for four seasons for 1-hr downwind areas of radii (a) 10-km, and (b) 35-km.
5.2.2.3 Cloud Fraction for Warm Cloud Visible Albedo

Figure 5.19 shows that for warm-cloud fraction the summer season is most representative of the overall averages. The spring and fall seasons have an inverted parabola response. The urban cloud fraction increases from low to medium, though by less than the rural, then drops on the high days. The low aerosol days in winter have higher cloud fraction downwind of urban areas, but this difference is reversed in on medium days. The plots in Figure 5.20 show overall higher cloud fraction in rural compared with the urban area. There are also negative differences across different seasons with higher aerosol levels.
Figure 5.19: 2002-2008 warm cloud fraction averages for 1-hr downwind areas of variable-km radii for three aerosol classes, ordered by season: urban-rural differences.
5.2.2.4 Cloud Fraction for Warm Cloud Visible Albedo Urban-Rural Differences

Figure 5.20: Urban-rural differences for 2002-2008 warm cloud fraction averages for 1-hr downwind areas of variable-km radii for three aerosol classes, ordered by season.
5.2.3 Cold Cloud Visible Albedo

5.2.3.1 Cold Cloud Visible Albedo

The overall results showed higher albedo at higher aerosol levels for cold clouds, though the progression from high to low is not continuous. Figure 5.21 below shows very high variability between seasons. Summer season albedo increases with aerosol, while fall decreases. Figure 5.21(c) shows sharply increasing rural albedo on higher aerosol days across summer seasons. Values increase downwind of the urban area but remain far lower, and increase less at lower downwind areas. Urban-rural differences in Figure 5.20 show similarly inconsistent patterns between different seasons. Rural albedo tends to be higher on higher aerosol days in fall and summer; urban albedo is higher in the spring season.
Figure 5.21: 2002-2008 warm cloud visible albedo averages for 1-hr downwind areas of (a) 10, and (b) 35-km radii for three aerosol classes, ordered by season. See Appendix E 2.3.1 for more downwind areas. (c) shows the summer season only for 5 through 35-km downwind-radius areas.
5.2.3.2 Cold Cloud Visible Albedo Urban-Rural Differences

![Graph showing urban-rural differences for 2002-2008 cold cloud 3.9-um reflectivity visible albedo averages for 1-hr downwind areas of (a) 10-km radius and (b) 35-km radius for three aerosol classes, ordered by season. See Appendix E 2.3.2 for the 5 and 20-km areas.]

Figure 5.22: Urban-rural differences for 2002-2008 cold cloud 3.9-um reflectivity visible albedo averages for 1-hr downwind areas of (a) 10-km radius and (b) 35-km radius for three aerosol classes, ordered by season. See Appendix E 2.3.2 for the 5 and 20-km areas.
Figure 5.23: 2002-2008 cold cloud fraction averages for 1-hr downwind areas of (a) 10 and (b) 35-km radii for three aerosol classes, ordered by season: urban-rural differences. See Appendix E.2.3.2 for the 5 and 20-km areas.
5.2.3.4 Cloud Fraction for Cold Cloud Visible Albedo Urban-Rural Differences

Figure 5.24: Urban-rural differences for 2002-2008 cold cloud fraction averages for 1-hr downwind areas of (a) 10 and (b) 35-km radii for three aerosol classes, ordered by season.
5.2.4 Sample image seasonal climatologies

The summers season SWIRR in Figure 5.25 shows high geographic variability of values in the low fAOD 20-km image. Cloud frequency has increasing urban-rural differences in the number plots. The cloud frequency images in Figure 5.26 show much higher frequency in the northern areas for the 20-km plot on high and medium aerosol days. Albedo is presented alongside SWIRR in Figure 5.25, and is evenly distributed in the climatologies, with a few high values on inland edges. The 5-km plots in Figure 5.27 show cloudy pixels in a narrower area, and staying closer to their origin in high aerosol days compared with low, though the number of days more than doubled in low compared with high. The cloud frequency in Figure 5.28 show again higher clustering, or lower wind-speeds on high aerosol days, but this is most likely due to total data availability at the lower downwind areas.

(a)  (b)
Figure 5.25: Warm-cloud SWIRR (left row), and warm-cloud albedo (right row) for high (a) and (b), medium (c) and (d), and low (e) and (f) aerosol one hour downwind of NYC and rural downwind areas of 20-km radii for summer seasons from 2002-2008.
Figure 5.26: Warm-cloud frequency per pixel per days of run for high (a), medium (b), and low (c) aerosol one hour downwind of NYC and rural downwind areas of 20-km radii for summer seasons from 2002-2008.
Figure 5.27: Warm-cloud SWIRR (left row), and warm-cloud albedo (right row) for high (a) and (b), medium (c) and (d), and low (e) and (f) aerosol one hour downwind of NYC and rural downwind areas of 5-km radii for summer seasons from 2002-2008.
Figure 5.28: Warm-cloud frequency per pixel per days of run for high (a), medium (b), and low (c) aerosol one hour downwind of NYC and rural downwind areas of 5-km radii for summer seasons from 2002-2008.
5.3 Discussion: Aerosol Climatologies averaged over all seasons

5.3.1 Warm Cloud SWIRR, albedo, and cloud fraction

Warm cloud overall results for the SWIRR-aerosol relationship, seen in Figure 5.1, Figure 5.2, Figure 5.3, and Figure 5.4 are mixed, with evidence of a particle radius and an albedo aerosol effect, and a negative result for increased cloud lifetime (Albrecht) effects. Urban clouds have a consistently higher SWIRR difference than their rural counterparts, which increase with higher fAOD. Albedo, SWIRR, and cloud fraction and urban-rural cloud-fraction all decrease with fAOD.

Figure 5.1 shows that the aerosol-SWIRR urban response is weakly negative for the 5-km radius urban area. Though it is more negative with higher downwind areas, it is the reverse of the particle radius effect predicted by Twomey (1977). For larger downwind areas (and more so for those not shown) the outcome is of even lower SWIRR with increasing fAOD. According to the Twomey theory, particulates are expected to offer additional CCN sites for cloud droplets, resulting in polluted clouds with more, smaller droplets than their less polluted counterparts. Jin et al (2005) found weekly signals of AOT, rainfall, cloud effective radius, and liquid water path for NYC. Jin’s study found the weekly AOT maxima on the same days as the highest could effective radius, and the lowest liquid water path. Those results are consistent with outcomes of this work. Other studies, as described below, have also found reductions in droplet concentration and cloudiness with aerosol (Koren et al., 2008).
The urban areas were, however, much more reflective in the SWIRR than their rural counterparts, suggesting a reduced particle radius effect is taking place. This interpretation holds true under the assumption of a background urban aerosol level that is elevated above rural areas. There is a concentration of known anthropogenic sources of aerosols in the NYC-metropolitan region. Aerosol variation in this study is only measured in NYC, but as seen in Figure 2.6, using observations over most years of this study period, average AOD is much higher on average in the urban compared with rural designations used for this study. The greatly increasing urban-rural SWIRR differences with increasing fAOD is consistent with a microphysical effect on cloud-droplet radius. It is also illustrated below in seasonal results.

The above outcome is paired with consistently lower urban warm cloud fraction compared with rural, which is the reverse of the cloud lifetime effect. This difference also increases with increasing aerosol levels. A reduced cloud fraction result is in agreement with a recent finding in NYC of reduced rainfall with fine mode aerosol (Hosannah 2014), as well as Jin’s finding of liquid water path. Sections 2.1.1 and 2.1.4 describe absorbing aerosol effects, in which shallow cloud fraction decreases with increasing AOD (Koren et al., 2004, Feingold et al., 2005, Koren et al., 2008). Clouds can be reduced by stabilization of the atmosphere, or by heating of the cloud layer itself.

Koren’s 2008 study showed significant cloud fraction reduction by absorbing aerosols over the Amazon, in particular when data was limited to lower cloud fraction. The aerosol data-
set in this study is similarly limited by the AERONET measure to days with relatively low cloud cover. Lower cloud fraction in the aerosol dataset is also evident by comparison with cloud fraction levels in the UHI dataset.

Under non-precipitating conditions, clouds can also thin in response to increasing aerosol through a combination of droplet sedimentation (Bretherton et al., 2007) and evaporation-entrainment (Hill et al., 2009) described in section 2.1.5. Hill’s study was a large-eddy simulation with bin microphysics, and found a 10% decrease in liquid water path in response to increased aerosol due to sedimentation and evaporation-entrainment. The study was of warm clouds, though of clean marine stratocumulus. These processes could be a factor in the cloud fraction outcome of the present work if interactions with clean marine clouds account for urban-rural particle-size differences. A UHI effect may have an influence on these outcomes, though in this section those cases cannot be separated. If present, urban-heating enhanced cloudiness could mask even greater urban cloud fraction reductions than found in these results.
The third finding for overall averages is of decreased warm-cloud visible albedo with increasing aerosol in almost all downwind-radius categories, as seen in Figure 5.5. This is inconsistent with both the albedo and the cloud lifetime effects. Figure 5.6 shows that urban albedo was lower than rural in many categories, but the magnitude of difference was smaller than for cloud fraction and SWIRR, and in a few categories in was higher. Droplet sedimentation and evaporation–entrainment would result in clouds with lower optical depth and visible albedo. Overall results show that warm cloud SWIRR and albedo for urban and rural areas consistently decrease with increasing aerosol, but urban-rural albedo differences for these variables show no relationship.

Evaporation–entrainment, droplet sedimentation, and absorbing aerosol effects can be consistent with the highest aerosol readings having been found with relatively stable droplet size. As described below in this section, cloud droplets may become saturated with aerosol CCNs at lower urban aerosol loadings. From that point onward aerosol effects other than of the ‘albedo effect’ dominate cloud processes, and a higher aerosol loading may enhance their strength.

5.3.2 Aerosols and relative humidity

Various caveats should be considered when evaluating outcomes. Control locations with equivalent distance to the shore were selected, but the topography of the three areas is different, and could influence outcomes. While there may be consistent relationships between variables, causation cannot be taken for granted. The aerosol dataset can be
expected to over-represent clear skies in the urban area, as clear conditions overhead of the AERONET station are required for that measure. Cloud fraction results may be directly influenced by this factor. Correlations between the aerosol and cloud or precipitation may not indicate a cloud response to the aerosol (Boucher et al., 2013, Painemal and Zuidema, 2010). Aerosol loading is greatly influenced by air-mass history and origin, and by cloud and precipitation processes (Boucher and Quaas, 2013, Clarke et al., 1999; Petters et al., 2006; Anderson et al., 2009), and both are affected by meteorology (Engström and Ekman, 2010; Boucher and Quaas, 2013). Days in which higher aerosols are present tend to have common synoptic features, and high humidity in particular (Niranjan et al., 2004), which can influence cloud properties independently of aerosol loading (Painemal and Zuidema, 2010). It has also been found that wet weather increases fossil fuel combustion-related fAOD (Jeoung et al., 2014), and this effect may be compounded in urban and industrial areas. Water vapor affects both rain and aerosol optical depth as hygroscopic aerosols increase in size. As seen below in Figure 5.29, the sub-micron range of aerosols are particularly sensitive to relative humidity. The fAOD measure used in this study is of particles that average about 0.5 to 1-µm.

In the referenced studies, however, aerosol levels correlate with decreased cloud droplet radius and higher humidity, whereas smaller particle size was found in this study to vary with aerosol level mainly in comparison with the rural sites. As seen in Figure 5.30, using fAOD data from this study, and ground-station data for humidity, the variables for the urban location do not show a strong relationship. The best-fit curve was found for wind direction of 240-270°, with an $R^2 = 0.26$. The correlation in some studies may have been
due to errors in aerosol measurements from scattering by clouds in the vicinity of the measurement.

Figure 5.29: Size segregated surface aerosol mass distribution obtained for different surface humidity conditions in the ten cut-off size ranges (from Niranjan, 2004)
Figure 5.30: 2004 fAOD and relative humidity using CCNY AERONET and ground-station weather data. NCDC temperature and dew-point temperature were used for relative humidity.

Over the time scale of these climatologies, and given the proximity of the control locations, synoptic conditions would be expected to be similar across the three regions, such that the urban-rural difference is suggestive of an urban aerosol, or some other urban effect. So even if high humidity and other synoptic-scale weather variables do contribute to aerosol loading, urban-rural differences in SWIRR and cloud fraction should be a result of conditions particular to the urban site. While urban warm cloud fraction is significantly lower under high fAOD loading, rural clouds increase slightly. The interpretation would then be that to the extent that high humidity results in higher aerosol loading in the rural regions as well, the lower anthropogenic contribution to their content mix is associated with larger rather than smaller warm-cloud cloud-top droplet radii, and a higher warm-cloud fraction. If high aerosol days are high-humidity days in the rural sites this is not in agreement with the reduced droplet radius found in studies cited above. Decreased SWIRR
with increased fAOD is observed for higher downwind radii in the urban region as well, and to the extent that wider areas capture a greater fraction of non-urban starting points, the same interpretation could be made. Urban warm clouds, by contrast with rural, maintain relatively constant SWIRR/ cloud-top radii, which may have been larger in the absence of anthropogenic aerosols or other anthropogenic effects, but are unlikely driven by higher humidity.

