UNIVERSITY OF CAMBRIDGE
Department of Architecture

Fundamentals of Urban Heat Islands
Concise guide for architects and urban planners

Kanchane R. Gunawardena

M.Phil(*Cantab*) Ph.D(*Cantab*)

The Martin Centre for Architectural and Urban Studies
Department of Architecture University of Cambridge,
1-5 Scroope Terrace, Cambridge, CB2 1PX, UK

EPSRC
Engineering and Physical Sciences
Research Council

Abstract

There is wide consensus amongst climatologists that cities exhibit distinct climates and are typically warmer than their surrounding rural areas. This phenomenon, known as the urban heat island effect, results from unintentional alterations to urban surface properties. These modifications lead to increased absorption of solar radiation, reduced cooling due to slower wind speeds, and lower evapotranspiration rates. Its occurrence presents adverse consequences to the health and comfort of urban built environment occupants, while also increasing energy consumption ensuing from the measures employed to seek relief. These consequences are highlighted as likely to exacerbate further when combined with the existing trend of increasing temperatures from wider climate warming. Adverse heat-related impacts are thus on an upward trend and are gaining wider attention, with the imperative to develop and implement mitigation and adaptation strategies having already gained significant political determination and investment in recent years.

Literature has been reviewed here from various knowledge domains, including public health, urban climatology, potamology, limnology, climate change science, and urban planning to provide a concise guide for architects and urban planners to consider when designing and implementing climate-resilient built environments.

Keywords: Urban heat island mitigation; urban cooling; green space; blue space; evapotranspiration; evaporation.

EPSRC
Engineering and Physical Sciences
Research Council

Acknowledgement

The work included in this concise guide was funded by the UKRI, Engineering and Physical Sciences Research Council of the United Kingdom (1930753).

Contents

Abbreviations

ASC	Adaptation Sub-Committee (of the Committee on Climate Change - UK)
CCC	Committee on Climate Change (independent body established under the Climate Change Act 2008 to advise the UK Government)
CIBSE	Chartered Institution of Building Services Engineers (UK)
DECC	Department for Energy and Climate Change (UK)
EPSRC	Engineering and Physical Sciences Research Council (UK)
HVAC	Heating, ventilation, and air conditioning
LAI	Leaf Area Index
LUCID	Local Urban Climate Model and its Application to the Intelligent Design of Cities - EPSRC project

Nomenclature

Symbol	**Description**	**Unit**
$Q_{F_{B,l}}$	Flux generated by lighting, appliances, and HVAC (collectively)	$W \cdot m^{-2}$
$Q_{F_{B,g}}$	Subsurface conduction from buildings	$W \cdot m^{-2}$
$Q_{F_{B,m}}$	Metabolic emissions of building occupants	$W \cdot m^{-2}$
$Q_{F_{B,r}}$	Shortwave and longwave radiation gained by buildings	$W \cdot m^{-2}$
Q_{F_B}	Flux from buildings	$W \cdot m^{-2}$
Q_{F_M}	Flux from human metabolism	$W \cdot m^{-2}$
Q_{F_T}	Flux from transportation	$W \cdot m^{-2}$
Q^*	Net radiation	$W \cdot m^{-2}$
Q_A	Advection	$W \cdot m^{-2}$
Q_E	Evaporation (latent heating)	$W \cdot m^{-2}$
Q_F	Anthropogenic heat	$W \cdot m^{-2}$
Q_G	Conduction	$W \cdot m^{-2}$
Q_H	Convection (sensible heating)	$W \cdot m^{-2}$
Q_S	Heat Storage	$W \cdot m^{-2}$
T_r	Rural temperature	°C
T_u	Urban temperature	°C
$\Delta T_{u-r\ (max)}$	Urban heat island maximum intensity	K
ΔT_{u-r}	Urban heat island intensity	K
K^*	Net shortwave radiation	$W \cdot m^{-2}$
L^*	Net longwave radiation	$W \cdot m^{-2}$
$K \uparrow$	Outgoing shortwave radiation	$W \cdot m^{-2}$
$K \downarrow$	Received shortwave radiation	$W \cdot m^{-2}$
$L \uparrow$	Outgoing longwave radiation	$W \cdot m^{-2}$
$L \downarrow$	Received longwave radiation	$W \cdot m^{-2}$
C	Heat capacity	$J \cdot K^{-1}$
VHC	Volumetric heat capacity	$J \cdot m^{-3} \cdot K^{-1}$
k	Thermal conductivity	$W \cdot m^{-1} \cdot K^{-1}$
K	Thermal diffusivity	$m^2 \cdot s^{-1}$
I	Thermal inertia	$J \cdot m^{-2} \cdot K^{-1} \cdot s^{-1/2}$
α	Albedo	-
ß	Bowen ratio	-
ε	Emissivity	-
n	Porosity	-

Note: All city climate zones referred to in this paper correspond to the Köppen Climate Classification System.

Definitions

Advection: The horizontal movement of a fluid (mass movement of molecules containing energy and other variables) [1].

Aquifer: A body of porous rock capable of storing significant amounts of water, underlain by impermeable material, and through which groundwater transports [2].

Boundary layer: Generally defined as a layer in a fluid where energy, mass, and momentum transfer processes are influenced by properties of the underlying surface [3].

Climate: The average of weather events over a long period [4], typically discussed in relation to the Köppen Climate Classification System.

Comfort: Described as a state of physical ease and freedom from pain or constraint [5].

Convection, forced: Fluid movement that results from external forces [4].

Convection, natural/free: The vertical movement of a parcel of air (mass movement of molecules containing energy and other variables) being at a different density than the surrounding fluid (buoyancy) [4].

Convection: Principally refers to vertical motion that results in the transport and mixing of fluid properties [4].

Ecosystem services: Processes or materials that are naturally provided by ecosystems, such as clean water, energy, climate regulation, phytoremediation, and nutrient cycling [6].

Health: The World Health Organisation (WHO) definition describes it as '*a state of complete physical, mental, and social wellbeing and not merely the absence of disease or infirmity*' [6].

Heatwave: The World Meteorological Organization (WMO) definition describes it as *'when the daily maximum temperature of more than five consecutive days exceeds the average maximum temperature by 5°C, the normal period being 1961-1990'* (ww.metoffice.gov.uk).

Lapse rate: Decrease of an atmospheric variable (typically temperature) with height [1].

Leaf Area Index (LAI): Dimensionless ratio between a single leaf's surface area [m^2] and per unit of ground surface area [m^2], that describes plant canopies [3].

Limnology: Scientific study of freshwater ecosystems, particularly lakes [2].

Mean radiant temperature (MRT): Mean temperature of all the surfaces that surround an object [7].

Operative temperature (T_{op}): Combines air temperature (T_a) with radiant effects (T_r) to provide a more realistic representation of the temperature perceived by occupants within a space [8]. As air velocity increases, (T_{op}) tends towards (T_a), at air speeds of 0.1 m·s^{-1} or less (typical in buildings) it approximates to the following: $T_{op} \approx (T_r + T_a)/2$ [9].

Thermal alliesthesia: *'The hedonic qualities of the thermal environment are determined as much by the general thermal state of the subject as by the environment itself'* [10].

Thermal comfort: Described as '*the condition of mind that expresses satisfaction with the thermal environment'* [11].

Wellbeing: The Oxford dictionary defines it as a state of mental and physical health, as well as social wellness, satisfaction with their lives, and experiencing a good quality of life [12].

Wet-bulb globe temperature (WBGT): An index, calculated for in-shade areas that is a function of all four environmental factors affecting heat stress. It includes dry-bulb, naturally ventilated wet-bulb, and black globe temperature. Since the index is concerned with extremes of heat stress, CIBSE consider such conditions as beyond those required for thermal comfort, or acceptable levels of overheating [9].

URBAN HEAT RISKS

'**Heat**' describes a form of energy that is transferred from one object to another following a temperature gradient by the processes of conduction, convection, and radiation. '**Risk**' is described as a measure of the probability that something of value such as life, health, property, or ecosystems, experiencing harm or damage from a given hazard [6]. '**Heat risk**' therefore refers to the harm or damage that may be experienced to things of value owing to the distinct hazard presented by heat [13]. As the harm or damage suffered is relative to the degree of exposure to the hazard, heat is considered in terms of excesses beyond thresholds. In certain instances, these are absolute thresholds, while in others they are relative to the adaptation mechanisms in place (see Table 1).

The principal risks from excess heat are:

- Direct adverse effects on health and mortality resulting from heat stress [14,15];
- Indirect effects on health and mortality resulting from reduced air quality [16,17]; and
- Increased energy consumption and associated carbon emissions, resulting from the actions taken to mitigate the adverse effects of excess heat [18,19].

Table 1. Key temperature thresholds affecting human physiological functions.

Physiological conditions	Core body temperature
Death from heat stroke Cellular proteins are damaged, and cells die	>42 °C
Hyperthermia at upper limits Exercise and common fever at lower limits	37.8-to-40°C
Core body temperature (normal)	36.1-to-37.8°C
Hypothermia Impaired central nervous system function	30-to-35°C
Loss of consciousness	30°C
Death due to ventricular fibrillation	<28°C
	Skin temperatures
Human skin temperature	33°C (surface temp.)
Triggers pain receptors in the skin	46°C (surface temp.)
Tolerance from thermal insulation of the air layer around the skin (short duration, e.g., sauna)	85°C+ (dry-air temp.)

Sources: ASHRAE [11] and Kuht & Farmery [20].

Urban health

In addition to population intrinsic factors such as age, gender, and health conditions (exemplified by studies: [21,22]), as well as socioeconomic factors such as affordability of adaptation measures (e.g., [23–27]), the geographical significance of urban areas has been repeatedly identified by epidemiological research to present greater heat vulnerability

[22,28], (see Table 2). This sensitivity is principally attributed to the influence of the urban heat island effect [21], the significance of which has been emphasised for several preceding heat-related health events [29]. The dynamic nature of this phenomenon however makes it difficult to quantify their precise spatial and temporal significance to specific heat events and their adverse health consequences. Studies have nevertheless presented ample observations to highlight oppressive night-time temperatures as being of greater significance to mortality than higher maximum daytime temperatures [30], which corresponds to the typically experienced nocturnal peak of heat islands [31]. The epidemiological evidence base therefore presents reasonable consensus to suggest the adverse health impacts of excess heat to be already common in urban areas, and stress that they are likely to exacerbate further when combined with increasing temperatures from a warming climate.

As a secondary consequence of higher temperatures, air quality may worsen in cities to the extent that it presents significant risk to public health. A notable example is presented by the 2003 pan-European heatwave, where between 420-770 excess deaths were attributed to poor air quality [17]. Air pollution manifests during warm weather as smog, generated by the photochemical reactions of atmospheric pollutants. A study of the American city of Los Angeles (warm and temperate) found that for every 1 K temperature rise above 22°C, the occurrence of smog increased by 5% [16]. Smog results from higher concentrations of nitrogen dioxide and particulate matter (PM) building-up under stable atmospheric conditions (i.e., minimal wind generated disturbances) [32]. The health effects of PM are more significant than other smog pollutants, with prolonged exposure increasing the risk of cardiovascular diseases and lung cancer. Another notable smog pollutant is ground-level Ozone (O_3), which is increased with higher temperatures as nitrogen oxides (NO_x) and volatile organic compounds (VOCs) react in the presence of solar radiation [33]. A study of the United Kingdom has suggested that a 5 K increase in mean temperature could result in 500 extra annual deaths attributable to Ozone exposure alone [34]. The risk to urban inhabitants is heightened as the air polluted from the many anthropogenic activities such as fossil fuel combustion and manufacturing, is then presented with an enhanced thermal load from the heat island to intensify chemical reactions and generate such adverse by-products. The excess of heat is therefore not only a direct stress on human health (thermoregulatory mechanisms), but also a secondary one as the resulting pollutant by-products affect biochemical processes to cause short-to-long-term health complications [13,35].

Table 2. Key building occupancy groups and their vulnerability to excess heat.

Groups	Key considerations
Medical conditions	▪ Pre-existing physical conditions; cardiovascular; neurological; endocrine disorders (diabetes, hyperthyroidism, hyperpituitarism); skin disorders impairing sweating; and infections (respiratory, gastrointestinal, septicaemia) [22]. ▪ Influence of drugs and substances that compromise thermoregulatory processes (e.g., phenothiazines, antidepressants, diuretics, alcohol, and narcotics). ▪ Obesity [36]. ▪ Serious physical disabilities [37].
Mental health conditions	▪ Serious psychological disabilities [37]. ▪ Depression, dementia, Parkinson's Disease, or other compromised cognitive states [38]. ▪ Perception of vulnerability [39].

Groups	Key considerations
Older people	▪ Ageing (senescence) resulting in reduced thermoregulatory capacity; begins from around 50-years of age [40]. ▪ Increased levels of dependency and isolated living arrangements [24,41].
Children	▪ Increased levels of dependency, limited ability to thermoregulate, and higher potential for dehydration [21]. ▪ Children under four years of age, who are obese, taking medication, with disabilities, or complex health needs at increased risk. ▪ Vigorous physical activity during outdoor temperatures exceeding 30°C increases vulnerability [32].
Gender	▪ European studies suggest women to be more vulnerable than men, even after accounting for age [22]. ▪ Women aged ≥65 at higher risk due to the negative effect of the menopause on thermoregulation and cardiovascular fitness [22]. ▪ Men at greater risk of heatstroke due to higher physical activity and exposure to outdoor warmer weather [22].
Socio-economic status	▪ Not fully understood and varies with context [42]. ▪ Not commonly reported in European studies [43], while poverty or lower socio-economic status (i.e., inability to purchase air conditioning), and lower education levels reported in American studies [24].
Regional	▪ Excess mortality with increasing temperature is apparent at higher thresholds in warmer climates, compared to milder climates [22]. ▪ In the United Kingdom, lower mortality thresholds are observed in the north relative to the south, e.g., Northeast mortality threshold is 20.9°C, while for the Southeast it is 23.5°C [44]. The Heatwave Plan for England accounts for these variations by establishing region specific thresholds [32].
Urban	▪ Increased urban sensitivity is largely attributed to the heat island effect, although not easily quantifiable [21].
Occupancy patterns	▪ How building occupancy patterns relate to temperature peaks [45]. ▪ Isolated or communal occupation [42]; social networks [37]; and engagement with social capital [25].

Urban energy use

In the wider European context, energy use has been largely discussed in relation to heating loads required to safeguard health and maintain comfort during the cold winter months. This focus is reflected in the available studies that have assessed climate-loading influence on building energy demand, particularly in relation to those assessing urban areas. One of the earliest of such studies examining London for example (maritime temperate), had considered data from 1951-60 to identify a beneficial 10% reduction in annual heating degree-days in central London, relative to a rural site [46]. Recent studies of London have highlighted a relative annual heating load reduction of 22% for commercial buildings [47], while an estimated 13% benefit is provided to residential buildings [48]. The heightened environmental thermal load presented by the heat island has therefore been established to contribute a beneficial reduction in urban energy usage during the colder winter months.

The historical emphasis on adapting the built environment to colder winter climate loads has led to considerable progress in achieving an energy efficient space-heating dominated building stock [34,49–51]. This progress however is not evident when considering warmer summertime climate loads, as until recently excess heat had not been the foremost concern [52,53]. This lack of adaptation means that internal environments of many buildings have already been found to overheat in the summer months to present significant risk to occupant health from heat stress, as well as leading to loss of productivity resulting from thermal discomfort [34]. If passive heat mitigation measures are not utilised to address these adverse conditions, the alternative of using active mechanical systems is likely to increase energy consumption, resultant waste heat, and carbon emissions [19,54,55]. The heightened environmental thermal load presented by the heat island may therefore contribute to an increase in urban energy usage during the warmer summer months.

In the United States, mechanical cooling or air conditioning has long been considered as the principal solution for mitigating heat-related health risks [22]. Heat vulnerability mapping has been used to justify this proliferation, with regions and cities with the highest air conditioning prevalence demonstrated to have some of the lowest cumulative heat vulnerability [28]. The widespread use of mechanical cooling has also been influenced by a potent sociocultural qualification that has associated its use with comfort and modernity. This has aided in the proliferation of the technology from fewer than 2% of American households in 1955, to reach 87% by 2009 [56]. The association is further amplified in warmer climate cities where air conditioning is regarded as a necessity. Studies have demonstrated that for every 1 K rise in daily maximum temperature (above 15-20°C), peak urban electricity demand had increased by 2-4%, with mechanical cooling responsible for 5-10% of the overall demand [16]. Several major American cities have thus repeatedly demonstrated higher urban temperatures to significantly enhance their summertime energy demands, which in turn has contributed to net annual increases in urban energy consumption [18,57].

In the European context, the use of mechanical cooling is evident in certain warmer climate cities in the south of the continent, although to a significantly lesser extent than in the United States. Such cities have been aware of cooling demand implications resulting from enhanced loads presented by their heat islands and wider climate warming. A study of Athens (subtropical Mediterranean) for example, had demonstrated a heat island intensity of 10 K to result in the doubling of the city's building cooling loads [54]. In most northern European cities, there is little use of air conditioning for the time being (e.g., ~3% of UK residential stock, [58]), with concern for demand increases also having received less attention. This status quo however is expected to change as ever-worsening public health risks may eventually compel its widespread introduction to reduce heat vulnerability [59]. For the south of England for example, modelling studies have estimated continued climate warming to result in 29-42% of households acquiring air conditioning by 2050 [60]. Recent studies in the United Kingdom have as a result begun to analyse urban climate loading influence on cooling demands with greater accuracy, particularly for commercial buildings where substantial summertime demand increases are expected. For example, a study that considered a prototypical office building in London had found its prevailing heat island to already present an annual cooling load increase of 25% [47].

The working principle of air conditioning is heat rejection, which is to absorb heat from the interiors of buildings and to release it to the surrounding outdoor environment as anthropogenic heat [18,31]. Building heat rejection from air conditioning systems is considered as a growing source of urban anthropogenic emissions, particularly evident in American cities [61], and with anticipated growth in the United Kingdom [50,60,62]. In warmer climate cities, increased reliance on such technologies has led to marked increases in summertime building emissions [63]. A simulation study of Phoenix (semiarid) had demonstrated the waste heat released from such systems to have negligible effect near the surface during the day (although the maximum is released), while during the night they increased air temperature by >1 K [64]. de Munck *et al.* [65] explained this nocturnal significance with reference to the greater depth of the boundary layer heat island during the day leading to rejected heat rising further up into the atmosphere to minimise the effect at the surface, while at night the contracted canopy layer heat island traps the rejected heat much closer to the surface. Another complication is presented by systems that use evaporative cooling to exchange heat with the external environment [63]. The rejected moisture and increase in humidity that results impedes comfort, from which the population must then seek relief by increasing the use of energy intensive air conditioning. The positive feedback loop that results is likely to lead to the urban climate becoming an unpleasant setting, where continuity of comfort necessitates transition from one mechanically cooled space to another. Such unhealthy urban surroundings in turn discourage inhabitants from engaging with the outdoors (e.g., walking or using public transport), thereby encouraging air conditioned vehicle use that leads to further energy consumption, heat rejection, and pollution [13,19].

Public health experts have warned the dependency on the technology to be also affecting the long-term adaptation of populations, with those who have frequented such environments finding their vulnerability to heat heightened in events of power outages, resulting from excessive energy demand typically experienced during warmer weather events [15]. Avoiding or reducing air conditioning use should therefore be a primary objective, which is particularly significant in urban areas where such climate conditioned buildings are increasing in representation [19]. The extensive and convenient use of the technology could be summarised to add to environmental, economic, and social burdens, while diverting attention away from alternative low impact adaptive measures. Consideration of passive cooling measures relating to morphology and materiality features, use of vegetation and blue space features in the design of not only individual projects, but also urban developments is advocated as the sustainable approach to mitigating urban heat-related risks [15].