5.3.3 Non-linear response and saturation

The AOD effect on cloud properties has been shown in many studies to be non-linear. Many observational studies find a strong effect at relatively low AOD concentrations. At AOD less than about 0.3 an increase in aerosol has been found to increase droplet concentration and decrease droplet size for constant liquid water (Boucher et al., 2013), though the effects may be localized. At high AOD droplet concentration tends to saturate (Boucher 2013, Verheggen et al., 2007). Verheggen (2007) similarly found that the activated fraction of aerosols in warm clouds decreased with increasing aerosol number concentration. The susceptibility of the activated particles to the total particle number was largest in clean conditions for warm clouds. The relatively unchanged reflectance with fAOD for urban clouds with the small downwind radius is consistent with this observation. Continuously elevated aerosols in the NYC-met region, in additional to the natural aerosol present could mean that these effects take place at the lowest fAOD loadings. Indeed, the low fAOD designation in this study was for a maximum of 0.2, and in Figure 2.1 (Section 2.1.4) cloud fraction sharply increases before AOD of 0.2, after which it steadily declines.
Koren’s 2012 study showed rainfall intensification, not suppression, was statistically correlated with increasing aerosol levels within the lowest aerosol levels, as described in section 2.1.6. This could be an alternative explanation for the higher downwind-radius results of this study for SWIRR and aerosol found in Figure 5.1, and also seen in seasonal results. In that study relatively low aerosol levels have the greatest impact on precipitation; the upper cut-off for AOD is 0.3. The 13:30 local time of this study makes it a good comparison with this one, as clouds that precipitate at 1:30 local time may show droplet size changes earlier, at noon local time. The present study uses 0.2 and 0.6 as cut-offs for low, medium and high fAOD, and droplet size levels off for the 5-km downwind area between medium and high fAOD, but decreased urban-rural cloud fraction continues at higher aerosol levels. Koren’s study could also explain the cloud fraction results in figures Figure 5.7 and Figure 5.8 as the cloud reduction due to raining out. More likely, this enhancement only takes place at lower fAOD levels, and evaporation-entrainment and other aerosol effects dominate at higher fAOD.

The increasing urban-rural SWIRR difference with fAOD can be explained by saturation effects. A possible interpretation is that in weather conditions associated with higher fAOD correlate with heavier cloud cover and increased likelihood of precipitation in rural regions. This idea would be borne out by the SWIRR and cloud fraction results, but not by the warm-cloud visible albedo, which drops between medium and high fAOD days. It is in agreement with cold-cloud albedo outcomes. The urban-rural reflectance difference could be evidence of a continuous effects on clean marine clouds even after they have saturated
clouds of land-based origin. This effect could be in proportion to higher aerosol levels even if saturation effects do take place on other clouds.

The cloud fraction outcome can then be explained again as a combination of droplet evaporation–entrainment, and aerosol sedimentation, and absorbing aerosol effects which take place in more polluted urban clouds. Aerosol released in cloud evaporation, along with unsaturated aerosol could also cause direct effects of absorption of solar radiation and local heating resulting in cloud dissipation.

5.3.4 Seasonal results

SWIRR-aerosol results break down into distinct seasonal patterns, with spring and fall often being similar, and most representative of the overall averages. The summer season run has more high aerosol data than other seasons, at 50 days with data in both urban and rural sites. In the 5 and 10-km downwind area locations urban SWIRR is fairly flat with changing aerosol levels while the rural values increase. Higher temperatures and longer days in the summer allow for more aerosol-sun interactions, and greater production of secondary aerosols. As described in Section 2.1.8, under peak demand on hot summer days in NYC energy consumption not only increases, but shifts towards more older, more polluting local power plants in order to meet the total regional demand, resulting in even higher aerosol levels.
Data in smaller downwind-area radii datasets is drawn from more exclusively urban and rural downwind starting points. For this reason patterns that are evident across downwind-area radii, but become more pronounced with smaller downwind areas are suggestive of an urban effect. This is true of summer SWIRR differences, as highlighted in red in Figure 5.31. Note both a continuously emerging pattern, as well as an increase in magnitude of the difference between the high at 35-km compared with 5 at the highest aerosol level. The same pattern is evident in both the fall and spring seasons, circled in green. The benefit of comparing various downwind-area radii is also illustrated here. Where the datasets are more limited, the lowest study area in which the pattern is apparent best illustrates the relationship. These figures together demonstrate that the same outcome in over-all averages is representative of a consistent pattern.

Summer warm-cloud visible albedo in the rural areas does not show a consistent response to aerosol levels, so an albedo effect is not demonstrated in this dataset, nor in urban-rural albedo differences. Albedo at higher downwind areas have some correspondence with variation in urban SWIRR. Cloud fraction increases for the rural cases, and decreases slightly for the urban region at the smaller downwind radii, as it does for overall results. An urban effect of reduced cloud fraction is consistent with a decreased albedo-SWIRR response in downwind areas more narrowly taken from the urban area. Again evaporation-entrainment, droplet sedimentation, and absorbing aerosol effects best explain these outcomes.
Summer aerosol mapped climatology results are in section 5.2.4. A saturation effect may be evident in differences in variability of SWIRR between lower and higher aerosol climatologies. In Figure 5.25 the urban downwind 20-km radius low aerosol image is more variable than on medium and high days.

Figure 5.31: Increasing urban-rural SWIRR-difference for summer with decreasing downwind radius. The same increasing response is also seen in the spring 10-km-radius area, and in the fall 20-km-radius area.

Spring, summer and fall seasons show a distinct inverse relationship in the variation of SWIRR and cloud fraction with aerosol levels, and even more so in their urban-rural differences, as shown in Figure 5.32. This relationship is evident at larger downwind-area
radii but becomes stronger as with smaller downwind areas. This outcome strengthens an interpretation of cloud fraction reductions as a response to microphysical effects on the cloud itself, and is consistent with cloud-top evaporation and entrainment effects. It neither supports nor detracts from the possibility that absorbing BC aerosols result in additional cloud dissipation through semi-direct radiative effects. As described in Section 2.1.8, black carbon aerosols are an important component of the NYC air pollution mix, such that absorbing aerosols likely exert an influence on cloud effects; BC itself can also serve as CCN in microphysical effects.
5.3.3 **Albedo and surface-cover effects**

In both the aerosol, and in the following UHI climatologies, minimal albedo effects are detected in numerical outcomes. Mapped climatologies, however, reveal relationships between SWIRR, albedo, and wind direction. Figure 5.33 shows the mapped climatology of an average over all seasons. Warmer seasons are over-represented in over-all averages as the summers have more data available. Prevailing winds are westerly for the summer season (Table 9). As seen in Figure 5.33, the off-shore air masses exhibit higher SWIRR.
and lower albedo. The higher albedo pixels are found most directly downwind of the sea, and generally have the lowest SWIRR values. Where a some edge pixels have both high or low albedo and 3.9-reflectance, their northeastern or southwestern location could have had their origins both in land and sea. Apart from edge pixels, note within the areas the east-west gradation in the urban and southern areas, and a north-south gradation in the northern station, in which the coastline runs more east-west. In UHI outcomes this pattern is both stronger and more prevalent. Onshore air masses can be expected to have higher moisture levels, and lower pollution, consistent with these outcomes. The albedo effect would predict the pixels with high SWIRR to exhibit relatively high albedo, the opposite of the inverse relationship found in this and other comparisons. This outcome could be explained by cloud dissipation effects of aerosols seen in this section, though it could also be explained by comparatively lower liquid water content of off-shore compared with on-shore clouds.

As with the summer-season mapped climatologies which had higher variability in the low aerosol category, a saturation effect, or higher sensitivity to aerosol at lower fAOD levels is evidenced by more distinct SWIRR and albedo gradations on days with low aerosol measurements compared with medium and high. The medium aerosol climatologies for all seasons do show strong variability, but less distinct gradations showing differences between off-shore and sea-breeze air-masses.
Figure 5.33: Inverse relationship between albedo and SWIRR, and the influence upwind air-masses from land compared with sea for low aerosol days. (a) Low aerosol warm-cloud SWIRR, (b) this dataset is characterized by higher overall SWIRR, (c) corresponding warm-cloud visible albedo, and (d) relatively small 3.9-μm urban-rural differences.
5.3.4 Cold clouds

As with SWIRR, the urban-rural differences for cold cloud albedo have a stronger response to fAOD than do absolute values. While urban albedo follows a U-shaped pattern with increasing aerosol, rural albedo increases consistently. The urban-rural difference becomes increasingly negative in every case in all but the 35-km radius area, as can be seen in Figure 5.6. The difference is positive for low aerosol, and the largest negative difference is for high fAOD. In the seasonal differences in Figure 5.22, the only exception to this pattern is the spring season. The spring season maintains a U-shaped urban-rural plot, but between the 35 and 5-km-radius area shifts strongly in the direction of a negative response, with the low doubling and the high roughly halving. Thus an inference can be made of an urban aerosol effect of decreased albedo, especially as this pattern becomes stronger with smaller downwind areas, as it does for the over-all average. This is true despite the fact that cold clouds are higher and less connected with ground-level variables, and more likely to have been advected from distant areas. The urban column aerosol measurement includes pollution carried aloft into the free atmosphere. The outcome could include contributions from remotely-sourced aerosols, while local urban aerosols are also incorporated into higher clouds at higher rates in the urban area.

As described in 2.1.7 more available ice nuclei (IN) particles can reduce the size of ice particles for the same water content as they do for liquid droplets, also increasing albedo. While clean clouds may mainly experience increased albedo (Verheggen 2007) the “cloud glaciation effect” can also ultimately result in reduced cloud lifetime in polluted clouds.
These processes are consistent with cold cloud results in this study, in which overall albedo increases with aerosol, but urban aerosols are associated with increasingly depressed albedo relative to urban locations with increasing aerosol. An anthropogenic aerosol mix including black carbon in particular can also contribute to a balance of cloud reduction (Lohmann, 2002). Regional synoptic conditions, including ones in which aerosol loading in the control areas is higher on days with high urban aerosol measures can account for the over-all increases in albedo with aerosol. Gryspeerdt (2012) found no significant sensitivity of particle number to AOD in mixed-phase clouds (deep convective and anvil cirrus).
6 URBAN HEAT ISLAND AND CLOUD-TOP PROPERTIES

6.1 Results: UHI Climatologies averaged over all seasons

6.1.1 Warm Cloud Shortwave Infrared Reflectance at 3.9-μm: numeric results

Please note: warm cloud-fraction results (for the same set of clouds) can be found in the warm-cloud visible albedo section. The 10 and 35-km downwind radius outcomes where not presented in the body can be found in appendices.

Figure 6.1 shows a consistently positive relationship between SWIRR and UHI for both urban and rural locations which held true for all downwind-area radii. Average values changed little between the 35 to the 5-km radius outcomes. High UHI values stayed in the 0.18 SWIRR range, and lows throughout were slightly above 0.14. Urban-rural difference values were positive in all cases, but values changed in pattern and magnitude with the downwind area size. Urban-rural differences for low UHI conditions changed the least with downwind area size, with a maximum at the 10-km radius of close to 0.005. Differences increased greatly for high and medium UHI as the downwind radius increased. The smallest difference was for the high UHI at the 35-km downwind radius area, and the largest were for high and medium UHI at the 5-km downwind areas. The magnitude of differences are smaller than those found in individual seasons (Figure 6.16), which exceed 0.013. For comparison, differences in the aerosol dataset exceed 0.030 (Figure 5.1 and Figure 5.16).
The corresponding UHI mapped climatologies, Figure 6.2, Figure 6.3, and Figure 6.4 show the 20, 10, and 5-km areas respectively. A striking feature of the medium UHI images in particular, as well as those of low UHI is the strong gradation towards higher SWIRR from west to east. Note again that the color scale changes with each image, and the median values of the high UHI image are higher than for medium and low. The high UHI images show a similar pattern, but less dramatic gradation, and still show the highest values on eastern edges. The urban downwind areas throughout the map images have higher frequency clusters slightly south-west, over Staten Island. The 5-km medium and high UHI bar plots had the highest urban-rural difference. The corresponding map plots again had a higher-frequency cluster over Staten Island, but average values there were in the 0.13 range. In high UHI plots the highest SWIRR values, exceeding 0.35, are seen for the urban regions only in the 10 and 20-km downwind areas. Greater dispersal towards the south-east in urban compared with rural areas, however, is seen throughout the medium and high UHI climatologies.