Urban climate and the built environment

Since amateur meteorologist Luke Howard's earliest observations of London's unique climate [66], the urban heat island effect has been investigated by numerous climate researchers, with notably greater interest from the 1950s onward (e.g., [1,18,46,57,67,68]). A substantial proportion of settlement-specific heat island observations have been presented by Anglo-European (e.g., [46,67,69]) and North American researchers (e.g., [1,70,71]). Their studies had notably considered atmospheric heat islands the most, with the surface manifestation addressed to modest extent (e.g., [72,73]), and the subsurface manifestation the least considered (e.g., [74–76]). With the studies that have considered atmospheric manifestations, Grimmond *et al.* [77] had highlighted the deficit of boundary layer observations relative to canopy layer observations. The reported heat island magnitudes are also not always comparable due to the many parameters that vary with each location and climate zone, as well as the varied methodologies utilised by different researchers to monitor, record, and report observations. A systematic review of 190 canopy layer studies published between 1950 and 2007 had found that nearly half of the heat island magnitudes reported to be scientifically inconsistent, with the two common areas of limitation being controlled measurement and transparency of the methods used [78]. In general, the majority of post-1960 studies have considered the assessment and/or the modelling and simulation of energy, mass, and momentum flux of urban climate domains as an essential task of their research framework, predicated on Sundborg's [67] formative definition of urban climate interactions in relation to their surface energy balance.

The heat island phenomenon is said to result from the inadvertent modification of the earth's surface properties that alters this energy balance, which accounts for the urban climate system's energy exchanges [1,67]. The partitioning and dynamics of this energy balance describes the distribution, intensity, and dynamic profile of the heat island experienced. The earlier described adverse effects on health and productivity, pollution, and increased energy consumption, combined with expected climate change is emphasised as a significant risk to the habitability of many future urban environments. As global urbanisation is on an upward trend [79], the imperative to better understand its features and mitigate its adverse impacts has gained greater potency in recent times [80].

While the mutual relationship between the urban built environment and its climate was first hypothesised by Howard [66], its specific aspects have taken many subsequent studies to clarify. These include associations to built environment density, the surface-to-volume ratio, height-to-width (aspect) ratio, and buildings-to-space ratio (i.e., sky-view factor) [1,46,81]. Street geometry has been considered by studies to be the most common and simplified representation of the urban built environment, the features of which have been found to offer strong correlations with the heat island experienced [82–84]. As an aggregated representation, the influence of urban 'grain' or 'texture' has also been assessed, particularly in relation to influence on the radiation balance [85–87]. Furthermore, a few studies have found strong correlations with city size, represented by their physical built environment extents as well as their population, with growth identified as having potential to increase the heat island effect experienced [88,89].

The dominant materiality of a city has been found to contribute significantly by several studies, mainly focusing on the material properties of albedo and heat storage, and their influence on the energy balance [18,71,90]. Urban surface features such as green space (e.g., [91–93]), and blue space (e.g., [94,95]), have also received attention, with contributions highlighted to present significant heat risk mitigation. A meta-analysis of urban green space studies by Bowler *et al.* [92] however had highlighted the evidence presented to be mostly based on observational studies of small numbers of green spaces, as well as the impact of specific greening interventions on the wider city and whether such effects are due to greening alone, as not sufficiently addressed. The Volker *et al.* [95] meta-analysis of urban blue space studies had noted research on temperature effects to be significantly fewer in comparison to green space studies, with those available focusing predominantly on daytime influence and dominated by remote-sensing based surface temperature observations (that disregard sensible to latent heat conversion) as opposed to air temperature observations, while the heterogeneity of types as also not well addressed.

URBAN CLIMATES

This chapter presents a description of the structures represented in urban climates, and how their interactions contribute to the formation and distribution of the heat island effect.

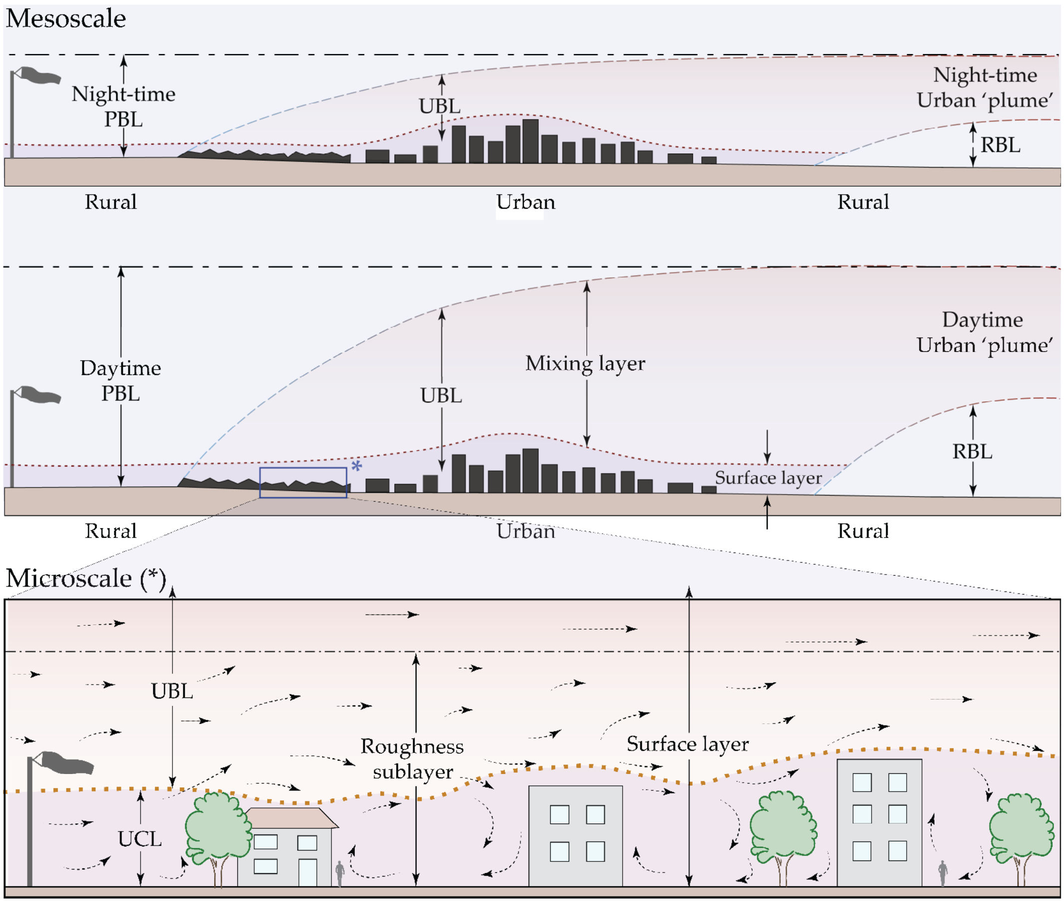


Figure 1. Boundary layer structures over an idealised city (at both day and night-time).

Climate structures (vertical)

The homosphere represents the lower atmosphere of the earth, where the weather variables and atmospheric gases are considered as mixed (extends to ~100 km in vertical elevation or altitude). Within this larger atmospheric region, four main subregions are distinguished and are identified in the descending vertical order as the lower part of the thermosphere, mesosphere, stratosphere, and the troposphere. As the lowermost region (extends to ~10 km), the troposphere contains ~75-80% of the planet's atmospheric mass, and nearly all its water vapour [4]. This layer is further partitioned in descending vertical order into the free troposphere and the planetary boundary layer. The latter planetary boundary layer (PBL) is the part of the troposphere that is influenced by contact with the earth surface, and its extents depend on the strength of the mixing generated by this surface (see

Figure 1). Its depth typically demonstrates diurnal and seasonal cycles. During the day, solar radiation generates strong thermal (convective) mixing that extends it to ~1-2 km in elevation, while at night the radiative cooling of the surface relative to the atmosphere causes the downward flux of heat, which suppresses mixing and contracts its depth to <100 m in elevation [1]. The same thermal processes contribute to greater mean depth during the summer, relative to winter. The increased surface roughness encountered in urban areas partitions this planetary boundary layer further into the urban boundary layer (UBL) and the canopy layer (UCL), in descending vertical order. The urban boundary layer is a mesoscale concept referring to the upper part of the planetary boundary layer, with its qualities influenced by the presence of an urban area at its lower boundary. The urban canopy layer in contrast is a microscale concept that describes the lower part of the planetary boundary layer consisting of the urban roughness elements (extends from the ground surface to the tops of buildings and trees), where the climate is dominated by the morphology and materiality of the immediate surroundings, and where people typically occupy. The urban canopy layer as a result represents the part of the atmosphere that is vital for ensuring good human health, comfort, and wellbeing in cities [70].

Climate scales (horizontal)

Although atmospheric phenomena exist as part of the earth continuum, meteorological and climatological studies class them into discrete spatial scales for ease of analysis. Most classification schemes use horizontal extents as the only measure, although little agreement exists on their precise limits. Oke [1] for example had identified reasonable consensus to describe the relevant scales in descending order as the planetary-scale (entire planet), large-scale/macroscale/synoptic-scale (10^5 to 10^8 m), mesoscale (10^4 to 2×10^5 m), local-scale (10^2 to 5×10^4 m), and the microscale (10^{-2} to 10^3 m); while Jacobson [4] had added the molecular-scale ($<10^{-3}$ m) to this spectrum to address the requirements of investigations such as those involving pollutants, where molecular diffusion is significant. In the assessment of urban built environments and their climates, these scales have been further modified and refined by various studies to address the spatial and temporal distinctions of interest.

Urban energy balance

> *"…the temperature of the city is not to be considered as that of the climate; it partakes too much of an artificial warmth, induced by its structure, by a crowded population, and the consumption of great quantities of fuel in fires..."*
>
> Luke Howard (1833, p. 2)

Luke Howard was the first climatologist to theorise that urban climates are governed by their surface energy exchanges [66]. Geographer Åke Sundborg [67] later explained the uniqueness of the urban climate and the heat island phenomenon in relation to the 'urban energy balance', which accounts for the incoming and outgoing energy flows from the urban surface system. Considering the First Law of Thermodynamics, which asserts that energy is always conserved, Sundborg [67] deduced that the energy absorbed by the urban surface system from solar radiation and the energy generated by human activity must be physically balanced. This equilibrium occurs through several processes that include the warming of air above the surface; evaporation of moisture; and heat storage in surface materials (see Equation 1). This complex interplay ensures that energy remains conserved within the urban environment, while the partitioning of this balance has consequence on how cities function (i.e., energy is used), and how their inhabitants flourish (i.e., ensuring good population health, thermal comfort, and wellbeing).

$$\textit{Net radiation } (Q^*) + \textit{Anthropogenic heat } (Q_F) = \textit{Convection } (Q_H) + \textit{Evaporation } (Q_E) + \textit{Heat Storage } (Q_S) \qquad \textit{Equation 1}$$

The radiation balance defines the energy availability within the climate system, Q^*.

For the daytime this radiation balance may be expressed as:

$$Q^* = (K\downarrow - K\uparrow) + (L\downarrow - L\uparrow) \qquad \textit{Equation 2}$$

$$Q^* = K^* + L^*$$

While for the night-time it may be expressed as:

$$Q^* = (L\downarrow - L\uparrow) \qquad \textit{Equation 3}$$

$$Q^* = L^*$$

where,

Q^*	Net all-wave radiation	L^*	Net longwave radiation
K^*	Net shortwave radiation	$L\downarrow$	Received longwave radiation
$K\downarrow$	Received shortwave radiation	$L\uparrow$	Outgoing longwave radiation
$K\uparrow$	Outgoing shortwave radiation		

The spatial variability of received shortwave ($K\downarrow$) and longwave radiation ($L\downarrow$) is determined by synoptic/large-scale atmospheric or earth-to-sun geometric relationships [4]. Latitude affects shortwave radiation, with greater received nearer to the equator than towards the poles. Higher surface elevation leads to lower atmospheric absorption, which translates to higher shortwave radiation and reduced longwave radiation received. Annual variation resulting from the earth-to-sun geometric relationship, facilitates greater shortwave radiation received in the summer than in winter. Diurnal variation resulting from the earth's rotation provides all shortwave radiation during the day, while none is available at night. For any similar latitude city, the difference in received shortwave and longwave radiation values are likely to be modest. Outgoing shortwave ($K\uparrow$) and longwave ($L\uparrow$) radiation however are modified by city-specific features such as the albedo (α), temperature, and emissivity (ε) of its atmosphere and surfaces, which contributes to greater variation [1].

Within the earth-atmosphere system, the annual radiation flows are balanced when all radiative processes are accounted. The surface subsystem however will be in surplus and the atmosphere in equal deficit [1]. This surplus of the surface is offset to the atmosphere by the transfer of energy by convection processes as sensible (Q_H) and latent (Q_E) heat, and conducted to or from the surface (Q_G). The relative capabilities of the surface and atmosphere to transport heat governs the exact partitioning of the radiative surplus or deficit, which demonstrates diurnal variation. Daytime heat dissipation from the surface is principally achieved by sensible and latent convection ($Q_H + Q_E$); while at night the radiation loss is principally replaced by conduction as an upward flux from the surface (Q_G), with the least contribution provided by sensible convection (Q_H). Although the conductive exchange (Q_G) is a significant short-term energy source or sink, when aggregated over the entire day its net contribution is typically minimal [1], and as a result is not included in the simplified form of the urban energy balance represented in Equation 1.

As the energy balance is applied to a scaled volume of the surface and atmosphere, the capability of this volume to either absorb or release energy must be acknowledged as an energy storage change (ΔQ_S). Accounting for this modifies the energy balance as follows:

$$Q^* + Q_F = Q_E + Q_H + Q_G + \Delta Q_S \qquad \textit{Equation 4}$$

If the energy storage ΔQ_S is positive, the surface system warms, while a negative value describes cooling. If the energy storage $\Delta Q_S = 0$, the volume exhibits no energy change ($Q_{in} = Q_{out}$), and thus no temperature change is observed. When the volume scale considered is not extensive, there is increased possibility of heat exchange occurring horizontally between adjoining surfaces with dissimilar energy partitioning. This net horizontal convective heat ($Q_H + Q_E$) exchange is represented by an advection term (ΔQ_A, see Equation 5), and its increased significance in such instances necessitates the consideration of a three-dimensional energy balance [1].

$$Q^* + Q_F = Q_E + Q_H + Q_G + \Delta Q_S + \Delta Q_A \qquad \textit{Equation 5}$$

Urban heat island effect

$$UHI\ \Delta T_{u-r} = T_u - T_r \qquad \textit{Equation 6}$$

The urban heat island (UHI) effect is described as the relative difference between urban (T_u) and rural (T_r) temperatures [1,89]. The phenomenon is best evident and most potent under synoptic-to-mesoscale anticyclonic conditions (i.e., large-scale atmospheric circulation analogous to a high-pressure system), when reduced wind velocities and cloud cover are typical. Its intensity ($UHI\ \Delta T_{u-r}$) is observed to be greatest in the summertime when increased solar radiation received increases the energy available within the urban system, and at night-time when heat storage release from urban form becomes the dominant heat source [1]. During the daytime however, lower, or even negative intensities may be experienced to present 'cool island' conditions.

The formation of an urban heat island is dependent on several climatic processes and may be described as a boundary layer or canopy layer occurrence. The former is governed by processes relevant at the local or mesoscale, while the latter by those at the microscale [70].

Heat island formation

During the daytime, solar radiation warms the land surface and the adjacent surface layer of the atmosphere to induce convective instability that results in air parcels or 'thermals' rising to the urban boundary layer, the process by which it expands to extend up to 1.5-2 km off the urban surface. As thermals rise, they mix with the atmosphere to form a boundary layer with constant temperature and almost neutral stability. The buoyancy-driven mixing process however drives warmer air to the top of the layer to form a stable capping inversion layer (i.e., warmer fluid above a relatively cooler fluid). This stable capping layer then becomes an almost impenetrable layer to thermals rising from below. The near impenetrability of this capping inversion layer is significant as it traps heat, water vapour, and pollutants released at the surface within the atmospheric domain of the boundary layer [1]. The difference between urban and rural conditions is characterised by the intensity and depth of the capping inversion layer. As the modification of surface properties in cities aid the release of more heat during the daytime, this causes its stable capping inversion layer to be warmer and thicker than in rural areas. This is referred to as the 'boundary layer heat island', and is the result of the relative difference in rural-to-urban 'warming'. It is mildly intense both day and night-time, displays no notable temporal features, and is mostly significant from a meteorological perspective [1,70].

At night-time, the land surface purges the heat absorbed from daytime solar irradiation. When the surface has achieved this radiative purging, this causes a downward flux of heat that cools the nocturnal atmosphere and in turn causes it to contract and attain a more stable temperature structure. The surface properties of rural areas enable them to purge heat rapidly, which presents little opportunity to generate substantially warm thermals. This results in the purged heat to settle into a near ground-level stable inversion. Urban surfaces however continue to emit stored heat to generate surface thermals, although at lesser intensity than during the daytime. The urban stable capping layer inversion therefore occurs at an altitude between the rural inversion layer and the urban boundary layer, typically at the top of the urban canopy layer (UCL). This difference in inversion altitudes is therefore attributed to the relative difference in rural-to-urban cooling, rather than heating [31]. The resulting urban inversion altitude is significant because it traps any subsequent rising thermals, water vapour, and pollutants rising from below to form the 'canopy layer heat island' [1,70]. Since the canopy layer represents the stratum of the urban climate that includes human habitation, this nocturnal canopy layer heat island is regarded as the most significant feature of the urban climate relevant to public health, thermal comfort, and built environment research.

Heat island categories

Although heat islands are discussed mostly in relation to atmospheric manifestations as above, they may be distinguished further based on the stratum of the urban surface studied to include surface and subsurface manifestations (see Figure 3). While all three are governed by urban surface-to-atmosphere energy exchanges, their interrelatedness in terms of spatial distribution, temporal variation, and intensities are not explicitly understood.

Atmospheric heat islands

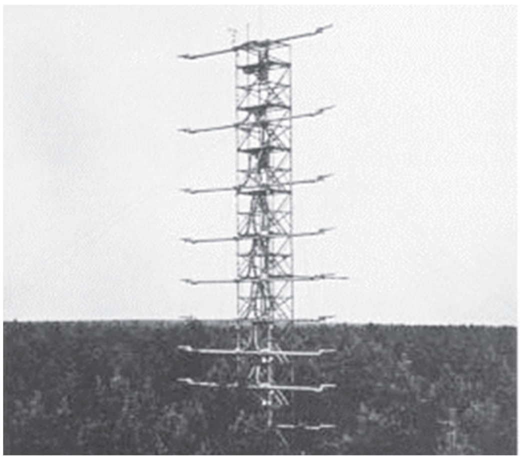


Figure 2. Weather tower over Thetford Forest.

There is strong association between surface and air temperatures, which is particularly evident in canopy layer observations immediately adjacent to the surface. Beyond this adjacency however, air temperatures are modified by convective mixing, and radiation absorption qualities enhanced by humidity and pollutants. Atmospheric heat islands as a result vary greater throughout the diurnal cycle than surface temperatures, with the timing of peak intensities dependent on the properties of urban and rural surfaces, season, and prevailing weather conditions [1].

Atmospheric heat islands are typically observed using fixed weather stations, mobile traverses, vertical sensing, or heat balance calculations (see Figure 3). Studies with fixed stations range from a single pair of urban and rural stations to multiple points distributed to a spatial grid. Single pair studies are subject to the characteristics of the selected sites, which may not always be representative of typical conditions [78]. Multiple stations offer better representation, with areas with established fixed stations enabling longitudinal analyses. Relative to fixed station approaches, mobile traverses are more economical and involve traversing on a predetermined route with readings taken at planned stops. Care must however be taken to avoid contamination from anthropogenic heat from transport infrastructure. Vertical sensing is utilised to record observations beyond the canopy layer and include boundary layer structures. Stationary towers with multiple vertical sensors, or instrumented vertical lift crafts such as balloons, helicopters, or drones may be utilised. While data obtained from latter vertical lift crafts could be used for cross-sectional studies, stationary tower data is of greater use for identifying longitudinal variations (e.g., Figure 2). Such studies however are uncommon owing to their high infrastructural cost, with recent approaches favouring drone usage given their rapid advancement and relative increase in affordability. Recent studies also avoid the need for collecting all data parameters by using energy balance calculations to determine residuals [1,96].