The 10 and 20-km medium images have higher frequencies slightly farther south-west, where there were lower reflectance averages. The low UHI outcomes did not vary as much in range between 20 to 5-km area radii, but also may show that a larger frequency over westward areas corresponded to lower SWIRR differences.
6.1.2 Warm cloud map climatologies: SWIRR, cloud frequency, and albedo

6.1.2.1 Warm-cloud SWIRR and warm-cloud fraction

6.1.2.1.1 High
Figure 6.2: Warm-cloud SWIRR (left row), and warm-cloud frequency per pixel per days of run (right row) for high UHI one hour downwind of NYC and rural downwind areas from 1999-2009 for downwind areas of radii 5-km (a) and (b), 10-km (c) and (d), and 20-km and (e) and (f).
Figure 6.3: Warm-cloud SWIRR (left row), and warm-cloud frequency per pixel per days of run (right row) for medium UHI one hour downwind of NYC and rural downwind areas from 1999-2009 for downwind areas of radii 5-km (a) and (b), 10-km (c) and (d), and 20-km and (e) and (f).
6.1.2.1.3 Low

Figure 6.4: Warm-cloud SWIRR (left row), and warm-cloud frequency per pixel per days of run (right row) for low UHI one hour downwind of NYC and rural downwind areas from 1999-2009 for downwind areas of radii 5-km (a) and (b), 10-km (c) and (d), and 20-km (e) and (f).
6.1.2.2 Warm-cloud visible albedo

Note: the same cloud frequency as in SWIRR applies.

6.1.2.2.1 High

(a) (b) (c)
Figure 6.5: Warm-cloud visible albedo for high UHI one hour downwind of NYC and rural downwind areas from 1999-2009 for downwind areas of radii 5-km (a), 10-km (b), and 20-km (c), (d), and (e). Erroneous values in the 20-km image are due to the addition of overlapping pixels; separated images of the same data are in (d) and (e).
6.1.2.2.2 Medium

Figure 6.6: Warm-cloud visible albedo for medium UHI one hour downwind of NYC and rural downwind areas from 1999-2009 for downwind areas of radii 5-km (a), 10-km (b), and 20-km (c).
6.1.2.2.1 Low

Figure 6.7: Warm-cloud visible albedo for low UHI one hour downwind of NYC and rural downwind areas from 1999-2009 for downwind areas of radii 5-km (a), 10-km (b), and 20-km (c).

6.1.3 Warm, cold and all-cloud visible albedo over all seasons

Rural albedo, as seen in Figure 6.8 increases with increasing UHI throughout all downwind-area sizes, though only by about 1%. Urban warm-cloud albedo has a variable response, with the highest values in medium UHI conditions, and the lowest for low UHI. Resulting
urban-rural differences, seen in Figure 6.9, have a repeated inverse-parabola response to UHI, with the highest values on medium UHI days. The sign of the low-UHI changes from positive to negative from the 20-km to the 5-km area.

A comparison of warm-cloud SWIRR map images with those of albedo shows the lowest range of albedo corresponding with the highest SWIRR values, as seen on the south-eastern edge of the 10-km high UHI images. SWIRR values are in the 0.3 range where albedo is at a minimum, in the 50-60% range, while on the south-western edge very low SWIRR in the 0.13 range coincide with nearly 100% albedo. The highest-frequency downwind area, a few miles southwest over New-Jersey and Staten Island, had SWIRR values in the 0.2 range, and albedo in the low 70’s.

The highest frequency area for medium UHI had SWIRR values of 0.15, and albedo again in the mid-70’s. Highs in SWIRR values for this image in the southeast side were 0.23 corresponding with albedo minima in the 50’s. The northwestern edge had SWIRR values of 0.11 SWIRR, and variable albedo including many very high values approaching 100%, along with some in the 60’s. For the highest frequency pixels in low UHI, SWIRR 0.2 correspond with 72% albedo.

Warm-cloud fraction was much higher over all in the UHI climatologies, a result of a bias in days AERONET aerosol measurements were available. Cloud fraction more than doubles between a low average in the 0.1 range under high UHI regimes, to highs reaching 0.25 with low UHI (Figure 6.10). Low UHI days consistently had negative urban-rural cloud
fraction, and medium days tended to have the highest positive difference, as seen in Figure 6.11. In high UHI conditions the urban-rural warm-cloud fraction difference was consistently positive but lower.

While albedo varies much less than cloud fraction with UHI. Their urban-rural differences of these both have an inverse-parabola response in some downwind areas, but in seasonal outcomes this is only consistent in albedo. Albedo and cloud fraction urban-rural difference generally vary inversely with the SWIRR difference for the 10 and 20-km areas. For the 5-km downwind area SWIRR and cloud fraction differences vary together

**Cold clouds**

Cold cloud albedo downwind of the rural area decreased with increasing UHI, and had a U-shaped curve for rural areas as seen in Figure 6.8. The result for urban-rural differences, in Figure 6.9, is an inverse parabola, similar to that for warm clouds.

Cold cloud fraction changed very little with downwind-area radius, as with warm-cloud fraction. High and medium remained in the 0.12 range, and the low at near 0.16 for rural and 0.17 for urban. Urban-rural differences were u-shaped, and mainly positive, and also varied very little with downwind area. They were greatest for the low UHI cases, about half as great for the high, at 0.005-0.007, and were close to zero for middle UHI days.
Cold-cloud visible albedo map images, Figure 6.12, Figure 6.13, Figure 6.14, show highs from stronger onshore winds on the west side of the images on low UHI and high days, and a gradation towards lower albedo values in offshore pixels. The middle UHI images had a more irregular pattern. Cloud frequency images in the right row show that while edge pixels are most reliably indicative of the direction of origin, they also represent the fewest days of data. In the case of cold cloud outcomes, wind-speed and direction measured at ground level and used to predict the downwind location are less accurate. Wind speeds increase greatly at higher altitudes, and if the direction is similar, then values on the downwind side of prevailing winds in the images may be more representative of the real outcomes.
6.1.3.1 Numeric results

Figure 6.8: Three UHI levels and warm, cold, and all-cloud visible albedo over all seasons from 1999-2009 for 1-hr downwind areas of (a) 5-km (b) 20-km radii. See Appendix F for the 10 and 35-km areas.
6.1.3.2 Numeric results for urban-rural differences

(a)

(b)
Figure 6.9: Urban-rural differences for three UHI levels and warm, cold, and all-cloud visible albedo over all seasons from 1999-2009 for 1-hr downwind areas of (a) 5-km, (b) 20-km, and (c) 35-km radii. See Appendix F for the 10 radius.
6.1.3.3 Numeric results for cloud fraction

Figure 6.10: Three UHI levels and cloud fraction for warm, cold, and all clouds over all seasons from 1999-2009 for 1-hr downwind areas of (a) 5-km (b) 20-km radii.
6.1.3.4 Numeric results for cloud fraction urban-rural differences

Figure 6.11: Urban-rural differences for three UHI levels and cloud fraction for warm, cold, and all clouds over all seasons from 1999-2009 for 1-hr downwind areas of (a) 5-km (b) 20-km radii.
6.1.4 Cold cloud image climatologies

6.1.4.1 High

Figure 6.12: Cold-cloud albedo (left row), and cold-cloud frequency per pixel per days of run (right row) for high UHI one hour downwind of NYC and rural downwind areas from 1999-2009 for downwind areas of radii 5-km (a) and (b), and 10-km (c) and (d).
Figure 6.13: Cold-cloud albedo (left row), and cold-cloud frequency per pixel per days of run (right row) for medium UHI one hour downwind of NYC and rural downwind areas from 1999-2009 for downwind areas of radii 5-km (a) and (b), and 10-km (c) and (d).
Figure 6.14: Cold-cloud albedo (left row), and cold-cloud frequency per pixel per days of run (right row) for low UHI one hour downwind of NYC and rural downwind areas from 1999-2009 for downwind areas of radii 5-km (a) and (b), and 10-km (c) and (d).
6.2 Results: Seasonal Urban Heat Island Climatologies

6.2.1 Warm Cloud SWIRR and Cloud Fraction

6.2.1.1 Warm Cloud SWIRR

In Figure 6.15 the summer season has much lower SWIRR values than other seasons, and almost no variability with UHI. Urban values are continuously higher than rural by large margins. The 5-20-km runs all show about a 30% greater reflectance in urban compared with rural areas. In the aerosol dataset summer values are also lower but less so compared with other seasons; urban-rural difference are always positive but vary with downwind area. As in the aerosol outcomes, this property is independent of cloud fraction, which drops dramatically with UHI.

SWIRR increases with UHI in the spring and fall seasons in both urban and rural locations. This relationship is not present in summer or winter outcomes. Figure 6.16 shows SWIRR urban-rural differences, again with large seasonal variations. Summer has the most consistent positive urban-rural differences; differences are mainly positive across seasons, and negative differences are small. With the exception of winter, high UHI conditions consistently have positive urban-rural differences. Winter generally has a inverse-parabola response, with the low UHI generally positive, and the high UHI close to or below zero. The highest urban-rural differences for all seasons take place under middle UHI conditions in the winter season. For the spring and fall seasons medium UHI conditions have low urban-
rural differences. Both have consistently positive differences for high UHI. In fall low UHI is consistently positive, while it is low or negative in the spring season.

For winter spring, and fall urban high UHI days had the highest SWIRR for the season. The corresponding high UHI days, in Figure 6.17, have the lowest albedos in winter and fall, and the middle values for spring. For rural areas spring and fall days have a lower peak SWIRR for high UHI days, slightly lower than the seasonal high for medium UHI. The corresponding rural albedo values are the lowest for the fall season, and the second-lowest in spring.

For spring, summer and fall cloud fraction, in Figure 6.19, steeply decreases with increasing UHI across all downwind area sizes, varying inversely with the SWIRR pattern for spring and fall. The urban-rural differenced values for cloud fraction and SWIRR, in Figure 6.18 and Figure 6.20, often vary inversely for the spring season, while they vary in the same direction in winter. Summer SWIRR differences are consistently positive, while cloud fraction differences for the summer are consistently either very low or negative, though variations in magnitude don’t correspond to one another.

In Figure 6.16 and Figure 6.18 spring high UHI conditions, positive SWIRR urban-rural differences corresponded with more negative albedo differences. For spring medium UHI days both SWIRR and albedo became less negative with smaller downwind areas. Across the winter numerical outcomes SWIRR differences generally often correspond with the sign of albedo. In winter there is also a general correspondence in the sign of urban-rural
differences of SWIRR and cloud fraction. Last, the winter season has the most positive albedo and cloud fraction urban-rural differences.

A correspondences between high SWIRR and low albedo is evident across most of the mapped images, seen clearly in the middle UHI image for winter 20-km climatologies, in Figure 6.27; parts of the low UHI image are exceptions. The area with greatest reflectance for high UHI days corresponded to an albedo of 72%, and a SWIRR of 0.22. As in other cases lower SWIRR is observed in pixels farther inland, while offshore easternmost pixels have the highest values. The pattern is true for all three downwind areas in the medium UHI image in Figure 6.27 but a greater proportion of high SWIRR is evident in the urban pixels, as reflected in the higher average shown in the bar plots (Figure 6.15 and Figure 6.16). The same inverse correspondence can be seen with a more careful comparison of pixels in the winter 5-km images, in Figure 6.25.

The same inverse correspondence is again found throughout the spring season. In Figure 6.29, the 5-km mapped climatologies, the medium and low climatologies have a clearer east-west gradation, but the a pixel-by-pixel inverse correspondence is also present in the high UHI images. In Figure 6.31 the spring 20-km downwind areas demonstrate the pattern most dramatically in the medium UHI category; two erroneous pixels in the 20-km SWIRR image distort the legend, but the same pattern is present there too. The 5-km downwind-areas in the fall season's Figure 6.37, and the 20-km areas in Figure 6.39 again show an inverse SWIRR and visible albedo relationship.
Winter 20-km mapped images in Figure 6.28 are a good example of higher wind-speeds associated with low UHI regimes, as seen in cloud frequency dispersal between the three images.

The summer season, as seen in numeric outcomes in Figure 6.15 and Figure 6.17, have the least variability across UHI levels in both SWIRR and albedo. An inverse correspondence between these variables is still present, but across the downwind areas a east-west gradation is least apparent in both the 5-km areas, Figure 6.33, and even less in the 20-km-radius areas in Figure 6.35.
Figure 6.15: Three UHI levels and warm-cloud SWIRR from 1999-2009 for four seasons for 1-hr downwind areas of radii (a) 5-km, and (b) 20-km.
6.2.1.3 Warm Cloud SWIRR Urban-Rural Differences

Figure 6.16: Urban-rural differences for three UHI levels and warm-cloud SWIRR from 1999-2009 for four seasons for 1-hr downwind areas of radii (a) 5-km, and (b) 20-km.
6.2.2 Warm Cloud Visible Albedo and Cloud Fraction

6.2.2.1 Warm Cloud Visible Albedo

Figure 6.17: Three UHI levels and warm-cloud visible albedo from 1999-2009 for four seasons for 1-hr downwind areas of radii (a) 5-km, and (b) 20-km.
6.2.2.2 Warm Cloud Visible Albedo Urban-Rural Differences

![Graph showing urban-rural differences in warm cloud visible albedo across seasons.](graph.png)
Figure 6.18: Urban-rural differences for three UHI levels and warm-cloud visible albedo from 1999-2009 for four seasons for 1-hr downwind areas of radii (a) 5-km, (b) 20-km, and (c) 35-km.
6.2.2.4 Cloud Fraction for Warm Cloud Visible Albedo

Figure 6.19: Three UHI levels and warm-cloud fraction from 1999-2009 for four seasons for 1-hr downwind areas of radii (a) 5-km, and (b) 20-km.
6.2.2.5 Urban-rural differences for warm cloud fraction

Figure 6.20: Urban-rural differences for three UHI levels and warm-cloud fraction from 1999-2009 for four seasons for 1-hr downwind areas of radii (a) 5-km, and (b) 20-km.m.
6.2.3 Cold Cloud Visible Albedo and Cloud Fraction

Decreasing overall visible albedo with increasing UHI can be seen most strongly in the summer and fall seasons in Figure 6.17. In the winter season only the urban location has this pattern; it is weakest in the spring season, especially for the lower downwind areas. Figure 6.18 shows large positive urban-rural differences for medium and low UHI in winter, and for medium UHI in fall. There are large negative differences in summer high UHI conditions. The spring, summer and fall seasons, in Figure 6.19 have consistently decreasing cloud fraction with increasing UHI, and a U-shaped response in the winter season. Figure 6.20 shows consistently positive urban-rural differences in the winter and fall seasons, with the greatest differences under low UHI conditions. These differences become smaller with decreasing study area for the fall difference, and larger for winter low UHI.
6.2.3.1 Cold Cloud Visible Albedo

Figure 6.21: Three UHI levels and cold-cloud visible albedo from 1999-2009 for four seasons for 1-hr downwind areas of radii (a) 5-km, and (b) 20-km.
6.2.3.2 Cold Cloud Visible Albedo Urban-Rural Differences

![Graph showing urban-rural differences for three UHI levels and cold-cloud visible albedo from 1999-2009 for four seasons for 1-hr downwind areas of radii (a) 5-km, and (b) 20-km.]

Figure 6.22: Urban-rural differences for three UHI levels and cold-cloud visible albedo from 1999-2009 for four seasons for 1-hr downwind areas of radii (a) 5-km, and (b) 20-km.
6.2.3.3 Cloud fraction for cold clouds

Figure 6.23: Three UHI levels and cold cloud fraction from 1999-2009 for four seasons for 1-hr downwind areas of radii (a) 5-km, and (b) 20-km.
Figure 6.24: Urban-rural differences for three UHI levels and cold-cloud fraction from 1999-2009 for four seasons for 1-hr downwind areas of radii (a) 5-km, and (b) 20-km.
6.2.4 Selected image climatologies

6.2.4.1 Winter 5-km radius downwind areas

(a) (b) (c) (d) (e) (f)
Figure 6.25 Warm-cloud SWIRR (left row), and warm-cloud albedo (right row) for high (a) and (b), medium (c) and (d), and low (e) and (f) UHI one hour downwind of NYC and rural downwind areas of 5-km radii for winter seasons from 1999-2009.

Figure 6.26: Warm cloud frequency per pixel per days of run for high (a) medium (b), and low (c) UHI one hour downwind of NYC and rural downwind areas of 5-km radii for winter seasons from 1999-2009.
6.2.4.1 Winter 20-km radius downwind areas

Figure 6.27: Warm-cloud SWIRR (left row), and warm-cloud albedo (right row) for high (a) and (b), medium (c) and (d), and low (e) and (f) UHI one hour downwind of NYC and rural downwind areas of 20-km radii for winter seasons from 1999-2009.
Figure 6.28: Warm cloud frequency per pixel per days of run for high (a) medium (b), and low (c) UHI one hour downwind of NYC and rural downwind areas of 20-km radii for winter seasons from 1999-2009.

6.2.4.1 Spring

5-km downwind
Figure 6.29: Warm-cloud SWIRR (left row), and warm-cloud albedo (right row) for high (a) and (b), medium (c) and (d), and low (e) and (f) UHI one hour downwind of NYC and rural downwind areas of 5-km radii for spring seasons from 1999-2009.
Figure 6.30: Warm cloud frequency per pixel per days of run for high (a) medium (b), and low (c) UHI one hour downwind of NYC and rural downwind areas of 5-km radii for spring seasons from 1999-2009.

20-km downwind
Figure 6.31: Warm-cloud SWIRR (left row), and warm-cloud albedo (right row) for high (a) and (b), medium (c) and (d), and low (e) and (f) UHI one hour downwind of NYC and rural downwind areas of 20-km radii for spring seasons from 1999-2009.
Figure 6.32: Warm cloud frequency per pixel per days of run for high (a) medium (b), and low (c) UHI one hour downwind of NYC and rural downwind areas of 20-km radii for winter seasons from 1999-2009.

6.2.4.1 Summer

5-km downwind
Figure 6.33: Warm-cloud SWIRR (left row), and warm-cloud albedo (right row) for high (a) and (b), medium (c) and (d), and low (e) and (f) UHI one hour downwind of NYC and rural downwind areas of 5-km radii for summer seasons from 1999-2009.
Figure 6.34: Warm cloud frequency per pixel per days of run for high (a) medium (b), and low (c) UHI one hour downwind of NYC and rural downwind areas of 5-km radii for summer seasons from 1999-2009.

20-km downwind
Figure 6.35: Warm-cloud SWIRR (left row), and warm-cloud albedo (right row) for high (a) and (b), medium (c) and (d), and low (e) and (f) UHI one hour downwind of NYC and rural downwind areas of 20-km radii for summer seasons from 1999-2009.
Figure 6.36: Warm cloud frequency per pixel per days of run for high (a) medium (b), and low (c) UHI one hour downwind of NYC and rural downwind areas of 20-km radii for summer seasons from 1999-2009.

### 6.2.4.1 Fall

5-km downwind
Figure 6.37: Warm-cloud SWIRR (left row), and warm-cloud albedo (right row) for high (a) and (b), medium (c) and (d), and low (e) and (f) UHI one hour downwind of NYC and rural downwind areas of 5-km radii for fall seasons from 1999-2009.
Figure 6.38: Warm cloud frequency per pixel per days of run for high (a) medium (b), and low (c) UHI one hour downwind of NYC and rural downwind areas of 5-km radii for fall seasons from 1999-2009.

20-km downwind
Figure 6.39: Fall season downwind SWIRR and warm-cloud visible albedo climatologist for 20-km downwind areas of NYC and two control regions for (a) high, (b) medium, and (c) low UHI.
6.3 Discussion: UHI Climatologies: seasonal, and averaged over all seasons

6.3.1 Warm Cloud SWIRR, albedo, and cloud fraction

Overall UHI outcomes demonstrate limited evidence for increased clouds from thermally generated UHI convection, as described in prior UHI work, in cloud fraction or albedo. There is a positive urban-rural cloud fraction difference for high UHI (Figure 6.11) in overall results. Urban-rural differences in seasonal plots, however, (Figure 6.20) show
mainly negative or low values across all seasons but winter, such that the positive average is a result of the high winter season contribution. For the winter season cloud fraction differences are strongly positive at all UHI levels. Winter albedo differences themselves could account for UHI values, and may dominate cloud fraction effects on heating under high UHI conditions. Winter has lower albedo with higher UHI, and negative urban-rural albedo for high UHI. Averaged seasonal results for albedo urban-rural differences are highly variable. A UHI cloud enhancement effect may be most clearly evident in high UHI urban-rural cloud fraction differences for the summer and fall seasons.

While these plots are of UHI and cloud properties, it is a larger dataset than that of aerosol, and since outcomes are also influenced by urban aerosols it can also be used to gain insights into possible aerosol-UHI interactions. UHI outcomes also may be better explained with possible aerosol effects in mind. Converse effects of UHI circulation masking cloud suppression may affect aerosol outcomes, but for that dataset it is harder to identify instances of a UHI influence.

Urban surfaces absorb solar radiation and release it later as sensible heat, while greater evapotranspiration in rural areas keeps ambient temperatures relatively cool on clear, sunny days. Cloud cover on overcast days, or clouds with higher albedo reduce solar input, and more is likely in low UHI conditions. The steeply dropping cloud fraction with higher UHI for both urban and rural areas in Figure 6.10 is therefore mainly a reflection of the weather regime in which UHI would be expected to develop. Spring, summer and fall seasons all exhibit this pattern, seen in Figure 6.19. Decreasing urban-rural albedo or
cloud fraction differences with increasing UHI would also be expected, as relatively high solar influx will also result in higher local temperatures. This can be seen in the 35-km downwind plot for albedo (Figure 6.9) for the average over all seasons. The winter season is again the primary contributor to this averaged outcome, as seen in Figure 6.18. A reversal of this pattern takes place for lower downwind radii, as low UHI albedo differences change from positive to negative with decreasing radius, and could be explained as aerosol cloud suppression.

Coakley (2007) found smaller droplet sizes in partly cloudy compared with overcast conditions, and also that droplet radius was overestimated in partly cloudy scenes. This is consistent with increasing SWIRR with UHI in the overall reflectance outcome in Figure 6.1, and in the inverse SWIRR-cloud fraction relationship in averaged overall values. Figure 6.10 shows decreasing averaged cloud fraction with higher UHI. In seasonal outcomes the spring and fall season exhibit the same patterns, seen in Figure 6.15, and Figure 6.19 and is minimal for winter and summer UHI. For winter, and for high UHI days in the fall SWIRR and cloud fraction varied in the same direction. For the aerosol dataset this inverse relationship was true both for almost all total and differenced values outside of the winter season.

High UHI with positive or increasing urban-rural cloud fraction differences are not expected, and such cases could indicate an urban cloud enhancement effect in the summer and fall seasons in Figure 6.20. Differences become less negative in the summer season with increasing UHI, and more positive in the fall. SWIRR results add to a picture of an
anomalous cloud-fraction difference which could be a urban heating effect. For summer and fall high UHI SWIRR differences are positive, which should predict depressed urban-rural cloud fraction.

In 5-km area outcomes in Figure 6.16 urban-rural SWIRR differences increase with UHI in both the spring and fall seasons. Based on Section 5 outcomes, higher aerosol urban-rural differences are likely to be present with higher SWIRR. Sunnier days would be expected to produce higher aerosol loads as sunlight interacts with primary pollutants, and energy use increase. Aerosol interaction with potentially thermally forced clouds could dampen possible cloud enhancement effects which might otherwise arise.

The urban effect on cloud fraction, as seen in changes with more narrowly focused study areas, is of a lower urban-rural difference for seasons other than winter. This can be seen most clearly by comparing the 35-km (or 20-km) urban-rural differences with the 5-km-downwind area radii outcomes. The exception, which could be explained by urban effects, is of fall high UHI.

**6.3.1.1 Albedo, SWIRR and land surface effects**

Even more distinctly than in many aerosol mapped climatologies, an inverse relationship between SWIRR and albedo is evident in the east-west gradations across images. Deeper clouds with larger droplet radii more likely to be capable of convection are found in onshore pixels on the western sides of climatologies. Off-shore pixels more influenced by
land cover, by contrast, have much lower albedo, and smaller particle size (high SWIRR). These air-masses carry more anthropogenic as well as other mineral and organic aerosols.

A pattern of strongly increasing SWIRR gradation towards the east side of the downwind images is seen in averaged season images such as Figure 6.3, as well as in the seasonal images that follow. It is least dramatic in the high UHI images, perhaps due to saturation effects. Figure 5.1 in the aerosol section showed a leveling off of SWIRR between medium and high aerosol levels. Note however that while gradations are stronger, the AOD scale in the high UHI legend reaches higher AOD values, and the predominant values seen are 0.35, while the highest value in the medium and low images is lower, at about 0.28. The location of the high SWIRR edge in Figure 6.2 is on the south-eastern side. This would correspond with the north-westerly winds in highest UHI recordings in a prior NYC-UHI study were on days (Gedzelman, 2003). While that study described a predominantly nocturnal UHI, the same conditions can also produce a daytime UHI.
In one example of many, the inverse gradation in Figure 6.41 is not only apparent in outlier pixels on the edges, but throughout the entire area of the three sets of climatologies. With a few exceptions higher SWIRR does not demonstrate an albedo effect. The 'albedo effect' refers to aerosols causing smaller droplet-size, which is associated with higher albedo.