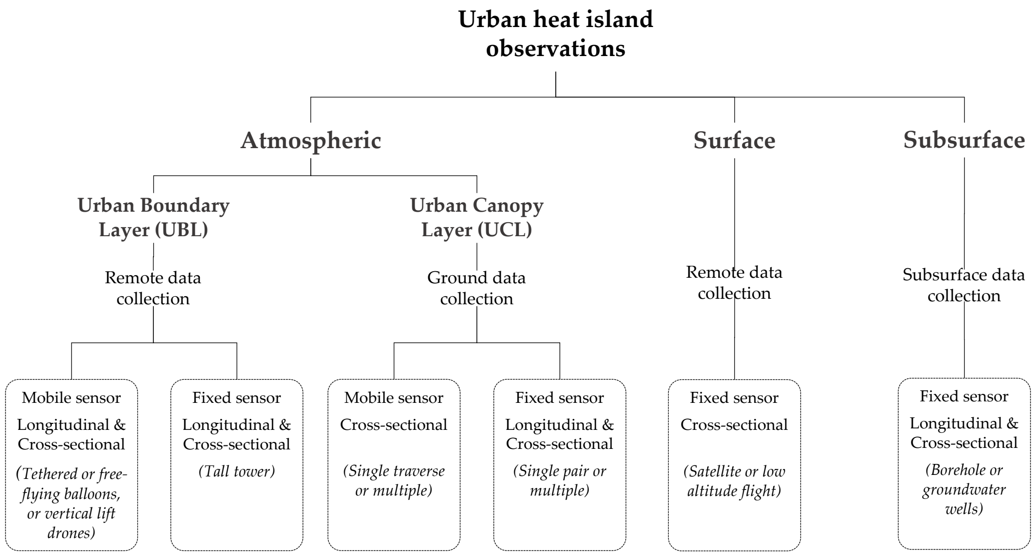


Figure 3. The study of urban heat islands.

Surface heat islands

Surface heat islands describe surface temperature differences between rural and urban areas. They are evident day and night-time, although warmer during the daytime (particularly in summer) as solar radiation heats surfaces, and relatively cooler at night-time as heat is purged back to the atmosphere [31,73]. Rural surroundings with shaded or moist surfaces typically remain closer to air temperatures, while exposed urban surfaces in dry summer conditions can reach up to 27-50 K warmer than the air [33]. The magnitudes of surface temperature differences are influenced by factors affecting radiation intensity and incidence (mainly shortwave radiation, $K{\downarrow}$), ground cover material properties, and weather patterns affecting the convective (Q_H) and evaporative (Q_E) cooling of surfaces.

Surface heat islands are typically measured using remote-sensing imagery that calculates the energy reflected and emitted from surfaces (see Figure 3). Images are taken from either satellite or aircraft (including drones), representing five different wavelengths of the visible and invisible energy spectrum. Satellite imaging typically presents the opportunity for two passes to provide cross-sectional comparison between day and night-time conditions. Low-altitude flight imagery in contrast provides flexibility to consider daily variation, with cross-sectional resolution dependent on the number of flights made. Both methods however require calmer weather with little to no cloud cover to produce clear images, which is generally satisfied by anticyclonic weather that is typical during strong heat island conditions. The findings of remote-sensing studies however should be treated with caution, as such images capture only cross-sectional data [95,97]. Furthermore, they disregard the sensible to latent heating conversion, while also neglecting the significance of vertical surfaces such as building facades and shaded areas [98].

Subsurface heat islands

Subsurface heat islands refer to belowground temperature differences between rural and urban areas. Principally affected by conduction heat flows, subsurface temperature differentials of up to 5 K have been reported [99]. This heat island manifestation is attributed to the cumulative effect of the mesoscale climate (i.e., surface and atmospheric heat islands), land-use modifications affecting conduction heat flows (Q_G), and natural and anthropogenic geothermal processes [74,75]. The adverse effects of it include modifications to groundwater chemical properties such as reduction-oxidation reactions that affect water quality, as well as modifying the diversity of aquifer bacteria and fauna to alter water purification and filtration processes. The manifestation may also present beneficial effects such as geothermal potential, while promoting biological decontamination in urban or industrial areas. It is typically studied by gathering subsurface temperatures from vertical boreholes, or groundwater observation wells used for monitoring water table quality and levels (see Figure 3). Vertical boreholes are typically used for cross-sectional studies, while pre-existing observation well networks facilitate longitudinal studies [99]. It must be noted that in terms of the research material available at present, this heat island manifestation has received the least attention in comparison to the other two.

Weather influence

Weather patterns significantly modify heat transfer processes between surfaces and the atmosphere, with wind velocity followed by cloud cover the main variables to consider [57,67,100]. Wind is generated when differential radiation balances across the earth-atmosphere system result in horizontal temperature variations that generate pressure differences and resulting atmospheric motion. Thermal energy from the sun is thus converted to kinetic energy of wind systems that transfer this energy into increasingly smaller scales of turbulent motion. Heat is eventually dissipated when the cascading scale of this kinetic energy reaches the molecular scale. For urban climates, both local-to-microscale and mesoscale wind systems are significant. The former wind systems are influenced by horizontal temperature variations resulting from contrasting surfaces across boundaries such as at the rural-to-urban interface, while the latter are influenced by the surface roughness of the urban built environment that alters both wind velocities with altitude and flow patterns. Increased wind velocity is the most sensitive variable in reducing heat island intensity, which highlights the significance of turbulent and advective activity (i.e., dynamic instability), in homogenising temperature structures [1].

Cloud cover is the second most significant weather variable to influence radiation exchanges and the heat island. Incoming shortwave radiation ($K\downarrow$) is affected by absorption and reflection from cloud tops, while net longwave radiation (L^*) is affected by their high absorption and emitting efficiency attributed to their effective status as black bodies. Increased cover reduces both shortwave penetration (heating), and net longwave radiation losses (cooling). Cloud type and the degree of cover is significant, with lower-altitude and thick 'stratus' more influential than a comparable extent of higher-altitude and thin 'cirrus' clouds. With complete overcast skies including low altitude thick cover, shortwave radiation may be reduced to 10% of the cloudless value, while partly cloudy skies add to short-term variability that aids the development of horizontal temperature stratification [1]. The overall influence of cloud cover is to moderate the variation of the surface radiation balance, which in turn moderates the diurnal temperature range experienced [89].

Both wind and cloud cover are implied in the term 'atmospheric stability', a measure of the relative tendency for air to transport vertically. General atmospheric stability has strong correlation with heat island intensity [100]. High wind velocity and cloud cover are observed to increase general atmospheric stability, and thereby reduce vertical transport. High wind velocities achieve this by introducing advection and turbulence (i.e., dynamic instability), while high cloud cover reduces the potency of convective thermals to result in reduced vertical transport. In contrasting calmer weather with low wind velocity and cloud cover, it is common for the planetary boundary layer to be considered (overall) as unstable during the daytime, and stable by night-time. However, this overall atmospheric stability must be considered as being made-up of several layers of differing stabilities. As discussed earlier, the formation of a heat island includes convective rise from an unstable surface layer, followed by an almost neutral mixed layer, which is then capped by the stable inversion layer that is the heat island [1].

Macro/large-scale high-pressure/anticyclonic systems affect temperatures in the planetary boundary layer, by permitting air to subside and warm on top of cooler surface air to create a subsidence inversion [4]. This descending warmer air evaporates clouds to increase shortwave ($K\downarrow$) penetration, which in turn generates the instability and vertical transport ideal for heat island formation. Conversely, low-pressure systems of similar scale cause air to rise and cool to generate clouds and hinder shortwave radiation penetration. The formation frequency of macro/large-scale high-pressure/anticyclonic systems is significant for assessing climate change influence on future heat island growth. A global trend of enduring and higher frequency of such systems could result in the increase of heat island magnitude in the future [101]. The exact association between the two phenomena however remain uncertain. This is due to the difference in resolution considered in climate change scenario analyses that typically disregard urban areas [69,102], given the applied surface energy balance schemes having been developed to capture rural land use contributions than building generated influences [103]. Notwithstanding this typical shortcoming, recent regional climate model simulation results of the United Kingdom that had explicitly accounted for urban grid cell influences, had suggested heat island intensities as likely to plateau with projected climate change [101].

Geographic influences

Heat island formation is significantly influenced by the geographic location and the surface topography and climate structures that describe a given city. Coastal or inland locality, altitude, presence of large green and blue spaces, and surrounding orography are all features that could present modifications to the energy balance and thereby influence the intensity and spatial and temporal variations of the heat island experienced. In coastal locations, coastal breeze systems of converging horizontal wind fronts that generate forced convection and humidity alterations, may contribute specific variations to differentiate their heat islands [104,105]. Similarly, cities in high altitudes have reduced atmospheric radiation absorption [1]; those in valleys with surrounding topographical barriers are affected by forced convection from airflow rising along such features [4]; while those with large areas of green and blue space features are affected by a pronounced latent flux [18]. Identifying the relevance of such geographical influences and their associated processes should thus be accounted in any attempt to characterise a given city's climate interactions.

BUILT ENVIRONMENT INFLUENCE

Although natural phenomena affect the energy balance of all types of landcover, anthropogenic modifications and activities present the predominant influence in cities. The features that characterise an urban built environment, the increased surface area and roughness from its morphology (mechanical influences); and use of engineered materials, improved drainage, and increased heat emissions from energy-use (collectively thermal influences), affect all components of the energy balance (Equation 1, p. 13). The formation of a heat island is the result of the net positive thermal balance of such alterations [1].

Mechanical effects of built form

Convection is the principal process by which the daytime energy surplus of the surface is transported to the atmosphere. It transports sensible and latent heat, other variables such as carbon dioxide and pollutants, and extracts momentum from mean atmospheric flow [1]. Convective instability of the atmosphere arises when natural convection is dominant, while dynamic instability arises when mechanical turbulence is dominant in the planetary boundary layer. With the former, the heating of the surface from solar radiation produces rising buoyancy-driven thermals, and when the generated vertical motion is dominated by this form of thermal turbulence, the boundary layer is said to be in a state of natural convection. In contrast, when wind flow passes over surface roughness elements such as buildings, mechanical turbulence is generated. This turbulence mixes mass, energy, momentum, and other variables vertically and horizontally, and if it principally dominates vertical motion, the boundary layer is said to be in a state of forced convection. Mechanical turbulence resulting from surface roughness represents only one form of forced convection. Specific conditions such as when airflow rises along a topographical barrier, or when horizontal wind fronts converge and rise (as discussed earlier), also generates forced convection [4]. Typical atmospheric convection profiles however are not so distinctly experienced, given the dynamics of varied wind velocity and direction; wake production from obstructing bodies; the distribution diversity of sources and sinks of mass, energy, momentum, and other variables; and surface movement, all contributing to both natural and forced convection coexisting as a 'mixed' regime [1,3].

Mechanical turbulence is induced by built environment surface roughness when wind drag and shearing stresses generated at the tops of surface objects produce groups of eddies of different scales [1]. This process blurs the simplified representation of the urban climate that includes the urban canopy layer and boundary layer discussed earlier, to present a roughness sublayer that describes the distinct spatial diversity of mean flow generated by individual roughness elements (Figure 1, p. 11). This roughness sublayer is identified by studies to extend up to ~2.5-3 times the height of built environment structures and includes the canopy layer depth [106]. Atmospheric stability affects its vertical extent, with generally higher depth evident for unstable conditions when convective instability contributes to the turbulence structure. Atmospheric flow in this layer is affected by factors such as differential mass, energy, and momentum flux generated from various sources and sinks (e.g., green, or blue space), and wake diffusion generated by turbulent wakes evident behind individual roughness features such as buildings. The magnitude of these wake diffusion eddies are associated to the dimensional parameters of roughness features, and is particularly significant for dispersed (low density) urban arrangements [106].

Dense urban built environment arrangements act as windbreakers to decrease mean wind velocity, with some studies highlighting up to 60% reductions [57]. This reduction occurs when the mean kinetic energy of wind flow is transformed by the interaction with increased surface roughness, into turbulent kinetic energy. Greater roughness of built-up areas and vegetation exerts greater flow transformation, which also increases turbulent diffusivities to enhance the sensible and latent flux regardless of temperature and vapour gradients [1]. Strong winds therefore promote turbulent mixing (i.e., dynamic instability) that inhibits the development of strong temperature stratification, while reduced velocities have the opposing effect and less heat as a result is transferred from the surface to the atmosphere [106]. Surface roughness generated forced convection influences however decrease more rapidly with altitude than the effects of thermal turbulence. Mechanical effects are therefore mostly dominant near the surface, and at greater altitudes thermal turbulence effects are of greater significance. This is reflected in the diurnal wind profiles of the atmosphere. During the daytime, most of the upper boundary layer is dominated by free convection, with typically large eddy structures resulting from thermals. While at nighttime, the stability of the atmosphere and lack of thermal turbulence from thermals mean that the convective mixing evident in this shallower boundary layer is entirely mechanical in source, including relatively smaller eddy structures [1].

Thermal influences of built form

Thermal effects that influence the urban climate are generated by geometrical and material features, as well as activities (i.e., energy use) that occur in the built environment.

Geometry and arrangement

Geometry influences the energy partitioning in the energy balance by modifying the net radiation balance. The key processes involved are obstructing incidence (shading) and enabling reflections (trapping). An arrangement that achieves a high aspect ratio (i.e., height-to-width ratio) can enable both these processes in opposing terms, with the net result determining the canopy layer temperatures experienced. In street canyons, the most simplified representation of urban geometry, buildings on either side shade their lower levels and the street surface during the day to limit direct shortwave ($K\downarrow$) and longwave ($L\downarrow$) radiation penetration and absorption. This canyon 'shading effect' can lower daytime temperatures experienced at the street level, at times to the extent that they may be significantly lower than surrounding urban and even rural areas. The relatively higher thermal inertia of urban materials means that lower daytime heat absorption translates to lower levels of longwave energy purged back to the atmosphere. This in turn serves to reduce the intensity of the nocturnal heat island [84]. Oke (1988) highlighted the significance of this shading effect to increase with latitude, and to be pronounced greater in winter when solar angles are lower. Furthermore, it is observed to increase with canyon aspect ratio and when oriented on the east-west rather than north-south axis, all of which determine the degree of shortwave and longwave radiation penetration [83].

High aspect ratio canyons and areas with tall building clusters tend to trap radiation by reflecting net shortwave (K^*) and net longwave (L^*) radiation flows (surfaces, atmosphere, and clouds) from surface to surface, leading to higher proportions of absorption [87]. This trapping influence has been assessed in the broader context of the city with reference to its general grain or texture (i.e., variation in urban arrangements). Complex urban grain with cavities or voids such as courtyards have been found to trap greater radiation than an open city with large blocks. An experiment utilising hypothetical street canyon models had identified a 20% radiation absorption increase attributed to enhanced surface heterogeneity, in comparison to a reference level surface [85]. Another modelling study had revealed that accounting for albedo, urban geometry could absorb up to 40% more shortwave radiation (K↓) than a comparative smooth reference surface (Steemers et al., 1998). The complexity of this urban grain affects the degree of radiation absorbed [83]. The same study considered sample urban fabrics from Toulouse (humid subtropical) and Berlin (maritime temperate) to find that the reduction in reflectance between the models varied, with 40% for Toulouse with its narrow streets and buildings, in contrast to 15% for Berlin with its wider open spaces [87]. The trapping effect of geometry is also significant for obstructing the release of longwave (L↑) infrared radiation back to the atmosphere (particularly purged by urban form at night), thereby contributing to a net longwave radiation (L*) increase. Urban areas with building clusters and deep canyons have as a result been shown to cool significantly slower, thereby contributing to an increase in the nocturnal heat island experienced [82].

Whether the shading or trapping effect becomes dominant depends on both the availability of shortwave radiation received (K↓; which is dependent on season, latitude, cloud cover, etc.), and the timing of the nocturnal heat island formation. A modelling study had found the shading effect to be significant at the beginning of the night, while the trapping of longwave radiation (L↑) later on in the night had presented a moderating influence on the heat island [84]. In addition to geometry considerations, materiality is highlighted as a key factor in determining the net effect of radiation flows. The relative daytime cooling effect of canyons for example can be further encouraged by increasing the albedo of surfaces (discussed below), with a recent study measuring potential reductions in air temperatures of up to 3-4 K with the use of lighter coloured surfaces [107].

Materiality

The materiality of the built environment influences the surface energy balance by affecting both net radiation and heat storage terms. The radiative properties of materials are considered as emissivity and albedo, while storage properties are affected by heat capacity (C) and thermal conductivity (k). A surface's ability to reject heat or emit longwave ($L\uparrow$) infrared radiation is measured as thermal emissivity (ε). As materials with high emittance values release heat more readily, the faster their surface cools [4]. Excluding metals, most other materials in urban environments tend to have high thermal emittance values (>0.70).

Solar reflectance or albedo (α) is regarded as the principal determinant of built environment surface temperatures, and is defined as the percentage of solar energy reflected by a surface. Higher the albedo of a material, the greater the solar energy that is reflected from its surface [1,4,18,90]. Material colour is correlated with albedo given that 43% of solar energy is in the visible wavelengths (400-700 nm), with lighter coloured surfaces presenting higher values (α >0.7) than darker surfaces (α <0.2) [71]. Albedo affects urban energy use both directly and indirectly. As for building specific energy use, the modification of radiation absorption directly affects heat transfer into occupied areas, thereby affecting indoor cooling loads. Indirectly the same process affects surface temperatures by reducing the convective heat flux and downwind ambient air temperatures. Reduced surface absorption from higher albedo eventually translates to reduced intensity of longwave radiation purged back to the atmosphere. This aids to reduce canopy layer temperatures particularly at night, which in turn reduces the thermal climate load on buildings and their resulting energy demands [18]. Albedo significance to surfaces vary with orientation and latitude. In tropical climates, the roof is the critical surface in sensible heat exchanges, while moving towards the poles increases the significance of vertical surfaces facing the equator [71].

Table 3. Typical albedo values and ranges for selected land-uses, from Taha *et al.* (1988).

Land use	**Albedo (approx.)**	**Land use**	**Albedo (approx.)**
Low-density residential	0.20	Park or green area	0.16
Medium-density residential	0.23	Urban areas	0.10-0.27
High density residential	0.25	Motorway or streets	0.30
Office buildings	0.22	Open green surfaces	0.35
Industrial	0.26	Seasonal parks	0.15

Heat storage in urban developments is influenced by the material properties of heat capacity and thermal conductivity. Heat capacity (C), commonly referred to as thermal mass, is the materials ability to store heat. The ease by which heat penetrates a material is addressed by thermal diffusivity (K), which is a function of thermal conductivity (k) and volumetric heat capacity (VHC). A higher value of diffusivity indicates that heat reaches deeper into the material with the temperature remaining constant [4]. Thermal inertia (I) is the measure of the responsiveness of a material to temperature variations. Materials with a high heat capacity also have high thermal inertia, meaning that their temperature fluctuations throughout the diurnal cycle are minimal [2]. Many urban materials tend to have higher heat capacities, thermal diffusivity, and thermal inertia than those found in rural contexts. In any urban development, the applied materials and their properties of emissivity and albedo, together with heat capacity and thermal conductivity, determine how solar energy is reflected, absorbed, and emitted by surfaces. The properties of the dominant material within an urban setting may therefore affect the intensity and timing of when the heat island peak is likely to be observed. Cities made of predominantly timber or earth (i.e., lower diffusivity values), are thus likely to reach their heat island peak intensity soon after sunset, while stone or concrete dominant cities (higher diffusivity values), are unlikely to reach their heat island peaks until sunrise [13,108–110].

A material structural property significant for heat dissipation from surfaces is porosity (n), defined as the ratio between void spaces to its total bulk volume. Higher porosity materials contain more air when dry, which results in lower thermal diffusivity. Subject to their albedo, this typically makes the thin surface layer warmer than the subsurface material bulk. Differential warming of this nature results in relatively higher surface temperatures, which under low wind velocities and convective instability contributes to the formation of strong thermals. Adding water to such surfaces marginally increases diffusivity, as the water significantly increases their heat capacity [1]. The principal cooling effect is achieved when the voids absorb water at the surface layer and cools by evaporation. Surfaces common to urban areas however are generally impervious (i.e., have lower porosity) and encourage faster water runoff that not only hinders this evaporative cooling contribution, but also adversely influence other subsurface processes [18].

The selection of materials for constructing urban developments is influenced by many other factors in addition to thermal properties [111]. Buildability and assembly detailing, economic cost and affordability, supply chains and availability, regulatory requirements, cultural and historic context considerations, and aesthetic demands, can all influence the materiality of a development or even the character of entire cities, subject to which consideration attains primacy. It is worth noting that materiality is an aspect of existing built environments that can be reasonably altered to enhance heat mitigation, undoubtedly to a greater degree of practicability than substantial alterations to existing urban geometries.