This interpretation of Figure 6.41 assumes that land-based air-masses have higher aerosol levels than those from the ocean. The edges that have move from land to sea do have lower droplet size but their albedo is also lower, while the inverse is true for clouds blown inland. Again this may be due to cloud dissipation effects of aerosols and/or of lower liquid water content of offshore clouds.

![SWIR and Albedo Images](image)

Figure 6.41: Example of the inverse relationship of SWIRR and albedo pixels, and the influence of on-shore compared with off-shore air masses for a UHI climatology. Showing 20-km downwind-area radii medium UHI climatologies.

The cloud frequency images in these figures, particularly for the 10 and 20-km radius areas show an overall westward shift in the NYC plots with decreasing UHI, corresponding to more easterly, onshore wind direction. Onshore breezes carry fewer aerosols, as well as
cooler temperatures; sea breezes also act to dampen the UHI effect (Gedzelman, 2003).

### 6.3.1.2 Summer season

The summer season has much lower overall SWIRR values (Figure 6.15) compared with other seasons, but fAOD levels are higher in the summer, as seen in the increased high fAOD data availability in this study. Jin (2005) found AOT as measured by MODIS 0.56-µm to routinely double between winter and summer, and on very hot days additional high-emissions energy production is shifted to the city. Aerosols act to decrease droplet radius, but as seen in the aerosol section, overall SWIRR decreases on higher aerosol days, while urban-rural SWIRR differences increase with fAOD. As seen in the aerosol section, the cloud response to aerosol can saturate at relatively low fAOD levels. Higher temperatures, humidity, and aerosols all increase in the summer, and can affect droplet radius. The high urban-rural difference in SWIRR (Figure 6.16) could be attributed to elevated urban aerosol levels, and to the particular urban aerosol composition.
In the summer season warm-cloud SWIRR and visible albedo mapped images, Figure 6.33 and Figure 6.35 the west-east SWIRR and albedo gradations found in overall averages and in other seasons is reduced or absent. Across UHI and many fAOD levels SWIRR and albedo bar-plots also show little variability in the summer season. This could be due to local saturation effects. Table 8 shows that winds in the summer season are predominantly West-Southwesterly. This is the same angle as the land-sea border, and could also result in more uniform reflectance and albedo.

Cloud fraction, in Figure 6.19, is lower in all stations with increasing UHI, as expected. Lower urban-rural differences would also be expected with increasing UHI, but Figure 6.20 shows that in summer, as in fall, differences increase with UHI. This could be interpreted as a cloud enhancement. The positive SWIRR difference would also predict a negative cloud fraction difference based on aerosol results. The outcome may be the balance between a decrease due to aerosols, and an increase due to heating effects, and could be an instance of urban cloud enhancement.

**6.3.1.3 Winter season**

The winter season follows a very different UHI pattern from the other three seasons, as do aerosol winter outcomes. Cloud fraction decreases in other seasons with higher UHI, but remains relatively low throughout the winter season (Figure 6.19). Decreasing overall albedo (Figure 6.17), and generally decreasing urban-rural albedo differences with increasing UHI (Figure 6.18) can account for UHI levels where cloud fraction does not.
Urban-induced cloud fraction enhancement may be evident in winter, in which cloud fraction urban-rural differences are consistently positive, and much higher than in other seasons (Figure 6.20). Urban-rural cloud fraction also increases with more narrowly focused study areas. Aerosol levels are lower in winter, and this result could be due to removal of dampening effects of aerosols on cloud fraction in other seasons. Despite low aerosol levels, and decreasing SWIRR in winter in response to aerosol, SWIRR is high across all UHI levels and locations in winter. So in this case factors other than aerosols may contribute to smaller warm-cloud droplet radius. Winter has relatively low variation in SWIRR with UHI, especially in the urban area.

While the role of aerosol is unclear in UHI outcomes, the UHI section draws from a larger dataset than that of aerosol, and the urban region has higher average aerosol levels than rural areas. With the exception of high UHI albedo, most cloud fraction and cloud albedo urban-rural differences are positive. The ‘cloud lifetime,’ and ‘cloud albedo effects’ would predict this response of clouds to higher SWIRR, or smaller particle radius. For albedo this is seen in a comparison of SWIRR in Figure 6.16 with albedo in Figure 6.18. Winter cloud albedo however correlates inversely with SWIRR in mapped climatologies, in Figure 6.25 and Figure 6.27. While this contradicts the basis for the cloud albedo effect, it is consistent with the result found in other plots throughout this study of higher aerosol and lower cloud particle radii together with lower albedo.

Cloud fraction for the urban areas increases somewhat between low and medium, and
levels off or decreases slightly from medium to high UHI. For the 5-km downwind area, optically thick (as for all clouds measured in this study) cloud fraction is higher in urban compared with rural sites at all UHI levels. Though the standard error of mean, as explained earlier, is always high in cloud fraction outcomes, the urban-rural differences are a substantial proportion of the overall measure. They are the greatest differences, positive or negative, across all seasons in all downwind areas. While the medium UHI cloud fraction difference decreases slightly, for high and low UHI differences greatly increase with decreasing study area size. As described earlier, lower cloud fraction is normally associated with higher UHI due to cloud effects on temperature, so a positive urban-rural cloud fraction for high UHI is anomalous. So there is a good chance that in winter under relatively high urban surface temperatures a higher than expected cloud fraction is a result of urban effect. Under low UHI conditions higher urban cloud fraction is expected, as is a clearer response with narrower study areas.

Other factors may contribute to the different form of UHI in winter, both for high and lower UHI. At 41 degrees latitude there is much lower solar radiation in winter than other seasons. This decrease, in combination with a thinner mixed planetary boundary layer may make the anthropogenic heat flux, along with urban storage and release of sensible heat, normally overwhelmed in daytime by solar radiation, a more important variable. Under lower UHI conditions it may be more quickly advected from its source, but still affect downwind atmospheric structure and stability. This could contribute to positive urban-rural cloud fraction at all UHI levels at the 5-km area. The urban canopy layer may also have stronger turbulence effects in a thinner PBL.
6.3.2 Cold clouds

While there can be climatological urban effects on cold cloud albedo in the form of deep convective clouds, these events are relatively rare, and land surface variables are less likely to determine trends in their development than in lower clouds. Similarities in the overall pattern of warm and cold cloud albedo and cloud fraction should be expected in that cold clouds can also be responsible for the conditions leading to high or low UHI regimes. Both warm and cold cloud properties are also mediated by the synoptic-scale weather regime, which are subject to boundary-layer interactions. The use of the same ground-station data to predict downwind locations of both warm and cold clouds leads to higher error in designations of cold clouds downwind (see Section 4.5).

For cold clouds different factors appear to contribute to UHI on different seasons. Figure 6.22 and Figure 6.21 show that in the winter urban-rural albedo differences often decrease with increasing UHI. As with warm clouds, in spring, summer and fall cloud fraction decreases greatly with higher UHI, seen in Figure 6.23; in Figure 6.24 cloud fraction differences generally decrease with UHI. The spring season has some positive urban-rural differences for high UHI in both albedo and cloud fraction that could signal an urban effect. Urban-rural cloud fraction in the winter is lower for high UHI but is positive for all UHI levels. This may indicate an urban effect.
One outcome of the image climatologies, Figure 6.12, Figure 6.13, and Figure 6.14, is the much greater dispersal in downwind air-masses in the NYC area compared with the rurals. Rural cloud-frequency images with larger datasets more readily take on a simple oval shape with concentric rings of decreasing cloud frequency, while urban winds are more dispersed and irregular. Bornstein and LeRoy (1990) have found moving thunderstorms to bifurcated and moved around the city due to a “building-barrier-induced divergence” effect. The urban wind measures may show turbulent effects of the roughness length of the urban surface.
7 CONCLUSIONS AND SUMMARY

7.1 Aerosols

Reflectance at 3.9-µm is lower overall on high aerosol days. This outcome may be explained by confounding synoptic variables. Urban-rural SWIRR differences increase significantly with fAOD. This elevated relative SWIRR is accompanied by proportionally reduced urban cloud fraction in the spring, summer, and fall seasons. The summer season demonstrates this result most strongly and consistently, though its overall SWIRR is lower than for other seasons. This outcome may be evidence of urban aerosol cloud dissipation effects. Existing theories consistent with this outcome are of evaporation-entrainment and sedimentation effects, or of semi-direct absorbing aerosol effects.

Albedo decreases overall with increasing aerosol. In mapped image climatologies off-shore clouds carrying more aerosols are often progressively optically thinner, and have reduced cloud droplet radius. In many of these climatologies SWIRR pixel values appear to inversely correlate with albedo, showing increasingly deep onshore clouds with larger cloud-top particle radii in pixels that were more likely to have been advected from the Atlantic. These patterns may be an outcome of cloud dissipation in offshore clouds with higher aerosol levels, or of the differing initial water content of the land compared with oceanic clouds.

A glaciation effect may explain the outcomes for cold-cloud albedo, which increases over-all in response to aerosol, but which have increasingly negative urban-rural differences with
increasing aerosol. The spring season in particular exhibits a consistently reduced urban-rural albedo difference with higher aerosol levels; all seasons show increasingly negative differences between the low and medium aerosol datasets.

7.2 UHI

Evidence of UHI enhancement of clouds is limited, but a few notable cases merit further exploration. A UHI cloud enhancement effect may be evident in high UHI urban-rural cloud fraction differences for the summer and fall seasons. The winter season is also notable for higher urban cloud fraction across the three UHI levels. The urban influence on cloud fraction independent of UHI is of cloud reduction in seasons outside of winter, and an increase in the winter season with narrower study areas.

Positive, and increasing urban-rural cloud fraction for high and increasing UHI in the fall season may be a case of urban cloud enhancement. All other things being equal, lower cloud fraction in the urban compared with rural region would be expected on days with a positive urban-rural temperature difference. An increasing difference with increasing UHI is present in this season. The inverse SWIRR- cloud fraction relationship found in the aerosol section of this study may not be present in this and other differenced outcomes due to UHI-related effects. The strengthening of an outcome as it more narrowly draws from pixels in the urban and rural study areas tend to strengthen the case for an urban effect. For seasons other than winter urban-rural differences in SWIRR increase with higher UHI,
with greater increases in smaller study areas. The fall albedo difference for high UHI increases over-all with decreasing downwind areas. All things being equal albedo differences would be expected to be negative under high UHI conditions. Albedo differences with increasing UHI also shift with decreasing downwind area from decreasing at the 35-km radius area to an increasing one for 20 and 10-km-radius areas.

The summer season also shows possible urban cloud enhancement effects in increasing urban-rural cloud fraction differences with increasing UHI. This season features a stronger shift from decreasing to increasing albedo difference with increasing UHI, despite positive SWIRR differences. The winter season is a third candidate for further exploration for possible UHI-enhanced cloud fraction. The highest cloud fraction urban-rural differences are consistently found in the winter season. The winter PBL is thinner and experiences much less solar influx, may be more sensitive to urban heating effects. Albedo differences are also high, but they decrease with increasing UHI and decreasing downwind-area radius.

For cold clouds the spring season shows neither overall, nor urban-rural decreases in cloud albedo that contribute to the higher urban temperatures measured for higher UHI days. This effect is even greater in smaller study areas, possibly signaling an urban cold cloud enhancement.
### 7.3 Overall urban effects

The mapped climatologies showing downwind effects of land compared with oceanic air masses are a unique and notable outcome of this work. These images demonstrate the contrast between land-surface compared with oceanic cloud characteristics at their coastal boundary, with higher SWIR and lower albedo downwind of land. They repeatedly show an inverse SWIR-albedo relationship, as in Figure 6.41. This outcome is in agreement with a cloud dissipation influence of urban and land-sourced aerosol effects. The pattern is found in both urban and rural climatologies, but the NYC region has higher average AOD, thus the effect will be stronger there. These observations pair well with numeric outcomes of higher urban SWIR and lower cloud fraction. Cloud dissipation effects found in numeric outcomes could explain the off-shore albedo effect.

The urban aerosol effect of reduced warm cloud fraction for spring, summer and fall is one of increased radiative forcing, as low clouds have an overall cooling effect. Warm cloud fraction also decreases with narrower study areas in all but one of the nine UHI cases outside of winter. NYC is one of the largest of many temperate-region coastal mega-cities with comparable patterns of land use and air pollution. As cloud reduction was found in proportion to increasing aerosol levels, this urban effect can be generalized to effects of anthropogenic aerosols downwind of cities at various scales in coastal temperate regions.