Human activity

A significant proportion of the energy consumed by the many activities in cities is eventually released to its climate as thermal waste. This waste thermal energy is referred to as 'anthropogenic heat' and is expressed as a flux for a given area (Q_F). It typically includes the three main contributing energy flux from, buildings (Q_{F_B}), transportation (Q_{F_T}), and human metabolism (Q_{F_M}), (see Equation 7). For large cities in industrialised nations, conservative urban anthropogenic emissions estimates range between 5-100 W·m^{-2}. The value varies given the complexity of the city, season, and diurnal cycles of activity. The complexity of London for example provides for a range of flux values across the different densities of human activity. A study had identified that over 50% of this city experiences annual heat flux of less than 8.0 W·m^{-2}, while only 2.5% experiences values greater than 50 W·m^{-2}; and where the density of activity is greatest as in the City of London, anomalous extreme values of greater than 210 W·m^{-2} could be expected [112].

$$Q_F = Q_{F_B} + Q_{F_T} + Q_{F_M} \qquad \textit{Equation 7}$$

The influence of season is significant in cold climate cities, where gains are generally larger in the winter due to intensive heating loads than in the summer. A study of American city cores for example, found anthropogenic heat emissions to range between 70-210 W·m^{-2} in winter, while in the summer the range was only between 20-40 W·m^{-2} [18]. Generic daily emission profiles tend to present local peaks in the morning and afternoon corresponding to peaks in building energy use and transportation. Weekday usage is also higher than weekends and holidays, when reduced activity is expected [63]. Such temporal variability is particularly significant when assessing microclimates, as the diurnal cycle for some activities may highlight short-term peaks in localised areas. A study of London for example had identified such extreme peaks of up to 550 W·m^{-2} [113].

Anthropogenic heat emissions influence the urban energy balance and contribute to the formation of heat islands by adding thermal energy to the urban system. Several studies have highlighted emissions to make significant contribution to the heat island magnitudes experienced. As examples:

- A study of Tokyo (humid subtropical) spatially mapped and numerically modelled thermal interactions to estimate that most of its nocturnal summertime heat island (2-3 K) was due to anthropogenic emissions [114];
- A simulation study of California (Mediterranean-to-semi-arid) had demonstrated that in a large city core, anthropogenic emissions could contribute to heat islands of up to 2-3 K [18]; and
- A study of London (temperate oceanic) had found that from the city's rejected anthropogenic emissions, around two thirds contributed to increasing the sensible heat flux that adds to its heat island [113].

Thermal energy produced and rejected by human activity can be partitioned as sensible and latent emissions. Sensible anthropogenic heat emissions can be directly released to the urban climate by exhaust chimneys and heating, ventilation, and air conditioning (HVAC) equipment. The indirect transfer of sensible heat however is more complex and only represented in the energy released from the building component (Q_{F_B}). Several processes are involved in this heat transfer. Heat within buildings is generated by lighting, appliances, and HVAC loads (collectively represented by $Q_{F_{B,l}}$), and by the metabolic emissions of their occupants ($Q_{F_{B,m}}$). These loads are influenced by environmental thermal loads that include shortwave and longwave radiation gained by the buildings ($Q_{F_{B,r}}$).

$$Q_{F_B} = Q_{F_{B,l}} + Q_{F_{B,m}} + Q_{F_{B,r}} + Q_{F_{B,g}} \qquad \textit{Equation 8}$$

The materiality and assembly of components of buildings, and how they interact with the surrounding climate, determine the degree of resistance (i.e., thermal inertia) that they offer to the transfer of generated heat energy through to the outdoor environment. Subject to this thermal inertia, the generated energy is indirectly transferred by conduction through the building envelope, followed by convection and radiation to the urban atmosphere, while a proportion of energy is also dissipated to the subsurface through substructure conduction ($Q_{F_{B,g}}$) [112]. This collective indirect transfer of heat has a time lag and varies in magnitude between actual energy consumed in buildings ($Q_{F_{B,l}}$), and the thermal energy rejected (Q_{F_B}). The energy transfer to the atmosphere may also be larger than the energy consumed, as it includes a proportion of the energy absorbed by the envelope that is purged back to the atmosphere ($Q_{F_{B,r}}$). The assumption that consumed energy ($Q_{F_{B,l}}$) is equal to energy released (Q_{F_B}), neglects the time lag and variance in magnitude of indirect sensible heat transfers, which in turn may lead to an over or underestimation of the rejected building component (Q_{F_B}). Furthermore, if anthropogenic latent emissions from evaporative cooling mechanisms (air conditioning), and the chemical reaction of hydrocarbon combustion from relevant sources are neglected, the rejected building component (Q_{F_B}) could be underestimated [63]. The significance of accurate quantification was stressed by a study of London, which had estimated Q_{F_B} to represent 80% of the ~150 terawatt-hours of waste energy that is annually rejected across the city [112].

GREEN AND BLUE SPACE INFLUENCE

The previous chapter discussed the influence of built environment morphology and materiality in modifying the urban energy balance and the resulting heat island. This section discusses the remaining features of green and blue space distribution and their contribution to the evaporative partitioning of the urban energy balance [115,116].

Evaporative cooling of the urban surface

The hydrological cycle is related to the surface energy balance by the evaporation of water that transfers energy from the surface to the atmosphere [86]. The combined evaporation from surface water, soil moisture, and transpiration from vegetation is described as 'evapotranspiration'. Annual global evapotranspiration is estimated to convert around 22% of the total solar energy received at the top of the earth's atmosphere [117]. A reduction in this contribution therefore affects the partitioning of the surface energy balance, as heat that would have otherwise been converted by this process instead contributes to the warming of the atmosphere and climate.

$$ß = Q_H / Q_E \qquad \textit{Equation 9}$$

To mitigate the adverse effects of excess heat, evapotranspiration can be increased by the addition of vegetation and significant waterbodies to an urban surface. The addition of such green and blue features enhances the conversion of sensible heating (Q_H) of the surface to latent heating (Q_E). The ratio of this flux is defined as the Bowen ratio (ß, see Equation 8), with average values ranging from ~0.1 for tropical oceans, ~0.4 to 0.8 for temperate forests and grasslands, ~2.0 to 6.0 for semi-arid areas, and greater than 10.0 for arid deserts [1]. Evaporative cooling of a surface occurs when this Bowen ratio is reduced, as the latent heat flux is increased (ß < 1). When the ratio presents a negative value as in certain arid climate green and blue space areas, the latent heat flux from the surface dominates to the extent that it generates a sensible flux inward to create a 'heat sink', which is commonly referred to as the 'oasis effect' [18].

Green space

Green spaces are represented by forests, parks, street trees and verges, private gardens, green roofs and walls, or any vegetated area that provides multiple ecosystem services to the urban environment (Figure 4). In addition to providing benefits that include reducing surface water run-off, enabling sustainable drainage, flood alleviation, increasing biodiversity, and general aesthetic and wellbeing enhancements, green space has been found to contribute to cooler microclimates [34]. They are as a result regarded as essential environmental capital that can be utilised to mitigate the adverse effects of heat islands, extreme heat events, and climate change [118]. A study of Glasgow (maritime temperate) for example, had estimated that an increase in green space of around 20% could serve to eliminate between one-third to half of the city's expected heat island effect in 2050 [119]. The introduction of such strategically planned interconnected networks of green space offering social, economic, ecological, and climate resilience benefits is therefore promoted in city-planning discourse as 'green infrastructure (GI) enhancements' [120].

Englische Anlagen park
Bern, Switzerland

Stavros Niarchos garden
Athens, Greece

Street trees
Barcelona, Spain

Private allotments
Bern, Switzerland

Figure 4. Examples of urban green infrastructure or greenspace features.

Properties of vegetation

Vegetation differs from other materials found in urban areas in terms of moisture content, thermal properties, and aerodynamics [93]. These unique properties allow vegetation to modify temperatures through different yet complementing processes that prevent their immediate surroundings from being warmed (relative cooling), and in certain conditions offer as enhanced cooling effect (heat sink). The processes directly influencing the urban energy balance relate to transpiration, shading, and surface roughness modification, along with the indirect influence from pollution filtering and surface runoff reduction [115].

Transpiration

Transpiration describes the most discussed cooling process, whereby water is transported through the plant and evaporated at their aerial foliage (stomatal apertures), by absorbing solar radiation that increases latent rather than sensible heat to keep the foliage and the surrounding atmosphere relatively cooler [71,121]. For most vegetated systems, 99% of the water absorbed and over half the energy absorbed is typically used for transpiration [1]. Vegetation type influences the cooling potential offered. While most C3 photosynthesising plants in cool and wet climates transpire significant volumes of water, relatively less is transpired by the less common C4 photosynthesising plants that are adapted to warmer climates. Drought-tolerant plants (with high stomatal resistance) typically found in arid climates that utilise Crassulacean Acid Metabolism (CAM) photosynthesis, minimise water loss by keeping their stomata closed during the day (open at night), which in turn provides reduced cooling as a result of their near negligible transpiration rates [91]. In most abundantly found plant types (C3 and C4), leaf stomatal apertures are typically closed in the absence of solar radiation. Latent cooling from transpiration is therefore principally relevant for daytime rather than night-time energy exchanges [3,122]. The rates of transpiration achieved during the day however depend on the vegetation properties of crown area, leaf area index (LAI), height of the canopy, stomatal resistance, and hydraulic resistance of the shoot and root; as well as soil conditions described by dryness, compaction, and hydraulic conductivity [123]. Transpiration effectiveness is also influenced by the background climate, with rates regulated by reducing or closing stomatal openings to prevent excessive water loss (reduction in moisture content affects guard cell turgidity, compelling them to become flaccid and close) [3]. Reduction in cooling effectiveness of plants subsequent to protracted heatwaves or drought conditions is therefore expected [124].

Shading

Shading from vegetation keeps the surroundings cooler by acting as a solar radiation interceptor that reflects and absorbs radiant energy, which limits shortwave (K↓) absorption by other urban surfaces, and their eventual longwave (L↑) purging to the canopy layer atmosphere [93]. The reflection of radiation reduces shortwave absorption, with rural vegetated areas said to reflect the incoming input back to the atmosphere by ~20-25% for grass -cover dominant areas and ~15% for tree-cover dominant areas [123]. A significant proportion of the absorbed shortwave energy is utilised by the vegetation for biological photosynthesis, while the residual is held in storage. The effectiveness of this shading effect is determined by leaf size, crown area, and leaf area index of the vegetation canopy [125]. Trees, and to lesser extent shrubs, present higher shading effectiveness in comparison to grass types. A large tree canopy can therefore create a distinct microclimate beneath it to present a localised trunk-space cool pockets [92].

Surface roughness

Surface roughness modifications are presented when the diversity and dynamics of vegetation canopies transform background wind flow to vary convective heat exchange. Canopy density and foliage features are again significant here, with grass cover offering a barrier of stagnant air nearer to the ground, while dense forests impede background wind flow to retain warmer insulated air beneath the canopy. Dispersed groves with canopy heterogeneity present much higher surface roughness that generates increased mechanical turbulence and in turn convective heat loss [1,126]. Isolated trees however present the best convective heat loss efficiency as they protrude into the boundary layer to present greater surface area exposure, and increased opportunity for contact with drier air flowing from non-vegetated areas to increase evapotranspiration. The three-dimensional morphology of a plant canopy and its exposure to the background climate are therefore significant determinants of the effective heat flux presented by a plant community [123].

In addition to the above three principal processes, pollution filtering and ground surface water runoff reduction by vegetation indirectly assists the cooling of the climate.

Pollution filtering is achieved by dry deposition, a process where the pollutant molecules or particles impact upon and stick to vegetation surfaces such as canopy leaves [127]. The removal of such pollutants reduces atmospheric scatter and the absorption of shortwave radiation and longwave infrared radiation emitted by urban surfaces, the atmosphere, and the sun. This in turn has influence on the net radiation balance and the rates of atmospheric warming or cooling experienced. The filtering contribution of urban vegetation is substantial, and higher for larger canopy trees than other types of vegetation. A modelling study for example had estimated the mature tree cover of the West Midlands in England to contribute a reduction in background particulate matter (PM_{10} ≤10 microns) concentration levels by as much as 4% [128].

Ground surface water runoff reduction is mainly achieved by the interception of rainfall by vegetation canopies. While at the surface, the root spread, and typical softer landscaping context aid in the reduction of runoff rates to encourage greater absorption. Although small-scale experiments have indeed verified the latter claim (e.g., [127]), recent studies have highlighted the permeability of wider urban surfaces to contribute greater [123].

Background climate influence

The effectiveness of the cooling contributed is determined by the background climate of the considered vegetated area:

Soil and atmospheric moisture content is particularly significant, with precipitation and irrigation providing greater soil water potential for transpiration, while high atmospheric humidity suppresses transpiration as the water potential gradient is reduced [125]. The availability of moisture also characterises the typical vegetation growth that results, with greater availability resulting in denser growth that generates greater surface roughness relative to drier climates [126].

Ambient temperature is a variable that determines the rate of sensible heat released from the vegetated surface. Seasonal sensible heat flux is therefore a minimum in winter, while the maximum is reached during the summer when the vegetation-to-atmosphere temperature gradient is typically higher.

Wind velocity is significant for modifying both moisture and temperature. At greater velocities, the convective heat transfer coefficient is primarily dependent on wind velocity as forced convection dominates heat transfer to aid greater sensible heat loss, irrespective of the temperature gradient. Wind flow is also advantageous in high humidity conditions as it assists to advect away saturated air, with higher velocities reducing the leaf boundary layer to enhance the water potential gradient and resulting latent heat flux [125].

In summary, the background climate variables of moisture content, ambient temperature, and their interaction with wind flow variables together influence the typical vegetation profiles that result for a given area. This in turn defines the availability and effectiveness of the cooling processes discussed above, as well as their distribution, discussed below.

Cooling distribution

The extent of cooling influence provided by green space is significant for understanding the likely public health and comfort benefits of green infrastructure enhancement strategies. A meta-analysis of urban park studies had identified that on average they were 1 K cooler during the day, with evidence of this influence extending to the surrounding areas to varying degrees [92].

> With London as an example, an early study of Kensington Gardens and Hyde Park found its recorded 3 K cooling influence to extend up to 200 m beyond its boundaries [46]. A recent longitudinal study of Kensington Gardens had highlighted a nocturnal cooling range between 20 and 440 m, with 83% of influence evident at 63 m from the boundary (approximately half the range), as well as a mean summer temperature reduction of 1.1 K, and a maximum reduction of 4 K observed on certain nights [91].

The nature of green space cooling influence experienced in surrounding areas can therefore vary significantly, with the variables affecting cooling penetration into the surrounding context (both horizontal and vertical distribution), requiring assessment.

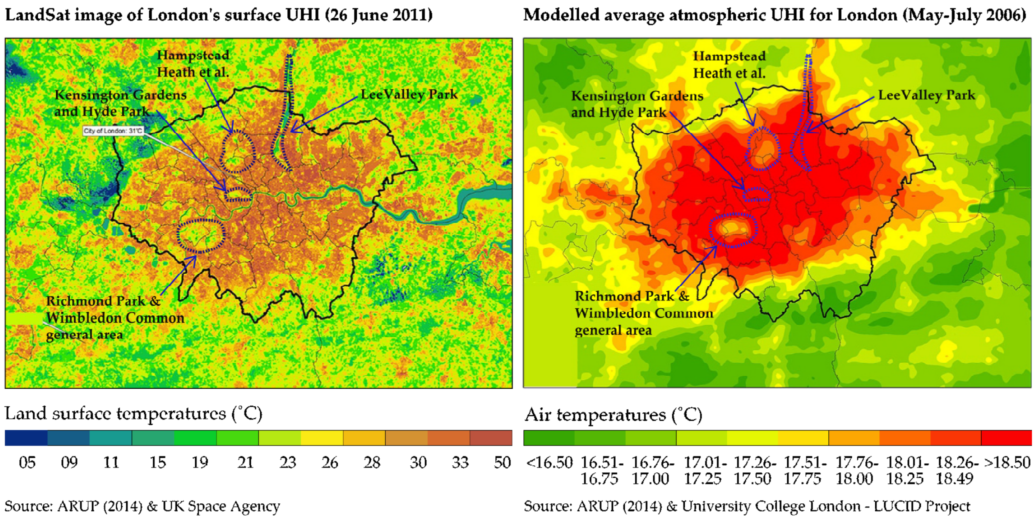


Figure 5. Summertime surface and modelled average atmospheric heat islands for London.

Although numerous studies of urban parks have demonstrated the horizontal distribution of their cooling effects, there is little quantitative evidence presented to clarify how such isolated instances affect the overall climate of a city [92]. From the above surface heat island observations from London (Figure 5), it can be hypothesised that the magnitude and geometrical distribution of green space features to have some degree of influence on citywide (urban boundary layer) cooling. There is however a significant gap in the literature presenting observational vertical distribution data, which makes the relationship between geometric parameters and vertical transport of cooling within the boundary layer, difficult to characterise. This lack of empirical data is attributed to the infrastructural cost necessary to carry out such vertical measurements particularly for longitudinal analysis, which is required to characterise the dynamics of vertical transport. Most studies therefore present and discuss findings in relation to canopy layer distribution and transport, almost exclusively with reference to the horizontal plane.

A higher proportion of the green space cooling effect is typically found to be maintained per metre distance beyond the boundaries of larger features [91], while those smaller than 0.05 km^2 tend to offer negligible distribution [129]. The geometry of the feature is highlighted as significant, with squarer and rounder-shapes found to offer greater cooling efficiency and distribution. This has been explained with reference to temperature and humidity gradients extending from the body to its surrounding landscape, which are typically identified to be relatively greater for such wider proportioned features [97,130]. The range of distribution is also dependent on the vegetation profile and its heterogeneity [124]. A modelling study in response had combined tree age and planting density to present a composite Leaf Area Index (*LAIsp*), as means to calculate the optimum cooling effect relative to park extents [131]. The results of this study supported a previous finding that had identified effective cooling distribution from networks of smaller 0.2-0.3 km^2 green spaces [131,132]. An earlier study that had considered the scale and interval of such features in network or cluster arrangements had identified spacings of <300 m to offer the optimal collective benefit [133]. Although these relationships have not been explicitly assessed in terms of vertical transport, examining Figure 5 in relation to Hampstead Heath and its contextual green spaces, forwards the hypothesis that such clustering arrangements present sufficient vertical transport necessary to affect citywide cooling [115].

The formation and function of wind systems play a significant role in the distribution of cooling contributions from vegetated areas. Macro-to-mesoscale prevailing wind flow and direction over the city affects downwind spread, aided by a combination of simple advection along aligned canyon geometries, and turbulent mixing above building roofs of cross canyons. This in turn establishes built environment morphology metrics such as the 'sky-view factor', 'canyon aspect ratio', and 'canyon orientation', as significant variables that modify cooling distribution [46,93].

The formation of microscale systems has also been identified to play a significant part in horizontal distribution. With low wind velocities typical of anticyclonic conditions, thermals rising from the surrounding urban areas generate low-level advection currents that draw air from parks as 'park breezes' [93,134]. This park breeze effect can generate an eddy system that completes its cycle with the subsidence of warmer urban air from above (see Figure 8). The occurrence of such systems may explain why the cooling rates within urban parks are seldom comparable to rural areas and are more closely associated with their urban surroundings. It may also explain why parks seldom appear on heat island intensity plots (e.g., Figure 5, p. 30), as the occurrence of such eddy systems are likely to hinder the vertical transport of the cooling plume.

Dynamic stability is vital for such conditions to manifest, as higher wind velocities (>5 $m \cdot s^{-1}$) tend to impede vertical movement and disrupt buoyancy-driven effects by introducing rapid turbulent mixing [93]. In a study of London's Kensington Gardens for example, horizontal cooling distribution was notably disrupted with higher wind velocities [91]. Low wind velocities evident under anticyclonic conditions typical of heatwaves and high heat island intensity, in contrast favour the formation of such buoyancy-driven eddy systems [93]. This suggests that the canopy layer cooling influence of green space is assisted by such microscale processes to offer their greatest distribution when it is most likely to be useful in relieving heat stress, which is a significant advantage to bear in mind when comparing against alternative heat mitigation strategies [91].