Cloud fraction reduction in the aerosol dataset should be considered in the context of predicted surface-temperature warming, and of increasing temperature extremes. Local
urban effects found in this study could compound forecasted regional increases in air temperatures and temperature extremes. A 1 degree Celsius increase is predicted for June-August by 2035 for this region relative to 1986–2005 temperatures (Oldenborgh, IPCC 2013). An increasing frequency of extreme heat events is also predicted for the Northeast United States (Figure SPM.4A, Field, IPCC 2012). Heat waves are associated with human discomfort, and higher mortality rates for vulnerable sectors of the population. The 2003 European heat wave resulted in 52,000 deaths (Larson, 2006). Extreme heat also puts a strain on the current northeast United States energy grid. Further aggravating this potential scenario, the current NYC-metropolitan power grid structure results in more air pollution on days of peak energy demand (see Section 2.1.8). This scenario could present additional challenges to urban residents and planners in terms of quality of life, health and safety, and energy production. A transition to renewable energy sources will reduce these effects, but the current trend is of increasing aerosol inputs as coastal cities grow in size and population.

The UHI dataset reveals greatly increased albedo, and consistently higher cloud fraction in the winter season. Cloud fraction urban-rural differences decrease with narrower study areas, but differences at the 5-km downwind-area radius still suggest an urban cooling effect in winter. In the aerosol dataset cloud fraction urban-rural difference is also strongly positive, and increases with narrower study areas; this suggests an urban effect colder weather in winter. This study highlights the summer and fall seasons as candidates for further study of UHI-enhanced cloudiness.
Cold cloud albedo may show more sensitivity to urban effects than do warm clouds. In the summer season rural albedo in particular increases with higher urban aerosol measures, and also has an increasingly negative urban-rural differences at smaller downwind areas. Across the seasons urban-rural albedo differences become more negative with smaller downwind areas, with larger decreases at higher aerosol levels.

This study is unique in its creation of long-term climatologies using satellite imagery in combination with ground-station data to explore downwind urban effects on low clouds in comparison with control regions. The use of years of daily satellite, wind, temperature and aerosol data reveal distinct urban signatures through urban-rural differences, and for more exclusively urban and rural study areas. A key result in urban-rural differences is a strong urban-rural cloud particle size difference with increasing aerosol levels, and the inverse relation of SWIRR with cloud fraction across the ‘clear-sky’ aerosol dataset. These statistical outcomes make up a robust dataset that contribute to our understanding of urban effects on weather and climate.

7.4 Future work

A few additions and modifications could greatly enhance the value of this work. An additional GOES dataset two hours prior to the one used, allowing for an upwind and downwind comparison of each day’s data could address distortions created by the different topographies of the three study areas. The additional division of the dataset into synoptic
regimes could isolate conditions in which observed effects are more likely to be found, and also greatly aid in explaining observed differences. Where aerosol data availability is sufficient, the combination of the two datasets could allow for a better understanding of their interactions.

The downwind area prediction was made with limited available data in the form of ground station wind speed and direction. The use of this data for higher cold cloud downwind projections is particularly likely to suffer from error due to this input. High resolution modeled downwind trajectory data could allow for a more accurate downwind location prediction, and for additional predictions further downwind.

While SWIRR can be used as a proxy for particle radius, the inverse relationship between it and SWIRR is non-linear. The transformation of reflectance into estimated particle radius based on a published radiative transfer model (Cattani, 2007) could allow for a greatly improved interpretation of this measure.

Scatterplots of SWIRR against albedo values of individual averaged pixels used in the mapped climatologies could be used to further investigate the inverse relationships observed in mapped plots. They could also be used to explore circumstances in which these outcomes were more prevalent. A restriction of outcomes to offshore pixels based on wind direction in this set, and across climatologies could more narrowly focus on microphysical effects of land-based aerosols. A restriction to only those over land could provide more useful information on effects on local residents.
The cold and warm cloud designations in this study only separate liquid water clouds from all others. The addition of a third mixed-phase, in addition to ice or other designations could better separate processes that take place in particular cloud types.

The 130-km distance from NYC for control locations was chosen in part to allow for the extraction of data of 65- km radius with limited overlap of the study and control regions. Enough data for many plots was available however for many 5 and 10-km plots, and a comparison with somewhat closer regions. Controls at closer distances to NYC will be closer to higher-density development of both the northern and southern study areas, as seen in the NDVI map, but could have the advantage of a more direct comparison with the urban study region.
APPENDIX A: SOCIAL AND ECONOMIC COSTS OF EXCESS URBAN HEATING

Urban heat islands result in increased energy use and detrimental health effects for local residents. UHI leads to higher consumption of energy with increased use of air conditioning in homes and buildings. This problem in New York City summers is a serious burden on its energy distribution grid. Today one-sixth of the electricity consumed in the U.S. is used for cooling purposes, at an annual cost of $40 billion (Rosenfeld et al, 1997). Reductions in urban air temperature by just a few degrees could save consumers millions of dollars on their utility bills each year. In 2007 the New York State Energy Research and Development Authority began a tree-planting pilot project as a mitigation measure to the problem (NYSERDA-DEC).

Reduction in urban air temperatures would also lower harmful air emissions, such as sulfur dioxides (SO2) and nitrogen oxides (NOx), which are produced when fossil fuels are burned to generate electricity. Hot, sunny days in urban areas are also ideal conditions for the formation of ground-level ozone (Lyamani, 2006). Ambient air pollutants (fine particulates, ozone, oxides of nitrogen, and diesel exhaust) increase incidence of asthma and are known to trigger asthmatic attacks. Airborne particles can cause or enhance respiratory, cardiovascular, infectious, and allergic diseases (Finlayson-Pitts, 2000, Bernstein, 2004, Tarlo 2004, Finlayson-Pitts 1997). Reduction in levels of ambient air
pollution was associated with fewer hospitalizations for asthma and other respiratory diseases (Landrigan 2006). New York City residents suffer high rates of asthma, which disproportionately affect the city’s poorer and minority residents (Maantay, 2007).

**APPENDIX B: AEROSOL MEASUREMENTS AND DATA SOURCES**

### B1 Aerosols properties and measurements:

Aerosol characteristics, given for a wavelength $\lambda$, include the vertical profile of the scattering $\sigma_s$ and absorption $\sigma_a$ coefficients and the scattering phase function. Instead of $\sigma_s$ and $\sigma_a$, one can use the extinction coefficient $\sigma_e = \sigma_a + \sigma_s$ and the single scattering albedo $\omega = \sigma_s / \sigma_e$; for nonabsorbing aerosols, $\omega = 1$. The aerosol depth is defined by

$$
\tau_{e,s,a} = \int_0^\infty \sigma_{e,s,a}(z) \, dz,
$$

Equation B0.1

for extinction, scattering, and absorption. The angular distribution of the scattered photons is characterized by the phase function $p(\theta)$, where $\theta$ is the scattering angle, between the incidence and the scattering directions. As $p(\theta)$ is normalized to $4\pi$ by integration over all directions, it does not depend on the total amount of the particles. If one is interested in the polarization effect of scattering, the scalar function $p(\theta)$ has to be replaced by a $4 \times 4$ phase matrix $P(\theta)$. 

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### B1.2 AERONET Aerosol Optical Depth Measurement

AOD is obtained from direct sunphotometer measurements data using the appropriate calibration constant. The optical depth due to water vapor, Rayleigh scattering, and other wavelength-dependent trace gases are subtracted from the total optical depth to obtain the aerosol component.

Aerosol Optical Depth (AOD) is the measure of aerosols (e.g., urban haze, smoke particles, desert dust, sea salt) distributed within a column of air from the instrument (Earth's surface) to the top of the atmosphere. AERONET uses voltage (V) measured by a sun photometer, which is proportional to the spectral irradiance (I) reaching the instrument at the surface. The estimated top of the atmosphere spectral irradiance (I_o) in terms of voltage (V_o) is obtained by sun photometer measurements at Mauna Loa Observatory in Hawaii. The total optical depth (τ_{TOT}) can be obtained using the following equation according to Beer-Lambert-Bouguer law:

\[
V(\lambda) = V_o(\lambda) \cdot d \cdot \exp[-\tau(\lambda) \cdot \tau_{TOT} \cdot m] \tag{Equation B0.2}
\]

where \( V \) is the digital voltage measured at wavelength \( \lambda \), \( V_o \) is the extraterrestrial voltage, \( d \) is the ratio of the average to the actual Earth-Sun distance, \( \tau_{TOT} \) is the total optical depth, and \( m \) is the optical air mass (Holben 1998).

Other atmospheric constituents can scatter light and must be considered when calculating the AOD. The optical depth due to water vapor, Rayleigh scattering, and other wavelength-dependent trace gases must be subtracted from the total optical depth to obtain the aerosol component:
\[ \tau(\lambda)_{\text{Aerosol}} = \tau(\lambda)_{\text{TOT}} - \tau(\lambda)_{\text{water}} - \tau(\lambda)_{\text{Rayleigh}} - \tau(\lambda)_{\text{O3}} - \tau(\lambda)_{\text{NO2}} - \tau(\lambda)_{\text{CO2}} - \tau(\lambda)_{\text{CH4}} \quad \text{Equation B0.3} \]

**B1.3 Criteria for Level 1.5 AOD retrievals:**

1. At least three wavelength combinations must include 440 and 870nm with either 490, 500 or 675nm.

2. The AOD for each channel must be greater than or equal to 0.02/m, where \( m \) is the optical air mass.

3. Outliers are removed according to the following criterion:

\[ \text{Abs}(\text{AOD}_{500\text{nm}} - \text{AOD}_{\text{SDA500nm}}) > (0.02 + \text{AOD}_{500\text{nm}} \times 0.005) \]

**B1.4 Urban Aerosol Characteristics**

Cities create multiple sources of pollution, mainly from the incomplete combustion of fossil fuels. Figure B.1 below shows the difference in air quality in a city (Munich) compared with a nearby alpine location. Total particle number concentrations was about $10^2$ cm$^{-3}$ in alpine compared with $10^4$ cm$^{-3}$ in urban air, and particle mass concentrations of 1 μgm$^{-3}$ and 10 μgm$^{-3}$, respectively.
B1.4.1 Aerosol composition in New York City

Aerosols in New York City are distributed according to below. This data was collected at a Queens College site (a borough of NYC), and secondary sulfate was found to be the principle source of PM2.5, contributing to 64.6% of its total mass concentration. Oil combustion and motor vehicles are sources of primary organic carbon, and black carbon.
Figure B0.2 Averaged contributions to PM2.5 mass concentrations measured at the Queens College site during July 2001 (Li et al., 2004).

**Appendix C: GOES data and cloud properties**

This data was obtained from the NOAA CLASS website, which provides GOES-East Satellite data in various formats. This study primarily used 16-bit AREA data, converted later to NetCDF format using the NOAA Weather and Climate Toolkit (WCT). Channels 2 and 4 were calibrated when being converted by the NOAA WCT. This program did not calibrate the visible, channel 1 so it was calibrated and converted from raw counts to albedo using online NOAA reference pages, as described below.
APPENDIX C: GOES DATA CALCULATIONS

C1: GOES Channel 1 Calibration

C1.1 Data download

Data was downloaded at a 4-km resolution from NOAA’s CLASS website (Comprehensive large Array-data Stewardship System).¹

Data was converted after downloading from raw to net-CDF format with the NOAA Weather and Climate Toolkit. The output of this was calibrated images for the 2 and 4 channels, but uncalibrated for the visible, channel 1.

Calibration was not done automatically for channel 1, so the code processing GOES files includes a pre-launch and post-launch calibration for the visible channel.

Pre-launch and post-launch calibration were added to this data after its conversion to net-CDF. The NOAA pages used for calibration are found in referenced endnotes.

The data I downloaded was 16-bit. I need 10-bit data for these calibrations, so the first operation was to divide by 32 to convert to 10-bit².
C1.2 Pre-launch calibration

The pre-launch calibration for both GOES-8 and GOES-12 for reflectance was done according to NOAA web page specifications:

\[ A_{\text{pre}} = k \left( X - X_{\text{space}} \right) \]  

Equation C0.1

Where \( A_{\text{pre}} \) is nominal pre-launch reflectance, the ratio of \( R \) relative to nominal solar radiance, i.e., the spectral radiance when the Sun is at local zenith and mean Sun-Earth distance (unit AU or Astronomical Unit). As a ratio, \( A_{\text{pre}} \) has no unit and value of 0-1.

\( K \) is the calibration coefficient, found in a lookup table. For GOES-8 \( k = 0.001062 \); for GOES-12 \( k = 0.001141 \); for GOES-13 \( k = 0.001160 \).

\( X_{\text{space}} \) is the instrument response to space scene where signal is expected to be zero. For the visible channel of GOES Imager, the instrument is clamped to the space such that \( X_{\text{space}} \) should always be 29.

C1.3 Post-launch calibration

For the post-launch calibration, a correction for satellite degradation over time,

For GOES-8 the equation used was:

\[ A \left( d; \text{post} \right) = 1.192 \times A(\text{pre}) \times (1 + 0.0001688 \times d) \]  \( \text{ (in percent)} \)  

Equation C0.2
Where the day, \( d \) refers to the days since the day of launch for GOES-8, April 13, 1994

For GOES-12, the equation\(^5\)

\[
A_{\text{post}} = a \exp(bt) A_{\text{pre}} \quad \text{or} \quad R_{\text{post}} = R_{\text{pre}} \times C, \text{ with } C = a\exp(bt)
\]

Equation C0.3

Was used, where \( A_{\text{post}} \) is the post-launch calibrated reflectance.

The coefficients \( a \) and \( b \) used here depend on the date, \( t \).

The variable \( t \) is the time (in years) from the satellite’s launch date of April, 1 2003 to the time of the satellite image in question. See endnotes and Appendices for tables and URL references.

For GOES-13, the equation\(^6\)

\[
A_{\text{post}} = a \exp(bt) A_{\text{pre}} \quad \text{or} \quad R_{\text{post}} = R_{\text{pre}} \times C, \text{ with } C = a\exp(bt)
\]

Equation C0.4

Was used, with the same format as for GOES-12, where \( A_{\text{post}} \) is the post-launch calibrated reflectance.

A solar elevation correction was made using Stull (2000). To confirm that it was working correctly output was compared with online calculators of solar elevation angle.

The three equations used for this calculation were:
Solar declination angle:

\[ \delta_s = \Phi_e \cdot \cos \left[ \frac{C \cdot (d - d_r)}{d_y} \right] \]  

Equation C0.5

Where \( \Phi_e \) is the ecliptic, 23.45° (0.409 rads)

\( d \) is the Julian Day

\( d_r \) is the summer solstice, 173 or 174

\( d_y \) is the number of days in the year

Local elevation angle:

\[ \sin(\Psi) = \sin(\phi) \cdot \sin (\delta_s) - \cos(\phi) \cdot \cos (\delta_s) \cdot \cos \left[ \frac{C \cdot t_{UTC}}{t_d} - \lambda_e \right] \]

Equation C0.6

Where \( \phi \) is the latitude

\( \delta_s \) is the solar declination angle from above

\( t_{UTC} \) is the local time in UTC

\( t_d \) is 24 hours, the length of one day

And \( \lambda_e \) is longitude
Last, the zenith angle:

\[ \zeta = \frac{C}{4} - \Psi \]  

Equation C0.7

Where \( C \) is 2\( \pi \) radians, or 360°

And \( \Psi \) is the local elevation angle from above

**C2: Baseline GOES Data Thresholds**

For both GOES channels 2 and 4 the baseline values for upper and lower limits were taken from a NOAA-OSO range of realistic temperature. Find the values applied in Table C1 below. These limits are also used in a NOAA-NESDIS temperature conversion scheme (7, 8, 9). These values, of a lower limit of 163° K, and an upper limit of 330° K, are the initial thresholds applied. Any temperature readings beyond this range are considered spurious or error.
Table C1. Baseline thresholds applied to GOES channels.

<table>
<thead>
<tr>
<th>GOES Channel</th>
<th>Central Wavelength (µm)</th>
<th>Lower Limit</th>
<th>Upper Limit</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.65 µm</td>
<td>0%</td>
<td>100%</td>
</tr>
<tr>
<td>2</td>
<td>3.9 µm</td>
<td>163º K (-110º C)</td>
<td>330º K (57º C)</td>
</tr>
<tr>
<td>4</td>
<td>10.7 µm</td>
<td>163º K (-21º C)</td>
<td>330º K (57º C)</td>
</tr>
</tbody>
</table>

C3: Solar Zenith Angle

First, the fractional year (γ) is calculated, in radians.

\[
\gamma = \frac{2\pi \times (\text{day} - 1 + \text{hour} - 12)}{365 \times 24}
\]

Equation C0.8

From (γ), we can estimate the equation of time (in minutes) and the solar declination angle (in radians).

\[
eq time = 229.18 \times (0.000075 + 0.001868 \cos \gamma - 0.032077 \sin \gamma - 0.014615 \cos 2\gamma + 0.040849 \sin 2\gamma)
\]

Equation C0.9

\[
decl = 0.006918 - 0.399912 \cos \gamma + 0.070257 \sin \gamma - 0.006758 \cos 2\gamma
\]

+ 0.000907 \sin 2\gamma - 0.002697 \cos 3\gamma + 0.00148 \sin 3\gamma

Equation C0.10
Next, the true solar time is calculated in the following two equations. First the time offset is found, in minutes, and then the true solar time, in minutes.

\[
\text{time\_offset} = \text{eqtime} - 4 \times \text{longitude} + 60 \times \text{timezone} \tag{Equation C0.11}
\]

where eqtime is in minutes, longitude is in degrees, timezone is in hours from UTC (Mountain Standard Time = +7 hours).

\[
st = \text{hr} \times 60 + \text{mn} + \text{sc} / 60 + \text{time\_offset} \tag{Equation C0.12}
\]

\(a 60), \text{sc}\) is the second (0 - 60).

The solar hour angle, in degrees, is:

\[
\text{ha} = (st / 4) - 180 \tag{Equation C0.13}
\]

The solar zenith angle (N) can be found from the following equation:

\[
\cos \phi = \sin(\text{lat}) \sin(\text{decl}) + \cos(\text{lat}) \cos(\text{decl}) \cos(\text{ha}) \tag{Equation C0.14}
\]

And the solar azimuth (2, clockwise from north) is:

\[
\cos(180 - \theta) = \sin(\text{lat}) \cos \phi - \sin(\text{decl})
\]

\[
\cos(\text{lat}) \sin \phi \tag{Equation C0.15}
\]
**C4: Earth’s Eccentricity**

Earth’s orbit around the sun is elliptical. The below calculation was used to determine the earth-sun distance for every day of the year, taken from Partridge (1976). In this approximate equation for the earth-sun distance, $R_{av}$ is mean sun-earth distance and $R$ is the actual sun-earth distance depending on the day of the year. The average distance used is 149,597,870.7 km.

\[
\left( \frac{R_{av}}{R} \right)^2 = 1.00011 + 0.034221 \cos(b) + 0.001280 \sin(b) + 0.000719 \cos(2b) + 0.000077 \sin(2b) \]  

Equation C0.16

where $b = 2\pi n / 365$ radians

and $n$ is the day of the year

**C5: Particle radius model: Computation of channel 2 reflectivity**

The following calculations are used in Daniel Lindsey’s Particle radius model, and are originally from Stevak and Doswell (1991). These were performed for AVHRR channels 3 (3.55-3.93μm ) rather than GOES channel 2 (3.80 - 4.00 μm), and for AVHRR channel 4 (10.3-11.3 μm) rather than GOES channel 4 (10.20 -11.20μm).
If \( \varepsilon_2 \) denotes the channel 2 emissivity and \( \alpha_2 \) is the channel 2 reflectivity, the total radiance \( N_2 \) measured by the satellite during the daylight hours can be expressed as

\[
N_2 = N_{2\text{ref}} + \varepsilon_2 N_2(T),
\]

Equation C0.17

where \( N_{2\text{ref}} \) stands for the reflected component and \( N_2(T) \) is the emitted component for a blackbody at temperature \( T(K) \), which is found by computing the Planck function appropriate for channel 2 (see Lauritson et al. 1979 or d’Entremont and Kleespies 1988).

The reflected component in Equation C0.17 can be determined from

\[
N_{2\text{ref}} = \alpha_2 N_2(T_s)(R/r)^2 \cos \xi,
\]

Equation C0.18

where \( T_s \) is the blackbody temperature of the solar photosphere (5800 K), \( R \) is the radius of the sun, \( r \) is the radius of the earth’s orbit, and \( \xi \) is the solar zenith angle. For convenience, let

\[
S_2(r, \xi) = N_2(T_s)(R/r)^2 \cos \xi,
\]

Equation C0.19

where \( S_3 \) is the solar flux at the top of the atmosphere, so that Equation C0.1 can be rewritten using Equations C0.18 and C.0.19 as

\[
N_2 = \alpha_2 S_2(r, \xi) + \varepsilon_2 N_2(T).
\]

Equation C0.20
Observe that this assumes the independence of $\alpha_2$ and $\varepsilon_2$ with respect to wavelength of the radiation within the window of channel 2. For sufficiently dense clouds, it is reasonable to assume a zero transmissivity, so Kirchhoff’s Law reduces to

$$\alpha_2 + \varepsilon_2 = 1.$$  \hspace{1cm} \text{Equation C0.21}

From Equations C0.20 and C0.21, it is easy to derive the following relations for computation of the AVHRR channel 3 emissivity and reflectivity:

$$\varepsilon_2 = \frac{N_2 - S_2(r, \xi)}{N_2(T) - S_2(r, \xi)}$$  \hspace{1cm} \text{Equation C0.22}

$$\alpha_2 = \frac{N_2 - N_2(T_s)}{S_2(r, \xi) - N_2(T)}.$$  \hspace{1cm} \text{Equation C0.23}

The value of $N_2$ is obtained from the calibration relation between counts and radiances for each pixel, while $S_2(r, \xi)$ is calculated from Equation C0.19 if the actual values of $r$ and $\xi$ are known.

Some simplification is necessary in practice, in order to determine the value of $N_2(T)$. To determine the temperature $T$, one can use the channel 4 (10.7\,\mu m) data provided $\varepsilon_2 = 1$ (i.e., a blackbody). Since real values of $\varepsilon_2$ are always less than one, this is a source of error in the computations. However, for sufficiently deep clouds, setting $\varepsilon_2$ to unity is a reasonable approximation (see Stephens 1978).
In summary, the technique employs the following simplifications.

(i) Absorption and dispersion of the incident, reflected, and emitted radiation are neglected.

(ii) Deviations from unit emissivity are neglected when determining the temperature $T$ from the channel 4 radiances.

(iii) The effect of water vapor on channel 4 data is neglected.

(iv) Zero transmissivity is assumed.

(v) Diffuse reflection (Lambertian surface) is assumed.

**APPENDIX D: CONTROL POINT DESIGNATION**

Various programs and online tools were used to designate the control points, based on distance from the New York City, distance from the shore, and degree of urbanization of the areas in question.

Figure D1 below shows the Matlab display of New York City’s latitude and longitude with designation for data extraction of a very small radius around that point. This served as a test to ensure that the latitude and longitude location for radius data extraction were correct in the program.
The following online tools were used to calculate the radius, the distance from shore of each point, and the location of those points on the circle. The radius chosen was acceptable based on the degree of vegetation seen on Google Earth. Alternative locations of a similar range in distance from NYC that were considered were more developed than the ones chosen.