Urban growth and green space

In contrast to most large cities, London is comparatively greener with ~47% of its area considered as green infrastructure, of which 33% is represented by distinct green spaces, and the residual 14% as domestic gardens [45]. Such urban green space areas however are under constant threat from various economic and spatial demands, particularly in urban centres where intensification of building arrangements is advocated by planning policy. Although the rate of decline has decelerated in recent years, urban green space in the whole of England has reduced by 7% (period between 2001 and 2013), with over two-thirds of this loss attributed to the paving over of domestic gardens, and the rest due to the development of greenfield sites. If further action is not taken, the UK Committee on Climate Change (CCC) has estimated that around 1,000 ha of cover could be lost annually [34].

The discussed environmental capital that green space contributes to the urban setting suggests that it is sensible to conserve what already exists, and where possible to enhance green cover in locations such as city centres and near healthcare facilities, where heat stress relief is critical. It is argued that when opportunities arise to reconfigure urban areas as with regeneration projects, net gain targets should be set and addressed [118].

Tree coverage

Climate studies have demonstrated air temperature differences in various areas of a city to be associated with the tree cover evident, with low cover contributing to higher temperatures [127]. A study of Hong Kong (humid subtropical) had identified tree cover to be more beneficial than grass surface enhancements, with a recommended coverage of a third of the urban area to contribute street level temperature reductions of 1 K (based on Hong Kong's built environment morphology) [135]. Relatively smaller tree planted green spaces have been identified to be of greater benefit than larger grassed areas, as they typically include a wider variety of plant life that offer all the beneficial processes of vegetation related cooling discussed earlier [91]. The air temperature modifications presented by outdoor tree cover also influences building energy use. This influence on the cooling and heating of their indoor environments have been assessed by small-scale experimental studies and larger-scale modelling to emphasise significant energy savings [121,127]. A study from the United States for example had demonstrated tree planting to offer a 25% reduction in net annual building cooling and heating energy consumption [61].

Dispersed development describes the market driven low-density expansion of urban built environments. It is typically criticised for increased land usage in comparison to compaction or densification development strategies, with much of this usage likely to be greenfield land leading to green space and tree coverage loss at the city peripheries [136]. An American study had shown the rate of rural green space loss in the most dispersing urban regions to be more than double the rate in the most compact urban regions, with association made between the frequency of extreme heat events experienced and the loss of regional vegetation cover [137]. The significance of safeguarding peripheral green space was further demonstrated by a study of the Frankfurt greenbelt (maritime temperate), which highlighted the zone to contribute a beneficial cooling effect of between 3-3.5 K.

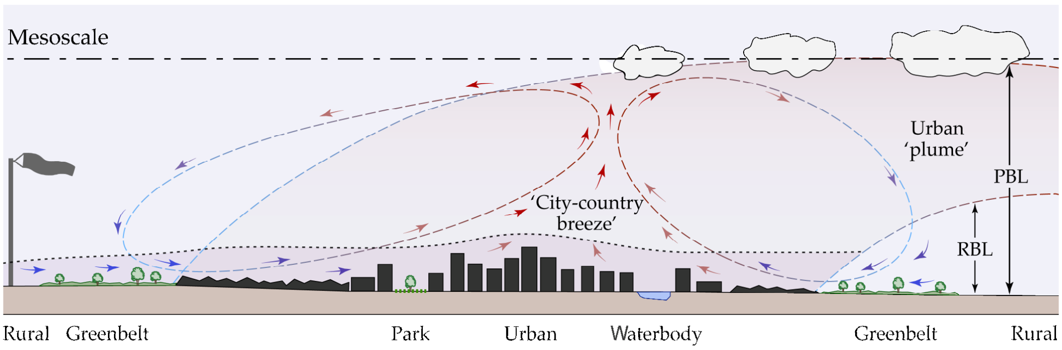


Figure 6. Heat island formation flow, also referred to as the city-country breeze.

The Frankfurt study discussed this cooling influence with reference to the formation of a mesoscale city-country breeze, also referred to as heat island flow [138]. Under anticyclonic conditions typical of high heat island intensity or heatwaves, this citywide system (see Figure 6) develops as thermals at the core of the city rise to the urban boundary layer, that in turn generates advection flow at the canopy layer level from the cooler surroundings of the greenbelt [1]. Urban growth strategies that expand into such peripheral areas can reduce this beneficial breeze by modifying the energy balance at peripheries to reduce the city-country temperature gradient and potential of the system, and by preventing the supply of relatively cooler air that would otherwise been provided by greenbelt vegetation. In contradiction with the typical intentions of urban heat risk mitigation, compact forms of development that encourage higher heat island intensity by concentrating built environment arrangements, favours the formation of this cooling breeze as it enhances the city-country temperature gradient, while dispersed developments weaken its influence (subject to the residual partitioning of the energy balance).

The relative reduction in evapotranspiration between the city and its context had long been considered as the dominant contributor to the boundary layer or daytime heat island [18]. A study of American cities however had demonstrated the daytime heat island to be principally dependent on the relative effectiveness by which urban and rural areas convect sensible heat to the climate, rather than by latent cooling contributions (heat storage remains dominant for the night-time heat island) [126].

> The modelling study identified that if urban areas are aerodynamically smoother than surrounding rural areas, heat dissipation to be relatively less efficient with potential for warming; and if reversed, could potentially lead to a cooling effect. The relative difference in convection efficiency between rural and urban conditions is dependent on the background climate. In humid temperate climates, convection was found to be less efficient at dissipating heat from urban areas than from rural ones, as rural areas tend to be aerodynamically coarser than urban areas due to the presence of generally denser and coarser vegetation canopies. The study highlighted urban built environments in such humid temperate American cities to have a reduced convection efficiency of 58%, contributing to temperature increases of up to 3.0 K that proportionally dominates their daytime heat island intensity. In dryer climates, the opposite occurs, as the built environment is coarser relative to the surrounding landscape, where drier conditions typically impede the growth of denser vegetation types. In such American cities, a 1.5 K decrease in heat island intensity was simulated. In certain cities, this decrease resulted in lower urban temperatures to indicate a daytime heat sink effect [126]. This phenomenon had previously been explained with reference to the oasis effect, resulting from the enhanced evaporative cooling contributed by urban vegetation [73]. The study however stressed that given the proportional contributions to the overall daytime heat island intensity, the evaporative cooling contribution to be less significant than the sensible convection contribution. These findings suggested that the addition of vegetation with the principal aim of improving evapotranspiration qualities of the urban surface, may prove to have less influence on the mitigation of the daytime heat island than previously held. At the boundary layer scale of the urban surface, vegetation seems to provide greater contribution to the cooling of the city by enhancing its surface roughness.

In humid climates where daytime heat island warming is observed to be substantial, the addition of vegetation to increase inner-city surface roughness remains as a viable strategy to mitigate the relative temperature difference [126]. If vegetation is to be used for this purpose, tree planting with increased diversity of species contributes greater roughness than planar greening approaches. The typologies of green infrastructure to prioritise when planning enhancements therefore requires consideration not only in terms of transpiration potential, but also the surface roughness delivered in their varied arrangements and seasonal forms. Certain planning processes have in recent times developed weighting systems to address such variations. For example, the Green Area Ratio (GAR) implemented in Berlin, and also adopted in Malmo, assigns weighting factors based on relative climate change mitigation potential [139]. Such planning instruments however must be continually updated with multidisciplinary evidence to ensure green infrastructure enhancements deliver their optimal heat mitigation potential.

Surface greening

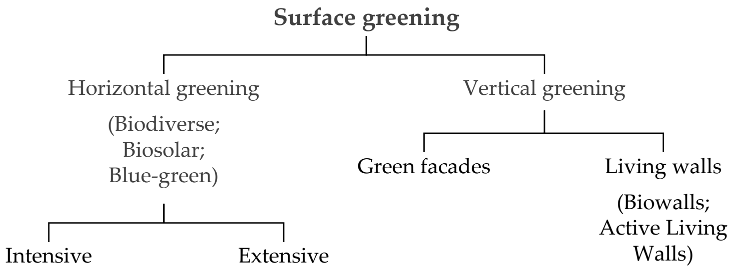


Figure 7. Principal surface greening categories (and variants).

Surface or planar greening solutions refer to the intentional application of vegetation communities to built environment surfaces (horizontal or vertical). A study of the built environment of Hong Kong (humid subtropical) demonstrated that for cities with similar high-density arrangements, planar greening solutions maybe the only available approach for enhancing urban green cover [135]. Such planar greening strategies affect the urban energy balance both directly and indirectly.

Surface greening solutions directly influence the climate by reducing surrounding surface temperatures, which modifies the longwave (L↑) radiation balance to affect microclimate temperatures [140]. A modelling study of Manchester for example, demonstrated the green roofing (i.e., horizontal greening) of its city centre buildings to contribute surface temperature reductions exceeding 6 K [118]. The aforementioned study of Hong Kong however highlighted the approach to be less effective for street level air temperature reductions, particularly when typical building morphology exceeds 10 m in elevation. In Hong Kong where the mean building height is ~60 m, green roofing influence at street level was deemed negligible [135]. Green infrastructure rating schemes such as the one used in Berlin, now acknowledge this elevational decoupling by excluding features constructed on high-rise buildings [139]. A review of green roofing studies similarly concluded proximity to be significant to their cooling contribution and experience, with limited vertical influence stressed. There is however little empirical evidence available to describe the vertical temperature structures and influence decay above such features, given that studies typically disregard monitoring beyond the immediate range [125].

The indirect energy balance influence of green roofing strategies is presented by the modification of heat transfer in and out of indoor environments, which in turn affects their cooling and heating energy demands, as well as the resulting heat rejection from HVAC equipment back to the urban climate [121,140]. This influence has been assessed by previous studies mostly in comparison to the alternative strategy of cool roofing (the application of high albedo surface coatings to roofs). Such comparative studies have demonstrated cool roofing to offer greater annual energy savings than extensive green roofs, although when considered in relation to the intensive type that includes greater vegetation cover and heterogeneity, the converse may be expected. The overall energy saving potential therefore increases with the vegetation leaf area index, and with cooling dominated buildings it is highlighted as the critical parameter [63]. From a purely economic perspective better value is presented by cool roofing, given that green roofing typically has higher construction and maintenance costs [140]. However, the overall value of green roofing approaches must be assessed beyond economic costs alone to also consider the balance of thermal insulation, runoff reduction, carbon uptake, and other ecosystem benefits offered.

Most studies considering vertical greening strategies that apply vegetation communities to building facades, have found their thermal influence to be limited to the immediate zone, with principal cooling contribution attributed to their function as sunscreens [141,142]. With the living wall subcategory that integrates fertigation systems to their vertically supported substrate zones, air temperature influence is reported to be better defined, although the range of influence is limited to <2 m from the installation surface [141]. The coverage area considered by such studies however is limited to either test-rigs or isolated facades. The influence presented by such solutions when deployed in canyon conditions, where surface temperatures are significant for modifying the heat storage potential and subsequent feedback to nocturnal canopy layer temperatures [83,143], is not sufficiently discussed at present. Studies available also tend to focus on monitoring or simulating immediate horizontal canopy layer interactions, rather than vertical transport further up into the atmosphere. In general, significantly less material is available at present on cooling contribution assessments of vertical solutions relative to horizontal greening, partly explained by the limited availability of suitable case studies to examine [115,116].

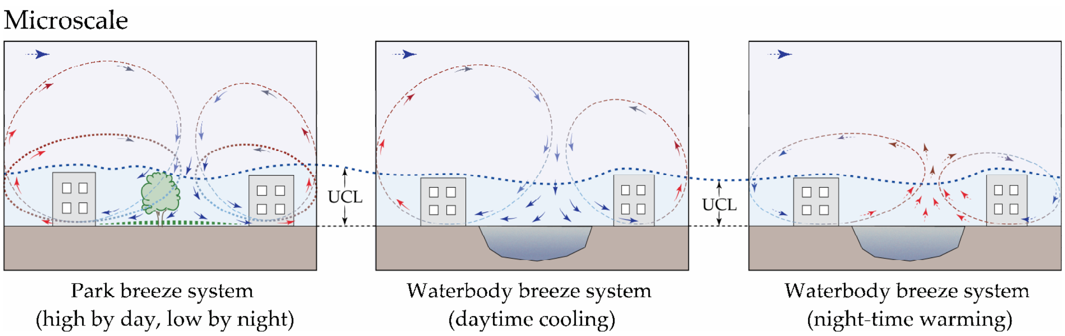


Figure 8. Illustration of green and blue space interactions with the urban climate.

Blue space

Urban blue space or blue bodies refer to surface features that include substantial volumes of water. The historical geopolitical significance of certain cities is predicated on the existence of such a natural feature. The River Thames in the port city of London for example is its dominant and central feature, and together with other blue space bodies such as tributaries, canals, and reservoirs, collectively account for circa 2.5% of the city's surface area [45]. The heat-risk mitigation contribution of the Thames was highlighted by a study that found the air temperature at its river banks to be 0.6 K cooler on average than in neighbouring streets [144]. In agreement with this finding, a meta-analysis of 27 studies (including remote-sensing studies) had identified blue space in general to provide an average cooling effect of 2.5 K, relative to their surroundings [95]. Urban planners and architects have long been conscious of such waterbody contributions in minimising urban heat stress, although the extent and dynamics of these contributions have seldom been discussed [145].

Properties of water

The ability of a waterbody to modify surrounding temperatures is determined by its inherent properties and their interaction with the background climate. The thermal properties of high specific heat capacity and enthalpy of vaporisation gives water a high thermal inertia, which plays a significant role in its ability to moderate temperature variations to act as a thermal buffer. Surface reflectance (albedo) of water varies, with low solar angles presenting high values, and the dominant medium-to-high solar angles presenting low values (α ~0.09). This means that for most of the day, water is an effective absorbing surface [1,71]. The amount of solar radiation penetration and absorption however varies with flow rate and dynamics (i.e., waviness), biochemical make-up, and the quantity of suspended particles present (i.e., turbidity). With most waterbodies, shortwave radiation penetration is unlikely to reach water column depths beyond 10 m. In terms of longwave infrared radiation, the incoming flux ($L\downarrow$) from the atmosphere is almost entirely absorbed at the surface with hardly any reflection, while the outgoing flux ($L\uparrow$) remains constant throughout the day (with larger bodies), owing to the limited range in diurnal surface water temperatures observed. The fluid properties of water enable the absorbed radiation to be transferred within the waterbody by conduction, radiation, convection, and advection processes, that in turn contributes to efficient heat transport and mixing. This permits heat gains or losses to be efficiently diffused throughout a large surface volume, and maintain surface water temperatures within a limited diurnal range [1].

When water absorbs thermal energy from ambient conditions and converts sensible heating to latent heating, evaporative cooling is achieved. For larger waterbodies such as oceans, more than 90% of the radiation balance available on an annual basis may be used to evaporate water, while for smaller bodies this conversion is likely to be >50%. This translates to lower Bowen ratios on average for such surface waterbodies. The evaporative flux also demonstrates diurnal variation. For a large part of the morning, the absorbed energy is mostly used to warm the water (i.e., increase energy storage). Towards the afternoon when the water surface temperature and the water-to-air vapour pressure deficit reaches their peak, a strong evaporative demand is generated. The energy stored within the waterbody is adequate to sustain this evaporative flux even throughout the night, although with diminishing intensity [1]. The sensible cooling effectiveness of a waterbody is dependent on the net effect of the radiation balance, and the climate variables that encourage the sensible-to-latent heat (evaporative) conversion [146].

Background climate influence

Higher waterbody temperatures translate to greater amounts of energy within the fluid system. This leads to higher surface temperatures that enhances the water-to-atmosphere temperature gradient. The presence of relatively cold air above the waterbody enhances the sensible flux as the water-to-atmosphere temperature gradient is enhanced, while the presence of warmer air has the opposing effect. The water-to-atmosphere moisture gradient or vapour pressure deficit (VPD), affects the potential for moisture to transfer into the atmosphere. Relatively drier air above the waterbody enhances the evaporation rate as the water-to-atmosphere moisture gradient (or VPD) is increased, while the presence of humid air has the opposing effect. Increased wind velocity above the waterbody can significantly alter both the sensible and evaporative heat flux by advecting away heat and moisture to enhance temperature and humidity gradients [146].

Types of waterbodies

The sensible cooling effectiveness of dynamic (open system) and static (closed system) types of blue space differ owing to their respective fluid flow characteristics.

Dynamic waterbodies

The thermal properties of dynamic bodies such as rivers, streams, and canals are influenced by fluid flow variables and climate parameters. Their fluid flow enables them to carry by advection absorbed radiation downstream (subject to flow-dynamics), and release energy in locations beyond the urban system [146]. Observations of such bodies have identified daily water temperatures to increase downstream, and when traversing through urbanised areas to demonstrate marked increases. A study by Galli [147] for example, observed stream temperatures in Washington DC (humid subtropical) to increase with impervious surface cover, a metric used to characterise and classify urbanisation. In a study of Long Island, New York (humid subtropical), urban streams were found to be 5-8 K warmer in the summer and 1.5-3 K cooler in winter than rural streams. This study also observed diurnal temperature fluctuations to be greater in urban streams, with notable contribution from summertime stormwater runoff from heated impervious surfaces leading to 10-15 K warmer temperatures than rural streams [148]. Although this form of storm runoff has a beneficial cooling influence on upstream urban surfaces, the process may lead to thermal pollution and resulting biochemical concerns further downstream [149].

The energy balance of a river is typically dominated by the net shortwave (K*) radiation balance, and followed by contributions from the net longwave (L*) radiation balance and evaporative flux [150,151]. A study of River Exe in Devon (maritime temperate) for example, demonstrated its net radiation balance to account for 56% of the heat gain, as well as 49% of its heat loss [152]. A study of twenty streams in Washington State however, observed riparian vegetation to gain increased significance in reducing the net radiation balance by reducing exposure to solar radiation incidence [153]. As most such bodies are of limited depth for most of their course, thermal exchanges at the riverbed-to-water interface may also require attention, particularly during seasonal transition periods in spring and autumn. A study of River Blithe in Staffordshire (maritime temperate) for example, found 82% of its energy exchange to have occurred at the atmosphere-to-water interface, while ~15% had occurred at the riverbed-to-water interface [150]. In smaller streams, the influence of other energy balance partitions may gain greater proportional significance, which also applies to hydrological factors such as discharge and groundwater exchange. For larger dynamic bodies however, high exposure to solar radiation and wind flow are likely to influence dominant heat exchange occurring at the atmosphere-to-water interface. This means that their water temperatures are principally modified by above-body climate conditions and their diurnal and seasonal cycles, with the possible exception of when substantial discharge or absorption from an external source is relevant [154].

Static waterbodies

As dynamic instability is limited in static bodies owing to their restricted flow, they tend to demonstrate higher sensitivity to energy exchange modifications at the atmosphere-to-water interface. Local climate conditions and their interactions with the water balance is therefore significant in defining the thermal properties of such aquatic systems. In deep static bodies the likes of lakes and reservoirs, thermal inputs contribute to temperature and density changes that result in strong thermal stratification of the water column. In such bodies, the 'epilimnion' describes the warmer upper layer; followed by the 'metalimnion' or 'thermocline', where the temperature begins to decrease rapidly with depth; and with the cooler, denser, and stable 'hypolimnion' at the bottom. The thermally active zone is represented by the epilimnion and the higher regions of the thermocline, and is significant for the thermal exchanges with the climate above the body. Several mixing mechanisms affect the thermal moderation of this active zone.

At the very surface of a body, evaporative mixing is evident as the latent flux generates instability that brings warmer water to the surface to maintain a relatively constant surface temperature [1]. Far greater significance to surface mixing is the mechanical energy transferred by windshear stress at the water surface that produces fields of waves and turbulence. The strength of this mixing is influenced by prevailing wind flow conditions at the waterbody and the factors that determine its exposure to this flow, such as fetch (distance measured in the upwind direction that generates surface property driven effects), and the presence of littoral (nearshore) obstructions. Strong wind driven turbulence may in certain instances transfer turbulent kinetic energy to the lower layers to destabilise stratification [155,156]. In addition to these forms of surface mixing, a diurnal mixing current may be generated at the littoral zone of a body. As the shallower littoral slope heats faster than the open water during the day, a horizontal current is generated from the zone towards the open water, while cooler water from the open water depth is drawn up the slope. As the littoral zone cools faster than open waters at night-time, the current is reversed. This diurnal littoral zone current however is typically not potent enough to destabilise stratification of the entire body, and thus is mostly significant for biochemical processes [157].

Whole-body mixing is evident in holomictic waterbodies (uniform temperature and density from surface to bottom), common to temperate climates. Seasonal changes in such climates produce the conditions for buoyancy-driven overturning of the stratification structure. The threshold temperature for this overturning is ~4°C, when pure water reaches maximum density. In spring, surface water that is cooler than this starts to warm, increase in density, and drops to generate convective instability. This overturning occurs until the epilimnion reaches ~4°C, after which warming increases stability and restricts vertical mixing to regain stratification. In the autumn as the surface water cools, its density increases to generate convective instability and overturning [1]. Waterbodies that demonstrate this manner of biannual seasonal mixing are referred to as 'dimictic' bodies, with cold winter climate bodies particularly observed to demonstrate relatively rapid overturning.

Whole-body mixing is also generated from anthropogenic interventions, usually employed to maintain the necessary balance of biochemical processes in certain bodies, such as reservoirs. Save for conditions with such manmade mixing regimes and seasonal overturning, all other natural mixing processes discussed above usually maintain the stratification structure for most of the year. During such periods, the surface layer volume remains dominant for thermal moderation and feedback to the climate above, as the stratified state restricts the use of the entire thermal capacity of the water column. Greater thermal capacity depth is only likely to become available towards the end of the summer, when internal conduction and radiation aids the transport of warming further down the water column to increase the thermally active surface layer volume.

In shallower static waterbodies, the limited whole-body volume presents reduced thermal capacity and inertia. Their peak surface temperature and resulting latent flux can therefore reach relatively earlier than a deeper waterbody [1]. The reduced volume also means that the relative significance of processes such as the conduction of heat across the water-bank boundary into the surrounding surface, littoral zone and wind-driven mixing discussed above, and heat storage from absorption by aquatic flora, fauna, and other matter, are all likely to be more pronounced [1,156]. The increase in proportional significance of these partitions mean that the net radiation balance converted to the evaporative flux will be lower than at larger bodies. Notwithstanding this relative difference, the latent conversion flux is still expected to represent a substantial proportion. For example, a study of a shallow lake in the Hudson Bay lowlands (subarctic), determined that on average 55% of its daily net radiation balance was utilised for the evaporative flux [158]. In addition to this thermal exchange at the atmosphere-to-water interface, the reduced depth of such bodies enables shortwave penetration to typically reach their full water column depth, with the penetration often potent enough to conduct a partition into the subsurface to present water-to-bed interface exchanges, as well as potential for warming from below. Accounting for wind-driven mixing influence, the occurrence of thermal stratification therefore had been discounted for such shallower bodies, with the full depth typically considered as a well-mixed epilimnion. Recent studies however have highlighted that during warm and calm periods (typical of heatwaves and high heat island intensity), shallower bodies to also demonstrate stratification, frequently and for substantial durations [156,159].

Smaller and shallower ponds have in recent times gained increased attention as means to sustainably manage urban drainage requirements, and are typically implemented in larger masterplan developments (referred to as Sustainable Drainage Systems, and abbreviated as SuDS). The urban setting affects the thermal properties of such ponds by the inflow of summertime surface runoff from surrounding developments, influence from anthropogenic discharges into the water balance (i.e., thermal pollution), and the morphology of the contextual built environment inhibiting wind-driven surface mixing. These influences may in turn present the opportunity for thermal stratification to develop. For example, a study of ten shallow urban ponds in Ontario, Canada (subarctic and humid continental), had identified the density changes produced by daytime heating as not often dissipated, with the stability and stratification achieved relatively extended during the mid-summer months, and to demonstrate vertical temperature differences >3-4 K between the top and bottom layers. They attributed this strong and persistent stratification evident to the turbidity of water columns (high levels of suspended sediments common in such ponds), and the reduced wind stress recorded at their surfaces [156]. In terms of geometric parameters to consider, the ratio between the body surface area and perimeter, and maximum depth had been identified as significant. Examining the former had revealed ponds with relatively large area but with simple geometry to experience pronounced stratification, as opposed to the condition in larger lakes where large surface area with longer fetch is typically associated with greater mixing [160]. Maximum depth on the other hand had presented the strongest correlation with stratification, with only bodies <1 m in depth typically identified to be near isothermal (i.e., demonstrating constant temperature) [156]. This means that only very shallow bodies utilise their entire water column thermal capacity for climate energy interactions. Beyond these limnological observations, the research considering urban waterbody temperature structures, particularly in relation to such small shallow artificial waterbodies, remains distinctly limited.

Cooling distribution

As discussed earlier in relation to green space, the cooling distribution of blue space may be similarly argued to be dependent on their scale, geometry, and arrangement. A study by Theeuwes *et al.* [94] for example, confirmed the significance of these parameters with the aid of a mesoscale model of hypothetical waterbodies simulated within an idealised city. Their results highlighted relatively larger-scale bodies to demonstrate greater cooling influence adjacent to boundaries and in downwind areas. The extent of the downwind spread was dependent on wind velocity and fetch length of the body, with the relative cooler air originating from the waterbodies transported to generate plumes several kilometres in length. The study also confirmed a previous remote-sensing finding that had identified several smaller regularly shaped waterbodies to present a smaller temperature effect (particularly during the day), although this to be distributed across a larger area of the city [94,97]. The study however offered little discussion on how the distance from the urban core affected this distribution, which in turn presents another distribution parameter that requires further assessment [161].

The study by Sun and Chen [97] of Beijing (humid continental), had established waterbody geometry to be significant for cooling distribution, with square and round geometries highlighted for providing greater efficiency. As discussed earlier in relation to green space, this is attributed to the increased temperature and humidity gradients that are likely to result between such wider-shaped waterbodies and their surrounding landscape. Furthermore, such regular geometries present consistent fetch distances for wind flow from any direction, which in turn presents greater opportunity for atmospheric advection. The significance of the width of a waterbody had also been assessed by a review of dynamic features in Beijing, which highlighted it as a key parameter affecting the temperature and humidity of their riparian zones. The study also found a significant and stable effect of decreasing temperatures and increasing humidity when the river width exceeded 40 m [162]. This significance of width may in part explain why the Lee Valley Regional Park (of which 22% is covered by reservoirs), does not appear to contribute a cooling benefit to the atmospheric heat island of London (see Figure 5, p. 35). Although it presents substantial surface area (>40 km^2), its width-to-length ratio is notably small.

The urban context is significant for modifying the climate variables that influence a waterbody. The Sun and Chen [97] study for example, noted substantially higher surface temperatures at waterbodies with surrounding building arrangements, attributed principally to their typical surface materiality. This contributes to steeper temperature gradients between the centre of the waterbody and its surrounding context to positively affect cooling distribution. The surrounding built environment morphology can influence this cooling distribution by shading and obstructing wind flow over the waterbody. Shading affects the net radiation balance to reduce the temperature gradient and availability of energy to evaporate water, while obstructing wind flow reduces the opportunity for atmospheric advection and surface mixing from waves. As discussed with green space features, built environment morphology also plays a role in directing or blocking advected cooling distribution from the waterbody, as well as in enhancing turbulent mixing.

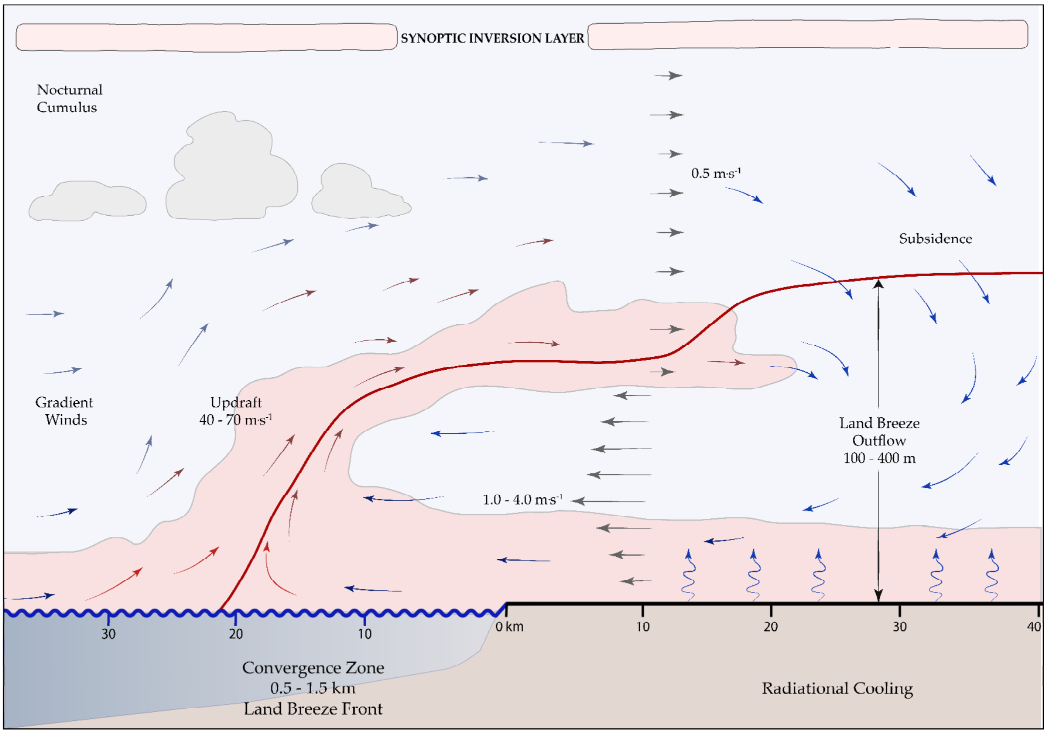
Cooling distribution demonstrates diurnal and seasonal variation. For example, a longitudinal canopy layer study of an urban river in Sheffield (maritime temperate) highlighted its cooling effect to be greatest in the morning, with warm days in May presenting ~2 K cooling over the river and 1.5 K in the riparian zone. At night-time however, no significant cooling was observed, while towards late June even daytime cooling had diminished to be similar to ambient air temperatures [146]. In agreement with these observations, a simulation study of a hypothetical city discussed earlier had identified blue space cooling to be mainly relevant during the daytime, while at night and particularly towards the end of the summer, a warming effect to be likely [94,161]. This diurnal and seasonal variation is explained with reference to the differences in the evaporative flux of such waterbodies. The moderate instability of the atmosphere in the morning hours increases the evaporative flux to the immediate atmosphere above the body to increase its vapour content. The early afternoon period often marks the peak of atmospheric convective instability, water surface temperature, and evaporative flux, although the vapour content of the immediate atmosphere above the body is reduced as the increased buoyancy of warmer vapour transports it to higher altitudes where concentrations are diluted and mixed [1]. This convective instability of the atmosphere of the day however is gradually reduced towards the evening and night, as the surface atmosphere cools to gain stability and resistance to vertical transport. The evaporative cooling during the evening period therefore increases the atmospheric saturation of the stabilising air mass above the body, and the ability to transport vapour to higher altitudes is reduced in relation to the rate at which it continues to be added from the waterbody. The generated vapour consequently converges into these immediate stable layers above the body to create a humidity maximum, and a reduced moisture gradient and resulting evaporative flux. In summary, the diurnal evaporative flux profile is typically expected to peak by day and continue throughout the night at a reduced rate [1]. This in turn is reflected in the diurnal profile of the cooling provided by a waterbody, with any reduction in the evaporative flux below the net longwave flux during the night resulting in a warming effect. The differential cooling of the waterbody in relation to surrounding urban surfaces (that cool faster), also reduces the waterbody-to-context temperature gradient. This in turn reduces the potential for night-time horizontal distribution by advection currents and prevents the removal of the saturated stable air mass above the body to further impede the evaporative flux and enhance the possibility of warming. The warming occurrence is particularly pronounced when waterbodies reach higher temperatures towards the end of the summer from stored thermal energy. The Theeuwes *et al.* [94] study also demonstrated that when the diurnal cycle of water temperature is accounted, this variation (although within a limited range), could result in a reduced cooling duration in the evening and a greater warming duration during the night.

The differences in thermal properties between waterbodies and their surrounding context can also generate breeze systems. These are discussed in previous studies in relation to sea, lake, and land breeze fronts (see Figure 9, p. 43). For example, an observational study of Tokyo found the sea breeze propagation into the coastal region to be slower in urban areas than rural areas [163], with a later simulation study having attributed this deceleration to convergence with heat island flow [104]. Studies examining lake-land breeze fronts have typically considered very large bodies, simply due to the pronounced fronts and effects observed. Lake Michigan in the United States for example (circa 60,000 km^2), has been the subject of several historical studies (e.g., [164–166]). A notable study of it had identified a strong correlation between the deceleration of the front's inland propagation and the maximum night-time heat island magnitude observed in the neighbouring city of Chicago. This highlighted the altitude of the frontal propagation of the land breeze (100-400 m), to correspond to the range permitting interference and convergence with canopy layer flow from the night-time heat island, while no significant association was observed with daytime heat island flow [167]. This in turn clarified sea breeze inflow to present minimal opportunity to disrupt the daytime heat island, where the thermal plume mostly occurs at a much higher altitude at the top of the boundary layer. There is therefore a notable difference in altitude between the two breeze systems, with the lake breeze able to take advantage of stronger thermals developed over land to achieve greater vertical distribution (100-1000 m in elevation), while the relatively weaker night-time thermals developed over the water-body lead to a relatively contracted distribution (100-400 m). To understand these distributions and their interaction with heat island thermal inversion plumes, detailed study of vertical temperature structures is necessary. The Keen and Lyons [165] study of Lake Michigan offered some data gathered from aircraft traverses, while Ryznar and Touma [166] had provided better representation using neighbouring towers. The latter study nevertheless cautioned its findings given that the towers were not over the lake itself, which in ideal circumstances is what is required to best characterise such breeze systems and their vertical temperature distributions. Save for these studies dating between 70s and 90s of larger sea and lake relationships, interest in the study of atmospheric feedback from waterbodies seems to have diminished, with little to no availability of data concerning the vertical transport and temperature structures above smaller-scaled urban features.

The possible occurrence of a microscale eddy system similar to park breezes discussed earlier in relation to green space, is not addressed in the available blue space literature. The formation of such a waterbody breeze system would in theory differ from a park breeze system, given the thermal inertia of water and its day-to-night cooling cycle that contributes to typically warmer night-time temperatures relative to the surrounding urban landscape. The system could thus be expected to reverse during the night as warm saturated air rises from the warmer waterbody, which in turn causes cooler air from the urban surroundings to advect towards this body (see Figure 8, p. 35). This in principle describes a smaller-scaled representation of the above discussed land breeze system (Figure 9, p. 43). The completion of this hypothetical waterbody eddy cycle is the night-time subsidence of warmer and humid air back to the surrounding context (Figure 8), in the same manner of a land breeze system. This in turn presents the possibility for the horizontal transport of an undesirable warming influence into the surrounding canopy layer areas.

The canopy layer trapping of heat could present significant threat to not only thermal comfort, but also human health as night-time temperatures have been epidemiologically established to be the most oppressive [30]. Even during the daytime, the expected waterbody cooling benefit from evaporative cooling dominance, may present a counterproductive net influence. A significant drawback of evaporative cooling is that it increases atmospheric water vapour (humidity), which is also a greenhouse gas [1]. The Theeuwes *et al.* [94] study revealed that in some instances ~60% of the comfort achieved by the sensible cooling effect of blue space, could be negated by this humidity uplift. Considering these diurnal thermal exchange dynamics of blue spaces, they may be regarded to offer a warming influence to urban environments when it is least desired (at night, towards the end of the summer, and during conditions typical of high heat island intensity and heatwaves), and as such offer limited contribution to urban heat risk mitigation demands when considered in isolation.

Land breeze system



Lake breeze system

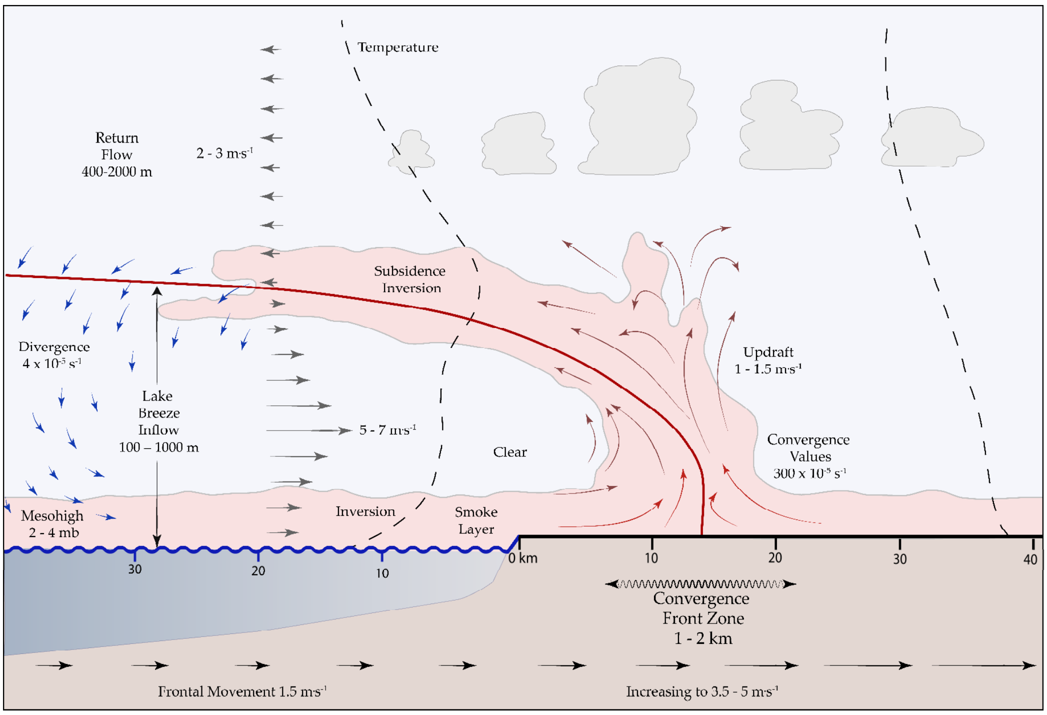


Figure 9. Ideal lake and land breeze systems; based on Keen and Lyons [165].

Synergistic cooling

Although both green and blue spaces are often commended for contributing significant environmental capital [95,146], comparative assessments are uncommon. A notable example was presented by a study of six parks and three lakes in Chongqing, China (humid subtropical), where the cooling in the parks was found to be more defined than at lakes, with the maximum recorded at 3.6 K for the parks and 2.9 K for the lakes [168]. The study however considered this comparison in isolation, with little discussion on the integrated dynamics between the two features. The Xu *et al.* [169] study in contrast considered these synergistic dynamics based on case study specific observations in Shanghai (humid subtropical), and presented a regression model to extend the observed 10-20 m zone of thermal comfort improvement with the use of littoral vegetation. Synergistic influence discussed in other studies remain limited to recommendations based on acknowledged principles, or as explanations for identified anomalous cooling enhancements. The Hathway and Sharples [146] study for example, observed the highest cooling distribution at ~30 m from the river centre to be evident at street canyons that were opened-up to provide access to riparian areas with greenery. Beyond such observations, there is little analysis offered to describe the synergistic processes involved, particularly for conditions where both features are integrated by design as blue-green ecosystems.

Synergistic processes are discussed mostly in potamological and limnological research, where attention is typically given to biochemical than thermal implications of green and blue space interactions. Most potamological studies target the assessment of agricultural and forested conditions. For example, a study from Washington State had considered twenty streams to identify the significance of riparian vegetation in reducing their net radiation balance, by reducing exposure to both solar radiation incidence and wind flow. They found clearcutting of this vegetation to have increased air temperature above the streams by up to 4 K within one winter, with the preserved vegetated buffers having provided some protection against mid-summer air temperature increases, while enhanced protection had been observed both early and late in the summer [153]. Following the removal of riparian vegetation, vapour density at a stream is also likely to increase as higher air temperatures lead to greater evaporation from the stream, and increased transpiration rates from remaining riparian vegetation. This increase in atmospheric moisture and the corresponding reduction in the soil, affects the hydrological balance [170]. In urbanised areas the balance is further affected by impervious surface cover altering the pathways of water movement, which can contribute to hydrological drought [171].