The tool to draw a radius of specified length around the location of the Central Park station was:

http://www.freemaptools.com/radius-around-point.htm?clat=40.779&clng=-73.969&r=130&lc=FF0000&lw=1&fc=FF00FF

(this URL shows the radius image itself)
The distance calculator to select particular points along the radius of equal length to the shore was: [http://www.daftlogic.com/projects-google-maps-distance-calculator.htm](http://www.daftlogic.com/projects-google-maps-distance-calculator.htm)

To find the latitude and longitude of designated points along the radius:

http://itouchmap.com/latlong.html

To confirm that the distances were correct by entering latitude and longitude values

http://jan.ucc.nau.edu/~cvm/latlongdist.html

Figure D2: a radius of 130 km around NYC. The URL for the radius of 130 km is:

http://www.freemaptools.com/radius-around-point.htm?clat=40.779&clng=-73.969&r=130&lc=FF0000&lw=1&fc=FFFF00
APPENDIX E: ADDITIONAL AEROSOL CLIMATOLOGY FIGURES

E1 Results: Aerosol Climatologies averaged over all seasons

E1.1 Warm Cloud Shortwave Infrared Reflectance at 3.9-μm

E1.1.1 Numeric results (see cloud fraction for warm clouds for fraction results)

Figure E 0.1: Three aerosol levels and warm-cloud SWIRR (left row), and urban-rural differences (right row) from 2002-2008 over all seasons for 1-hr downwind areas of 50-km radius.
E1.2 Warm, cold and all-cloud visible albedo over all seasons

### E1.2.1 Numeric results

**Figure E 0.2:** Three aerosol levels and warm, cold, and all-cloud visible albedo over all seasons from 2002-2008 for 1-hr downwind areas of (a) 5-km and (b) 20-km radii.
E1.2.2 Numeric results for urban-rural differences

Figure E.0.3 Urban-rural differences for three aerosol levels and warm, cold, and all-cloud visible albedo over all seasons from 2002-2008 for 1-hr downwind areas of (a) 5-km and (b) 20-km radii.
E1.2.3 Numeric results for cloud fraction

Figure E0.4 Three aerosol levels and cloud fraction for warm, cold, and all-clouds over all seasons from 2002-2008 for 1-hr downwind areas of (a) 5-km (b) 20-km radii.
E1.2.4 Numeric results for cloud fraction urban-rural differences

Figure E0.5 Urban-rural differences for three aerosol levels and cloud fraction for warm, cold, and all-clouds over all seasons from 2002-2008 for 1-hr downwind areas of (a) 5-km (b) 20-km radii.
E1.2.5 Cold Cloud Visible Albedo

Figure E.06 Three aerosol levels and cold-cloud SWIRR from 2002-2008 for four seasons for 1-hr downwind areas of radii (a) 5-km, and (b) 20-km.
Figure E 0.7 Urban-rural differences for three aerosol levels and cold-cloud SWIRR from 2002-2008 for four seasons. One-hour downwind areas of radii (a) 10-km, and (b) 35-km.
Figure E 0.8 Three aerosol levels and cold cloud fraction from 2002-2008 for four seasons for 1-hr downwind areas of radii (a) 5-km, and (b) 20-km.
E1.2.8 Cloud Fraction for Cold Cloud Visible Albedo Urban-Rural Differences

Urban-Rural differences for 05-km Radius Cold Cloud Visible Albedo:
Cloud Fraction All Aerosol Levels from 2002-2008 by Season

Urban-Rural differences for 20-km Radius Cold Cloud Visible Albedo:
Cloud Fraction All Aerosol Levels from 2002-2008 by Season

Figure E 0.9 Urban-rural differences for three aerosol levels and cold cloud fraction from 2002-2008 for four seasons for 1-hr downwind areas of radii (a) 5-km, and (b) 20-km.
E1.3 Aerosol and All Cloud Visible Albedo

E1.3.1 All Cloud Visible Albedo

Figure E 0.10 Warm-cloud visible albedo for low aerosol one hour downwind of NYC of 65-km downwind-area radii.
E2  Seasonal Aerosol Climatologies

E2.1  Warm Cloud SWIRR and Cloud Fraction

E2.1.1  Warm Cloud SWIRR

05-km Radius Warm Cloud 3.9-μm Reflectance:
All Aerosol Levels from 2002-2008 by Season

20-km Radius Warm Cloud 3.9-μm Reflectance:
All Aerosol Levels from 2002-2008 by Season
Figure E 0.11 Three aerosol levels and warm-cloud SWIRR from 2002-2008 for four seasons for 1-hr downwind areas of radii (a) 5-km, and (b) 20-km.

E2.1.2 Warm Cloud SWIRR Urban-Rural Differences

Urban-Rural differences for 05-km Radius Warm Cloud 3.9-μm Reflectance:
All Aerosol Levels from 2002-2008 by Season

Urban-Rural differences for 20-km Radius Warm Cloud 3.9-μm Reflectance:
All Aerosol Levels from 2002-2008 by Season
Figure E 0.12 Urban-rural differences for three aerosol levels and warm-cloud SWIRR from 2002-2008 for four seasons. One-hour downwind areas of radii (a) 5-km, and (b) 20-km.

### E2.2 Warm Cloud Visible Albedo and Cloud Fraction

#### E2.2.1 Warm Cloud Visible Albedo

<table>
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<th>NY</th>
<th>Rural</th>
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<tr>
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<td>Winter Low</td>
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<tr>
<td>Fall Low</td>
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</tr>
</tbody>
</table>

05-km Radius Warm Cloud Visible Albedo: All Aerosol Levels from 2002-2008 by Season

NY
Rural
Figure E 0.13  Three aerosol levels and warm-cloud visible albedo from 2002-2008 for four seasons for 1-hr downwind areas of radii (a) 5-km, and (b) 20-km.

E2.2.2 Warm Cloud Visible Albedo Urban-Rural Differences
Figure E 0.14 Urban-rural differences for three aerosol levels and warm-cloud visible albedo from 2002-2008 for four seasons for 1-hr downwind areas of radii (a) 5-km, and (b) 20-km.

E2.2.3 Cloud Fraction for Warm Cloud Visible Albedo

05-km Radius Warm Cloud Visible Albedo Cloud Fraction:
All Aerosol Levels from 2002-2008 by Season
Figure E.0.15 2002-2008 warm cloud fraction averages for 1-hr downwind areas of variable-km radii for three aerosol classes, ordered by season: urban-rural differences.

**E2.2.4 Cloud Fraction for Warm Cloud Visible Albedo Urban-Rural Differences**

Urban-Rural differences for 05-km Radius Warm Cloud Visible Albedo: Cloud Fraction All Aerosol Levels from 2002-2008 by Season
Figure E 0.16  Urban-rural differences for 2002-2008 warm cloud fraction averages for 1-hr downwind areas of variable-km radii for three aerosol classes, ordered by season.

APPENDIX F: ADDITIONAL UHI CLIMATOLOGY FIGURES
F1 Averages over all seasons

F1.1 Warm-cloud SWIRR mapped climatologies

F1.1.1 Medium

Figure E 0.1 Warm-cloud SWIR reflectance at 3.9-µm for medium UHI one hour downwind of (a) NYC with averaged pixels, and (b) rural locations, of 65-km downwind-area radii.

F1.1.2 Low

(b) (c)
Figure E 0.2 Warm-cloud SWIR reflectance for low UHI one hour downwind of rural areas for (a) NYC and urban (b) downwind areas of radii of 65km.

**F1.2 Warm cloud visible albedo mapped climatologies**

**F1.2.1 high**

![figure](image)

Figure E 0.3 Warm-cloud visible albedo for high UHI one hour downwind of (a) NYC and (b) rural for downwind areas of radii of 65km.

**F1.2.2 Medium**

![figure](image)
Figure E 0.4 Warm-cloud visible albedo for medium UHI one hour downwind of (a) NYC and (b) rural for downwind areas of radii of 65km.

F1.2.3 Low

Figure E 0.5 Warm-cloud visible albedo for low UHI one hour downwind areas of radii of 65km.
F1.3 Warm, cold and all-cloud visible albedo over all seasons

F1.3.1 Numeric results

Figure F0.6  Three UHI levels and warm, cold, and all-cloud visible albedo over all seasons from 1999-2009 for 1-hr downwind areas of (a) 10-km and (b) 35-km radii.
F1.3.4 Numeric results for urban-rural differences

Urban-Rural differences from 1999 to 2009 Over All Seasons by Cloud Type

Figure F0.7 Urban-rural differences for three UHI levels and warm, cold, and all-cloud visible albedo over all seasons from 1999-2009 for 1-hr downwind areas of (a) 10-km and (b) 35-km radii.
Figure F0.8 Three UHI levels and cloud fraction for warm, cold, and all-clouds over all seasons from 1999-2009 for 1-hr downwind areas of (a) 10-km and (b) 35-km radii.
Figure F 0.9 Urban-rural differences for three UHI levels and cloud fraction for warm, cold, and all-clouds over all seasons from 1999-2009 for 1-hr downwind areas of 10-km radius.
F1.4 Cold cloud image climatologies

F1.4.1 High

Figure F 0.10 Cold-cloud visible albedo for high UHI one hour downwind of (a) NYC and (b) rural for downwind areas of radii of 65km.

F1.4.2 Medium

Figure F 0.11 Cold-cloud visible albedo for medium UHI one hour downwind of (a) NYC and (b) rural for downwind areas of radii of 65km.
F1.4.3 Low

Figure F 0.12 Cold-cloud visible albedo for low UHI one hour downwind of (a) NYC and (b) rural for downwind areas of radii of 65km.

F2 Seasonal Urban Heat Island Climatologies

F2.1 Warm Cloud SWIRR and Cloud Fraction

F2.1.1 Warm Cloud SWIRR

10-km Radius Warm Cloud 3.9- μm Reflectance: All UHI Levels from 1999-2009 by Season

- NY
- Rural
Figure F.0.13 Three UHI levels and warm-cloud SWIR reflectance from 1999-2009 for four seasons for 1-hr downwind areas of radii (a) 10-km and (b) 35-km.

F2.1.3 Warm Cloud SWIR Urban-Rural Differences

Urban-Rural differences for 10-km Radius Warm Cloud 3.9-μm Reflectance:
All UHI Levels from 1999-2009 by Season
Figure F.0.14 Urban-rural difference for UHI levels and warm-cloud SWIR reflectance from 1999-2009 for four seasons for 1-hr downwind areas of radii (a) 10-km and (b) 35-km.
F2.2 Warm Cloud Visible Albedo and Cloud Fraction

F2.2.1 Warm Cloud Visible Albedo

Figure F.0.15 Three UHI levels and warm-cloud visible albedo from 1999-2009 for four seasons for 1-hr downwind areas of radii (a) 10-km and (b) 35-km.
F2.2.2 Warm Cloud Visible Albedo Urban-Rural Differences

Figure F.0.16 Urban-rural difference for UHI levels and warm-cloud visible albedo from 1999-2009 for four seasons for 1-hr downwind areas of radii (a) 10-km and (b) 35-km.
F2.2.4 Cloud Fraction for Warm Cloud Visible Albedo

10-km Radius Warm Cloud Visible Albedo Cloud Fraction:
All UHI Levels from 1999-2009 by Season

35-km Radius Warm Cloud Visible Albedo Cloud Fraction:
All UHI Levels from 1999-2009 by Season

Figure F.0.17 Three UHI levels and warm-cloud fraction from 1999-2009 for four seasons for 1-hr downwind areas of radii (a) 10-km and (b) 35-km.
F2.2.5 Urban-Rural Differences for Cloud Fraction for Warm Cloud Visible Albedo

Urban-Rural differences for 10-km Radius Warm Cloud Visible Albedo: Cloud Fraction All UHI Levels from 1999-2009 by Season

Urban-Rural differences for 35-km Radius Warm Cloud Visible Albedo: Cloud Fraction All UHI Levels from 1999-2009 by Season

Figure F 0.18 Urban-rural difference for UHI levels and warm-cloud fraction from 1999-2009 for four seasons for 1-hr downwind areas of radii (a) 10-km and (b) 35-km.
F2.3 Cold Cloud Visible Albedo and Cloud Fraction

F2.3.1 Cold Cloud Visible Albedo

Figure F0.19 Three UHI levels and cold-cloud visible albedo from 1999-2009 for four seasons for 1-hr downwind areas of radii (a) 10-km and (b) 35-km.
F2.3.2 Cold Cloud Visible Albedo Urban-Rural Differences

Urban-Rural differences for 10-km Radius Cold Cloud Visible Albedo:
All UHI Levels from 1999-2009 by Season

Urban-Rural differences for 35-km Radius Cold Cloud Visible Albedo:
All UHI Levels from 1999-2009 by Season

Figure F 0.20 Urban-rural difference for UHI levels and cold-cloud visible albedo from 1999-2009 for four seasons for 1-hr downwind areas of radii (a) 10-km and (b) 35-km.
Figure F0.21 Three UHI levels and cold-cloud fraction from 1999-2009 for four seasons for 1-hr downwind areas of radii (a) 10-km and (b) 35-km.
Figure F0.22 Urban-rural difference for UHI levels and cold-cloud fraction from 1999-2009 for four seasons for 1-hr downwind areas of radii (a) 10-km and (b) 35-km.
Endnotes

1  http://www.nsof.class.noaa.gov/saa/products/welcome
   NOAA’s Comprehensive Large Array-data Stewardship System
2  Personal communication with Alex.Graumann Jan 12, 2011 (“10 bit” “32”); also F:\h\data\goes\calibration\G12post-launch-GOES Imager Visible Pre-launch Cal.mht
3  The table can be found at:
   http://www.star.nesdis.noaa.gov/smcd/spb/fwu/homepage/GOES_Imager_Vis_PreCal.php
4  This is equation 10 in
   http://www.oso.noaa.gov/goes/goes-calibration/vicarious-calibration.htm
5  The STAR calibration page G12post-launch-GOES Imager Visible Pre-launch Cal.mht is no longer up, a similar one that replaced it, is
   with the same table, and the main calculation showing the same correction as a single number.
7  http://www.oso.noaa.gov/goes/goes-calibration/gvar-conversion.htm
8  http://goes.gsfc.nasa.gov/text/imager.calibration.html
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