At static bodies, various types of aquatic vegetation (macrophyte) are common in the littoral zone. These range from terrestrial plants, emergent plants (e.g., reeds), free-floating leaved plants (e.g., waterlilies), and submerged plants (e.g., milfoil) [155]. In addition to their numerous biochemical contributions, they are significant for moderating the thermal properties of the waterbodies they inhabit. Littoral vegetation shading, besides modifying the radiation balance of a body as discussed earlier, can cause differential heating and cooling to result in internal convection flows that aid mixing. Where vegetation density and the leaf area index (LAI) is extensive, this shading could be expected to reduce water temperatures in the littoral zone [172]. For example, a study of Priest Pot, a small lake in the Lake District National Park (maritime temperate), found the dominant solar radiation influence at this body to be modified by vegetation to alter its light climate within the water, which in turn encouraged greater mixing and reduced stratification. They however stressed this interaction between littoral zone radiation penetration, vegetation shading, and internal mixing to be complex and specific to each body, with lower or higher relative littoral zone temperatures probable.

Littoral vegetation also contributes to the dampening of wind-stress induced mixing. When wind flow encounters the water surface, the shear stresses generated induce fields of waves and turbulence. The resulting turbulent kinetic energy generated at the surface then penetrates the water column to provide mixing [155]. Surface wind-stress is however dampened when it encounters aquatic vegetation, which results in a reduced surface layer depth and the potential strengthening of temperature stratification [172]. Significant to the degree of mixing generated are both the cover of vegetation present and the fetch (i.e., the length of water over which a given flow has contact). Longer the fetch, the greater the opportunity for turbulent mixing. In larger lakes with increased fetch, littoral vegetation serves to dampen surface-generated turbulent kinetic energy from penetrating the water column. The effectiveness of this is dependent on plant separation, with higher density achieving greater dampening. A study of the aforementioned Priest Pot lake found wind mixing to generally dampen owing to its extensive surrounding tree cover, while the nearby Esthwait Water Lake (Figure 10), with its larger open setting and fetch, allowed active turbulence to develop over open water, save for its dampening at the littoral zone by vegetation [155,173]. In contrast, a study of Lake Purrumbete in Victoria, Australia (maritime temperate), identified greater littoral plant spacing to reduce its influence to the extent that it was similar to open water [155].

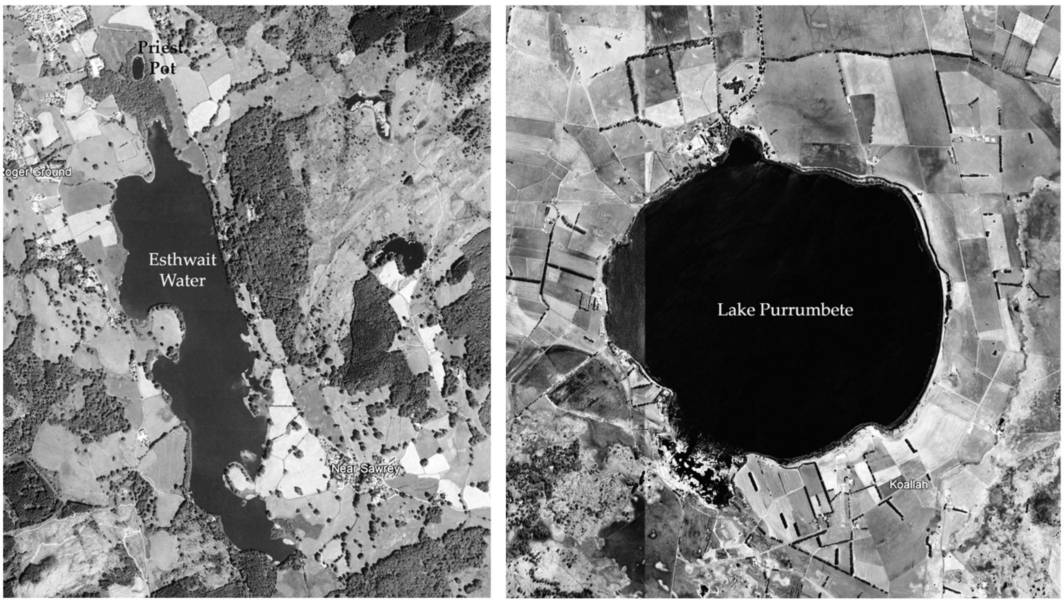


Figure 10. Priest Pot, above Esthwait Water Lake (left); and Lake Purrumbete (right).

The observations from such studies highlight littoral vegetation to have significant influence on the mixing regimes evident at a given waterbody, which in turn has influence on the climate feedback they provide. This latter aspect however is not addressed in potamological and limnological research, which are more concerned with climate effects on such bodies rather than their influence or feedback to the climate above. This in turn highlights an area that requires further attention.

SUMMARY

The following presents a summary of the material discussed in this concise guide.

Heat-related risks

- The heat island effect is the dominant contributor to the unique climate experienced in cities, and is strongly associated with urban growth. As most cities demonstrate conformity with the established urbanisation growth trend, their populations are at heightened risk from its increasing effects, which will be exacerbated further if development and growth plans do not include effective heat island mitigation strategies.
- In addition to continued climate warming, many regions are likely to experience increases in the frequency and severity of extreme heat events. The prediction of increased events is not favourable news for growing cities, as the anticyclonic conditions that favour the formation of such heat events also exacerbate the severity of heat islands.
- Mitigating and adapting to a warming climate and the heat island effect are highlighted as likely to require a range of substantial resilience measures, addressing both citywide and microclimate processes.
- Mitigation and adaptation are of equal priority when considering climate resilience measures. In relation to heat-related risks, this translates to measures that address the mitigation of prolonged heat storage in cities (long-term), as well as the moderation of heat extremes that enable adaptation (short-term).

Urban climate studies

- The heat island is best observed and most potent under synoptic-to-mesoscale anticyclonic conditions, when reduced wind velocities and cloud cover are typical. Its intensity is greatest in the summer, and at night-time. Mitigation and adaptation measures that can deliver their maximum cooling potential under such conditions should be prioritised.
- The distinct day and night-time difference in heat island intensities experienced is associated to the diurnal change in the convective instability of the urban atmosphere. The increased solar radiation received during the day generates stronger thermals that expand the urban boundary layer and contributes to the dominant thermal inversion occurring at a higher altitude above the city surface (i.e., further away from the habitable zone). At night-time the slow release of heat stored in urban form becomes the dominant heat source for the less potent nocturnal thermals that contribute to the dominant thermal inversion occurring at a lower altitude, typically at the top of the canopy layer (i.e., closer to the habitable zone). The mitigation of the lower altitude nocturnal effect must take precedence, as the proximity influence heightens occupant vulnerability to adverse health and comfort effects. Understanding the variation in the inversion altitude is also significant for determining how both energy sources and sinks transport their vertical influences to modify the urban climate.
- Weather patterns significantly modify heat transfer processes between surfaces and the atmosphere. Targeting measures that increase wind flow and velocity (i.e., generate dynamic instability), followed by cloud cover (i.e., affecting the radiation balance), can

disrupt heat island formation. These include measures that enhances mechanical effects on surface flow (i.e., increase surface roughness), and the evaporative flux.

- Built environment morphology that induces mechanical effects on surface flow (i.e., enhances surface roughness), and use of engineered materials, improved drainage, and increased heat output from energy use that contributes thermal effects to the urban climate, are well addressed by preceding observational research and physical modelling exercises. The influence of green and blue space distribution however remains thinly discussed in comparison, and represents an emerging area of interest for researchers.

Green and blue space contribution

- Green and blue space features and their immediate atmospheric environments are mutually dependent. The state of the immediate atmosphere is modified by the latent and sensible flux from surface water and vegetation, while they also respond to changes in the climate in ways that modify this flux output. Although one-way interactions are well discussed in studies, reciprocal dependencies are given less attention.
- For both green and blue space features, the thermal effects are influenced by intrinsic characteristics such as scale, geometry, spread and interval of features, internal structures or stratification, and surface roughness and fetch length; as well as prevailing background climate conditions such as wind flow, morphology and materiality of the context, and feature-to-context temperature and humidity gradients. These characteristics and conditions in turn influence their thermal feedback and its horizontal and vertical transport, with the latter highlighted as not well addressed by previous research.
- Recent studies suggest evapotranspiration from green spaces to contribute only marginally to the mitigation of the higher altitude daytime heat island effect. In lower altitude canopy layer climates however, significant cooling influence is acknowledged, with horizontal transport of this cooling contributing to the moderation of heat extremes. The greater use of such features is therefore a viable approach for addressing urban heat risks, with diverse and intensive planting observed to offer enhanced surface roughness, shading, and evaporative cooling.
- The occurrence of microscale breeze systems modifies horizontal and vertical cooling transport. With contribution from such systems, green space features have been observed to extend horizontal cooling greatest during conditions typical of high heat island intensity and heatwaves, which in turn provides relief when it is most likely to be required. Blue space features in contrast may transport a nocturnal warming effect that worsens towards the end of the summer, when heat island intensity and risk from heat stress is most oppressive (epidemiologically identified). This suggests that when considered in isolation, green space features offer greater benefit to heat risk management than blue space features.
- Notwithstanding the nocturnal warming effect, blue space cooling contributions and transport during the daytime is substantial. With consideration for body depth and mixing regimes, they remain as a viable approach for managing urban heat risks, particularly when employed in conjunction with green space features to provide mutually dependent environmental capital that offers many benefits, including synergistic cooling.
- With both green and blue features, the addition of multiple smaller interventions that take advantage of dominant wind patterns offer greater cooling transport across a larger canopy layer area than with a solitary larger feature. This validates the use of green and blue space as infilling features, even in high-density compaction (regeneration) schemes.

BIBLIOGRAPHY

[1] T.R. Oke, Boundary Layer Climates, 2nd ed., Methuen, London, 1987.

[2] M. Allaby, A Dictionary of Geology and Earth Sciences, Oxford University Press, Oxford, 2013.

[3] J. Monteith, M. Unsworth, Principles of environmental physics, 4th ed., Academic Press, an imprint of Elsevier, Oxford, 2013.

[4] M.Z. Jacobson, Fundamentals of atmospheric modeling, Cambridge University Press, Cambridge, 2005.

[5] A. Stevenson, ed., Oxford Dictionary of English, 3rd ed., Oxford University Press, Oxford, 2010.

[6] C. Park, M. Allaby, A Dictionary of Environment and Conservation, 3rd ed., Oxford University Press, Oxford, 2017.

[7] C. Gorse, D. Johnston, M. Pritchard, A Dictionary of Construction, Surveying and Civil Engineering, 2nd ed., Oxford University Press, Oxford, 2020.

[8] CIBSE, Guide A: Environmental design, 8th ed., Chartered Institution of Building Services Engineers, Norwich, 2018.

[9] CIBSE, J.F. Nicol, TM52 The limits of thermal comfort: avoiding overheating in European buildings, Chartered Institution of Building Services Engineers, London, 2013.

[10] T. Parkinson, R. De Dear, Thermal pleasure in built environments: Physiology of alliesthesia, *Building Research and Information*. 43 (2015) 288–301.

[11] ASHRAE, ASHRAE Handbook 2013 Fundamentals, American Society of Heating, Refrigerating and Air Conditioning Engineers, Atlanta, 2013.

[12] A. Rogers, N. Castree, R. Kitchin, A Dictionary of Human Geography, Oxford University Press, Oxford, 2013.

[13] K.R. Gunawardena, Residential overheating risk in an urban climate [M.Phil Thesis], University of Cambridge, 2015.

[14] J.A. Patz, D. Campbell-Lendrum, T. Holloway, J.A. Foley, Impact of regional climate change on human health, *Nature*. 438 (2005) 310–317.

[15] S. Hajat, S. Vardoulakis, C. Heaviside, B. Eggen, Climate change effects on human health: projections on temperature-related mortality for the UK during the 2020s, 2050s and 2080s, *Journel Epidemiol Community Health*. 0 (2013) 1–8.

[16] H. Akbari, Energy Saving Potentials and Air Quality Benefits of Urban Heat Island Mitigation, *Lawrence Berkeley National Laboratory*. (2005).

[17] J.R. Stedman, The predicted number of air pollution related deaths in the UK during the August 2003 heatwave, *Atmospheric Environment*. 38 (2004) 1087–1090.

[18] H. Taha, Urban climates and heat islands: albedo, evapotranspiration, and anthropogenic heat, *Energy and Buildings*. 25 (1997) 99–103.

[19] K. Steemers, Energy and the city: density, buildings and transport, *Energy and Buildings*. 35 (2003) 3–14.

[20] J. Kuht, A.D. Farmery, Body temperature and its regulation, *Anaesthesia & Intensive Care Medicine*. 15 (2014) 273–278.

[21] S. Hajat, R.S. Kovats, K. Lachowycz, Heat-related and cold-related deaths in England and Wales: who is at risk?, *Occupational and Environmental Medicine*. 64 (2007) 93–100.

[22] R.S. Kovats, S. Hajat, Heat stress and public health: A critical review, *Annual Review of Public Health*. 29 (2008) 41–55.

[23] L.G. Chestnut, W.S. Breffle, J.B. Smith, L.S. Kalkstein, Analysis of differences in hot-weather-related mortality across 44 U.S. metropolitan areas, *Environmental Science & Policy*. 1 (1998) 59–70.

[24] E. Klinenberg, Heat Wave: A Social Autopsy of the Disaster in Chicago, University of Chicago Press, Chicago IL, 2002.

[25] M. Pelling, C. High, Understanding adaptation: What can social capital offer assessments of adaptive capacity?, *Global Environmental Change*. 15 (2005) 308–19.

[26] S. Lindley, J. O'Neill, J. Kandeh, N. Lawson, R. Christian, M. O'Neill, Climate change, justice and vulnerability, Joseph Rowntree Foundation, York, 2011.

[27] I. Preston, N. Banks, K. Hargreaves, A. Kazmierczak, K. Lucas, R. Mayne, C. Downing, R. Street, Climate change and social justice: an evidence review, Joseph Rowntree Foundation, York, 2014.

[28] C.E. Reid, M.S. O'Neill, C.J. Gronlund, S.J. Brines, D.G. Brown, A. V. Diez-Roux, J. Schwartz, Mapping community determinants of heat vulnerability, *Environmental Health Perspectives*. 117 (2009) 1730–1736.

[29] R. Watkins, J. Palmer, M. Kolokotroni, P. Littlefair, The London heat island: Results from summertime monitoring, *Building Services Engineering Research and Technology*. 23 (2002) 97–106.

[30] L.S. Kalkstein, R.E. Davis, Weather and Human Mortality - an Evaluation of Demographic and Interregional Responses in the United-States, *Annals of the Association of American Geographers*. 79 (1989) 44–64.

[31] T.R. Oke, The Energetic Basis of the Urban Heat-Island, *Quarterly Journal of the Royal Meteorological Society*. 108 (1982) 1–24.

[32] PHE, Heatwave Plan for England: Protecting health and reducing harm from severe heat and heatwaves, (2018) 46.

[33] US EPA, Urban Heat Island Basics, in: Reducing Urban Heat Islands: Compendium of Strategies, US Environmental Protection Agency., Washington DC, 2008.

[34] CCC ASC, Managing climate risks to well-being and the economy, Committee on Climate Change, London, 2014.

[35] K.R. Gunawardena, Heat vulnerability: risk to health and wellbeing in the built environment [M.Phil Paper], University of Cambridge, 2015.

[36] C. Koppe, S. Kovats, G. Jendritzky, B. Menne, W.H. Organization, Heat-waves: risks and responses, World Health Organization, Copenhagen, 2004.

[37] M. Benzie, A. Harvey, K. Burningham, N. Hodgson, A. Siddiqi, Vulnerability to heatwaves and drought, Joseph Rowntree Foundation, York, 2011.

[38] R.S. Kovats, H. Johnson, C. Griffith, Mortality in southern England during the 2003 heat wave by place of death, *Health Stat Q*. 29 (2006) 6–8.

[39] V. Abrahamson, J. Wolf, I. Lorenzoni, B. Fenn, S. Kovats, P. Wilkinson, W.N. Adger, R. Raine, Perceptions of heatwave risks to health: interview-based study of older people in London and Norwich, UK, *J Public Health (Oxf)*. 31 (2009) 119–126.

[40] E. Grundy, Ageing and vulnerable elderly people: European perspectives, *Ageing & Society*. 26 (2006) 105–134.

[41] UN, United Nations Publications, World Population Ageing 2013, United Nations publication., New York, 2013.

[42] S. Brown, G. Walker, Understanding heatwave vulnerability in nursing and residential homes, *Building Research & Information*. 36 (2008) 363–72.

[43] ZCH, Impacts of Overheating, The Zero Carbon Hub, London, 2015.

[44] B.G. Armstrong, Z. Chalabi, B. Fenn, S. Hajat, S. Kovats, A. Milojevic, P. Wilkinson, Association of mortality with high temperatures in a temperate climate: England and Wales, *J Epidemiol Community Health*. 65 (2011) 340–345.

[45] ARUP, Reducing urban heat risk: a study on urban heat risk mapping and visualization, London, 2014.

[46] T.J. Chandler, The Climate of London, Hutchinson & Co Ltd, London, 1965.

[47] M. Kolokotroni, Y.P. Zhang, R. Watkins, The London Heat Island and building cooling design, *Solar Energy*. 81 (2007) 102–110.

[48] ACN, ACN Overheating Meeting: Overheating in Cities and Neigbourhoods, Adaptation and Resilience in the Context of Change Network, London, 2011.

[49] CCC ASC, Adapting to climate change in the UK, Measuring progress, Committee on Climate Change (Adaptation Sub-Committee), London, 2011.

[50] CCC ASC, Meeting Carbon Budgets - Progress in reducing the UK's emissions, Committee on Climate Change (Adaptation Sub-Committee), London, 2015.

[51] S. Diamond, C. Das Bhaumic, A. Mylona, G. Virk, Zero Carbon Hub evidence review: assessing overheating risk, London, 2015.

[52] C. Smith, G. Levermore, Designing urban spaces and buildings to improve sustainability and quality of life in a warmer world, *Energy Policy*. 36 (2008) 4558–4562.

[53] GB DCLG, Investigation into overheating in homes: literature review, Department for Communities and Local Government, London, 2012.

[54] M. Santamouris, N. Papanikolaou, I. Livada, I. Koronakis, C. Georgakis, A. Argiriou, D.. Assimakopoulos, On the impact of urban climate on the energy consumption of buildings, *Solar Energy*. 70 (2001) 201–216.

[55] M. Santamouris, On the energy impact of urban heat island and global warming on buildings, *Energy and Buildings*. 82 (2014) 100–113.

[56] G. Henderson, Home air conditioning in Europe – how much energy would we use if we became more like American households ?, *ECEEE 2005 Summer Study*. (2005) 541–550.

[57] H.E. Landsberg, The Urban Climate, Academic Press, New York, 1981.

[58] GB DECC, Department of Energy and Climate Change, Energy Consumption in the UK: Domestic data tables, Department of Energy and Climate Change, London, 2014.

[59] J. Palmer, N. Terry, S. Firth, T. Kane, D. Godoy-Shimizu, P. Pope, Energy use at home: models, labels and unusual appliances, DECC and DEFRA, London, 2014.

[60] B. Boardman, S. Darby, G. Killip, M. Hinnells, C.N. Jardine, J. Palmer, G. Sinden, 40 Percent House, Environmental Change Institute University of Oxford, Oxford, 2005.

[61] H. Akbari, Shade trees reduce building energy use and CO2 emissions from power plants, *Environmental Pollution*. 116 (2002) S119–S126.

[62] A. Pathan, A. Young, T. Oreszczyn, UK Domestic Air Conditioning: A study of occupant use and energy efficiency, *Air Conditioning and the Low Carbon Cooling Challenge*. (2008).

[63] D.J. Sailor, A review of methods for estimating anthropogenic heat and moisture emissions in the urban environment, *International Journal of Climatology*. 31 (2011) 189–199.

[64] F. Salamanca, M. Georgescu, A. Mahalov, M. Moustaoui, M. Wang, Anthropogenic heating of the urban environment due to air conditioning, *Journal of Geophysical Research-Atmospheres*. 119 (2014) 5949–5965.

[65] C. de Munck, G. Pigeon, V. Masson, F. Meunier, P. Bousquet, B. Tréméac, M. Merchat, P. Poeuf, C. Marchadier, How much can air conditioning increase air temperatures for a city like Paris, France?, *International Journal of Climatology*. 33 (2013) 210–227.

[66] L. Howard, The climate of London: deduced from meteorological observations made in the metropolis and at various places around it, Harvey and Darton, J. and A. Arch, Longman, Hatchard, S. Highley [and] R. Hunter, London, 1833.

[67] A. Sundborg, Climatological studies in Uppsala with special regard to the temperature conditions in the urban area, Appelbergs Boktryckeri Aktiebolag, Uppsala, 1951.

[68] A.J. Arnfield, Two decades of urban climate research: A review of turbulence, exchanges of energy and water, and the urban heat island, *International Journal of Climatology*. 23 (2003) 1–26.

[69] R.L. Wilby, Past and projected trends in London ' s urban heat island, *Weather*. 58 (2003) 251–260.

[70] T.R. Oke, The distinction between canopy and boundary-layer urban heat islands, *Atmosphere*. 14 (1976) 268–277.

[71] H. Taha, H. Akbari, A. Rosenfeld, J. Huang, Residential Cooling Loads and the Urban Heat-Island - the Effects of Albedo, *Building and Environment*. 23 (1988) 271–283.

[72] L. Gartland, Heat islands: understanding and mitigating heat in urban areas, Routledge, Oxford, 2008.

[73] S. Peng, S. Piao, P. Ciais, P. Friedlingstein, C. Ottle, F.M. Bréon, H. Nan, L. Zhou, R.B. Myneni, Surface urban heat island across 419 global big cities, *Environmental Science and Technology*. 46 (2012) 696–703.

[74] G. Ferguson, A.D. Woodbury, Subsurface heat flow in an urban environment, *Journal of Geophysical Research-Solid Earth*. 109 (2004) 1–9.

[75] G. Ferguson, A.D. Woodbury, Urban heat island in the subsurface, *Geophysical Research Letters*. 34 (2007) L23713.

[76] K. Menberg, P. Bayer, K. Zosseder, S. Rumohr, P. Blum, Subsurface urban heat islands in German cities, *Science of the Total Environment*. 442 (2013) 123–133.

[77] C.S.B. Grimmond, M. Roth, T.R. Oke, Y.C. Au, M. Best, R. Betts, G. Carmichael, H. Cleugh, W. Dabberdt, R. Emmanuel, Climate and more sustainable cities: climate information for improved planning and management of cities (producers/capabilities perspective), *Procedia Environmental Sciences*. 1 (2010) 247–274.

[78] I.D. Stewart, A systematic review and scientific critique of methodology in modern urban heat island literature, *International Journal of Climatology*. 31 (2011) 200–217.

[79] UN, World Urbanization Prospects: Highlights, United Nations, New York, 2014.

[80] CCC ASC, Progress in preparing for climate change, Committee on Climate Change (Adaptation Sub-Committee), London, 2015.

[81] K. Steemers, M. Ramos, M. Sinou, Urban Diversity, in: K. Steemers, M.A. Steane (Eds.), Environmental Diversity in Architecture, Routledge, Abingdon, 2004: pp. 85–100.

[82] T.R. Oke, Canyon Geometry and the Nocturnal Urban Heat-Island - Comparison of Scale Model and Field Observations, *Journal of Climatology*. 1 (1981) 237-.

[83] T.R. Oke, Street Design and Urban Canopy Layer Climate, *Energy and Buildings*. 11 (1988) 103–113.

[84] N.E. Theeuwes, G.J. Steeneveld, R.J. Ronda, B.G. Heusinkveld, L.W.A. van Hove, A.A.M. Holtslag, Seasonal dependence of the urban heat island on the street canyon aspect ratio, *Quarterly Journal of the Royal Meteorological Society*. 140 (2014) 2197–2210.

[85] M. Aida, Urban Albedo as a Function of the Urban Structure - a Model Experiment .1., *Boundary-Layer Meteorology*. 23 (1982) 405–413.

[86] T.R. Oke, The Urban Energy-Balance, *Progress in Physical Geography*. 12 (1988) 471–508.

[87] K. Steemers, N. Baker, D. Crowther, J. Dubiel, M. Nikolopoulou, Radiation absorption and urban texture, *Building Research & Information*. 26 (2010) 103–112.

[88] T.J. Chandler, City growth and urban climates, *Weather*. 19 (1964) 170–171.

[89] T.R. Oke, City size and the urban heat island, *Atmospheric Environment*. 7 (1973) 769–779.

[90] H. Akbari, S. Menon, A. Rosenfeld, Global cooling: increasing world-wide urban albedos to offset CO2, *Climatic Change*. 94 (2009) 275–286.

[91] K.J. Doick, A. Peace, T.R. Hutchings, The role of one large greenspace in mitigating London's nocturnal urban heat island, *Science of the Total Environment*. 493 (2014) 662–671.

[92] D.E. Bowler, L. Buyung-Ali, T.M. Knight, A.S. Pullin, Urban greening to cool towns and cities: A systematic review of the empirical evidence, *Landscape and Urban Planning*. 97 (2010) 147–155.

[93] T.R. Oke, The Micrometeorology of the Urban Forest, *Philosophical Transactions of the Royal Society of London Series B-Biological Sciences*. 324 (1989) 335–349.

[94] N.E. Theeuwes, A. Solcerova, G.J. Steeneveld, Modeling the influence of open water surfaces on the summertime temperature and thermal comfort in the city, *Journal of Geophysical Research-Atmospheres*. 118 (2013) 8881–8896.

[95] S. Volker, H. Baumeister, T. Classen, C. Hornberg, T. Kistemann, Evidence for the Temperature-Mitigating Capacity of Urban Blue Space - a Health Geographic Perspective, *Erdkunde*. 67 (2013) 355–371.

[96] K. Gunawardena, Methodologies for assessing urban microclimates, in: University of Cambridge, Department of Architecture, PhD Symposium, University of Cambridge, Cambridge, 2018: pp. 1–20.

[97] R. Sun, L. Chen, How can urban water bodies be designed for climate adaptation?, *Landscape and Urban Planning*. 105 (2012) 27–33.

[98] J.A. Voogt, T.R. Oke, J.A. Voogt, T.R. Oke, Complete Urban Surface Temperatures, *Journal of Applied Meteorology*. 36 (1997) 1117–1132.

[99] K. Menberg, P. Blum, A. Schaffitel, P. Bayer, Long-term evolution of anthropogenic heat fluxes into a subsurface urban heat island, *Environmental Science and Technology*. 47 (2013) 9747–9755.

[100] D.O. Lee, Rural atmospheric stability and the intensity of London's heat island, *Weather*. 30 (1975) 102–109.

[101] T. Kershaw, M. Sanderson, D. Coley, M. Eames, Estimation of the urban heat island for UK climate change projections, *Building Services Engineering Research and Technology*. 31 (2010) 251–263.

[102] D.B. Crawley, Estimating the impacts of climate change and urbanization on building performance, *Journal of Building Performance Simulation*. 1 (2007) 91–115.

[103] S.I. Bohnenstengel, S. Evans, P.A. Clark, S.E. Belcher, Simulations of the London urban heat island, *Quarterly Journal of the Royal Meteorological Society*. 137 (2011) 1625–1640.

[104] H. Yoshikado, Vertical Structure of the Sea Breeze Penetrating through a Large Urban Complex, *Journal of Applied Meteorology*. 29 (1990) 878–891.

[105] G. Pigeon, A. Lemonsu, N. Long, J. Barrié, V. Masson, P. Durand, Urban thermodynamic island in a coastal city analysed from an optimized surface network, *Boundary-Layer Meteorology*. 120 (2006) 315–351.

[106] M. Roth, Review of atmospheric turbulence over cities, *Quarterly Journal of the Royal Meteorological Society*. 126 (2000) 941–990.

[107] R. Watkins, J. Palmer, M. Kolokotroni, Increased Temperature and Intensification of the Urban Heat Island: Implications for Human Comfort and Urban Design, *Built Environment*. 33 (2007) 85–96.

[108] K.R. Gunawardena, T. Kershaw, Urban climate influence on building energy use, in: M.P. Burlando, Massimiliano; Canepa, Maria; Magliocco, Adriano; Perini, Katia, Repetto (Ed.), International Conference on Urban Comfort and Environmental Quality URBAN-CEQ, Genoa University Press, Genoa, 2017: pp. 175–184.

[109] K. Gunawardena, T. Kershaw, N. McCullen, The influence of the urban climate on building energy use, in: 8th International Conference on Structural Engineering and Construction Management, ICSECM, Kandy, 2017.

[110] K.R. Gunawardena, N. Mccullen, T. Kershaw, Heat island influence on space-conditioning loads of urban and suburban office buildings, in: Cities and Climate Conference 2017, Potsdam Institute for Climate Impact Research, Potsdam, 2017: pp. 1–13.

[111] K. Gunawardena, The future of cob and strawbale construction in the UK, University of Bath, Bath, 2008.

[112] M. Iamarino, S. Beevers, C.S.B. Grimmond, High-resolution (space, time) anthropogenic heat emissions: London 1970-2025, *International Journal of Climatology*. 32 (2012) 1754–1767.

[113] S.I. Bohnenstengel, I. Hamilton, M. Davies, S.E. Belcher, Impact of anthropogenic heat emissions on London's temperatures, *Quarterly Journal of the Royal Meteorological Society*. 140 (2014) 687–698.

[114] F. Kimura, S. Takahashi, The Effects of Land-Use and Anthropogenic Heating on the Surface-Temperature in the Tokyo Metropolitan-Area - a Numerical Experiment, *Atmospheric Environment Part B-Urban Atmosphere*. 25 (1991) 155–164.

[115] K.R. Gunawardena, M.J. Wells, T. Kershaw, Utilising green and bluespace to mitigate urban heat island intensity, *Science of the Total Environment*. 584–585 (2017) 1040–1055.

[116] K.R. Gunawardena, T. Kershaw, Green and blue-space significance to urban heat island mitigation, in: S. Emmit, K. Adeyeye (Eds.), Integrated Design International Conference (ID@50), University of Bath, Bath, 2016: pp. 1–15.

[117] G.Y. Qiu, H.Y. Li, Q.T. Zhang, W. Chen, X.J. Liang, X.Z. Li, Effects of Evapotranspiration on Mitigation of Urban Temperature by Vegetation and Urban Agriculture, *Journal of Integrative Agriculture*. 12 (2013) 1307–1315.

[118] S.E. Gill, J.F. Handley, A.R. Ennos, S. Pauleit, Adapting cities for climate change: the role of the green infrastructure, *Built Environment*. 33 (2007) 115–133.

[119] R. Emmanuel, A. Loconsole, Green infrastructure as an adaptation approach to tackling urban overheating in the Glasgow Clyde Valley Region, UK, *Landscape and Urban Planning*. 138 (2015) 71–86.

[120] T. Matthews, A.Y. Lo, J.A. Byrne, Reconceptualizing green infrastructure for climate change adaptation: Barriers to adoption and drivers for uptake by spatial planners, *Landscape and Urban Planning*. 138 (2015) 155–163.

[121] Y.J. Huang, H. Akbari, H. Taha, A.H. Rosenfeld, The potential of vegetation in reducing summer cooling loads in residential buildings, *Journal of Climate and Applied Meteorology*. 26 (1987) 1103–1116.

[122] L. Shashua-Bar, M.E. Hoffman, The Green CTTC model for predicting the air temperature in small urban wooded sites, *Building and Environment*. 37 (2002) 1279–1288.

[123] R. Ennos, G. Percival, Quantifying the cooling benefits of urban trees, in: Trees, People and the Built Environment, Forestry Commission, Birmingham, 2011: pp. 113–118.

[124] S.E. Gill, M.A. Rahman, J.F. Handley, A.R. Ennos, Modelling water stress to urban amenity grass in Manchester UK under climate change and its potential impacts in reducing urban cooling, *Urban Forestry & Urban Greening*. 12 (2013) 350–358.

[125] M. Santamouris, Cooling the cities - A review of reflective and green roof mitigation technologies to fight heat island and improve comfort in urban environments, *Solar Energy*. 103 (2014) 682–703.

[126] L. Zhao, X. Lee, R.B. Smith, K. Oleson, Strong contributions of local background climate to urban heat islands, *Nature*. 511 (2014) 216–219.

[127] E.G. Mcpherson, D.J. Nowak, R. a Rowntree, US Department of Agriculture, Chicago's Urban Forest Ecosystem: Results of the Chicago Urban Forest Climate Project, *Urban Ecosystems*. (1994) 201.

[128] A.G. McDonald, W.J. Bealey, D. Fowler, U. Dragosits, U. Skiba, R.I. Smith, R.G. Donovan, H.E. Brett, C.N. Hewitt, E. Nemitz, Quantifying the effect of urban tree planting on concentrations and depositions of PM 10 in two UK conurbations, *Atmospheric Environment*. 41 (2007) 8455–8467.

[129] K. Doick, T. Hutchings, The Forestry Commission, Air Temperature regulation by urban trees and green infrastructure, The Forestry Commission, Farnham, 2013.

[130] J.X. Shi, J.S. Deng, X.M. Wang, C.F. Luo, X.C. Qiu, Thermal effect and adjusting mechanism of rural landscape patterns, *Scientia Silvae Sinicae*. 47 (2011) 7–15.

[131] B. Vidrih, S. Medved, Multiparametric model of urban park cooling island, *Urban Forestry & Urban Greening*. 12 (2013) 220–229.

[132] L. Shashua-Bar, M.E. Hoffman, Vegetation as a climatic component in the design of an urban street - An empirical model for predicting the cooling effect of urban green areas with trees, *Energy and Buildings*. 31 (2000) 221–235.

[133] T. Honjo, T. Takakura, Simulation of thermal effects of urban green areas on their surrounding areas, *Energy and Buildings*. 15 (1990) 443–446.

[134] C. Jansson, P.E. Jansson, D. Gustafsson, Near surface climate in an urban vegetated park and its surroundings, *Theoretical and Applied Climatology*. 89 (2006) 185–193.

[135] E. Ng, L. Chen, Y. Wang, C. Yuan, A study on the cooling effects of greening in a high-density city: An experience from Hong Kong, *Building and Environment*. 47 (2012) 256–271.

[136] M.M.H. Echenique, A.J.A. Hargreaves, G. Mitchell, A. Namdeo, Growing Cities Sustainably. Does Urban Form Really Matter?, *Journal of the American Planning Association*. 78 (2012) 121–137.

[137] B. Stone, J.J. Hess, H. Frumkin, Urban form and extreme heat events: Are sprawling cities more vulnerable to climate change than compact cities?, *Environmental Health Perspectives*. 118 (2010) 1425–1428.

[138] A. Bernatzky, The Contribution of Trees and Green Spaces to a Town Climate, *Energy and Buildings*. 5 (1982) 1–10.

[139] M. Keeley, The Green Area Ratio: an urban site sustainability metric, *Journal of Environmental Planning and Management*. 54 (2011) 937–958.

[140] R. William, A. Goodwell, M. Richardson, P.V. V Le, P. Kumar, A.S. Stillwell, An environmental cost-benefit analysis of alternative green roofing strategies, *Ecological Engineering*. 95 (2016) 1–9.

[141] N.H. Wong, A.Y.K. Tan, Y. Chen, K. Sekar, P.Y. Tan, D. Chan, K. Chiang, N.C. Wong, Thermal evaluation of vertical greenery systems for building walls, *Building and Environment*. 45 (2010) 663–672.

[142] K. Perini, M. Ottele, A.L.A. Fraaij, E.M. Haas, R. Raiteri, Vertical greening systems and the effect on air flow and temperature on the building envelope, *Building and Environment*. 46 (2011) 2287–2294.

[143] B. Bueno, J. Hidalgo, G. Pigeon, L. Norford, V. Masson, Calculation of air temperatures above the urban canopy layer from measurements at a rural operational weather station, *Journal of Applied Meteorology and Climatology*. 52 (2013) 472–483.

[144] H.M. Graves, R. Watkins, P. Westbury, P.J. Littlefair, Construction Research Communications Ltd, Cooling buildings in London: overcoming the heat island, Construction Research Communications Ltd, London, 2001.

[145] A.M. Coutts, N.J. Tapper, J. Beringer, M. Loughnan, M. Demuzere, Watering our cities: the capacity for water sensitive urban design to support urban cooling and improve human thermal comfort in the Australian context, *Progress in Physical Geography*. 37 (2012) 2–28.

[146] E.A. Hathway, S. Sharples, The interaction of rivers and urban form in mitigating the Urban Heat Island effect: A UK case study, *Building and Environment*. 58 (2012) 14–22.

[147] J. Galli, Thermal Impacts Associated with Urbanization and Stormwater Management Best Management Practices, Department of Environmental Programs, Metropolitan Washington Council of Governments, Washington DC, 1990.

[148] E.J. Pluhowski, Urbanization and Its Effect on the Temperature of the Streams on Long Island, New York, US Government Printing Office, 1970.

[149] M.J. Paul, J.L. Meyer, Streams in the urban landscape, *Annual Review of Ecology and Systematics*. (2001) 333–365.

[150] E.C. Evans, G.R. McGregor, G.E. Petts, River energy budgets with special reference to river bed processes, *Hydrological Processes*. 12 (1998) 575–595.

[151] D. Caissie, The thermal regime of rivers: a review The thermal regime of rivers: a review, *Freshwater Biology*. 51 (2015) 1389–1406.

[152] B.W. Webb, Y. Zhang, Spatial and seasonal variability in the components of the river heat budget, *Hydrological Processes*. 11 (1997) 79–101.

[153] J. Dong, J. Chen, K.D. Brosofske, R.J. Naiman, Modelling air temperature gradients across managed small streams in western Washington, *Journal of Environmental Management*. 53 (1998) 309–321.

[154] R. Gu, S. Montgomery, T. Al Austin, Quantifying the effects of stream discharge on summer river temperature, *Hydrological Sciences Journal*. 43 (1998) 885–904.

[155] M.J. Coates, A.M. Folkard, The effects of littoral zone vegetation on turbulent mixing in lakes, *Ecological Modelling*. 220 (2009) 2714–2726.

[156] K. Song, M.A. Xenopoulos, J.M. Buttle, J. Marsalek, N.D. Wagner, F.R. Pick, P.C. Frost, Thermal stratification patterns in urban ponds and their relationships with vertical nutrient gradients, *Journal of Environmental Management*. 127 (2013) 317–323.

[157] H.G. Stefan, G.M. Horsch, J.W. Barko, A model for the estimation of convective exchange in the littoral region of a shallow lake during cooling, *Hydrobiologia*. 174 (1989) 225–234.

[158] R.B. Stewart, W.R. Rouse, A simple method for determining the evaporation from shallow lakes and ponds, *Water Resources Research*. 12 (1976) 623–628.

[159] K.L. Abis, D. Mara, Temperature measurement and stratification in facultative waste stabilisation ponds in the UK climate, *Environmental Monitoring and Assessment*. 114 (2006) 35–47.

[160] A. Mazumder, W.D. Taylor, Thermal structure of lakes varying in size and water clarity, *Limnology and Oceanography*. 39 (1994) 968–976.

[161] G.J. Steeneveld, S. Koopmans, B.G. Heusinkveld, N.E. Theeuwes, Refreshing the role of open water surfaces on mitigating the maximum urban heat island effect, *Landscape and Urban Planning*. 121 (2014) 92–96.

[162] C.Y. Zhu, S.H. Li, P. Ji, B.B. Ren, X.Y. Li, Effects of the different width of urban green belts on the temperature and humidity, *Acta Ecologica Sinica*. 31 (2011) 383–394.

[163] H. Yoshikado, H. Kondo, Inland penetration of the sea breeze over the suburban area of Tokyo, *Boundary-Layer Meteorology*. 48 (1989) 389–407.

[164] W.A. Lyons, The climatology and prediction of the Chicago lake breeze, *Journal of Applied Meteorology*. 11 (1972) 1259–1270.

[165] C.S. Keen, W.A. Lyons, Lake/land breeze circulations on the western shore of Lake Michigan, *Journal of Applied Meteorology*. 17 (1978) 1843–1855.

[166] E. Ryznar, J.S. Touma, Characteristics of true lake breezes along the eastern shore of Lake Michigan, *Atmospheric Environment (1967)*. 15 (1981) 1201–1205.

[167] J.M. Keeler, D.A.R. Kristovich, Observations of urban heat island influence on lake-breeze frontal movement, *Journal of Applied Meteorology and Climatology*. 51 (2012) 702–710.

[168] C. Li, C.W. Yu, Mitigation of urban heat development by cool island effect of green space and water body, in: Proceedings of the 8th International Symposium on Heating, Ventilation and Air Conditioning, Springer, 2014: pp. 551–561.

[169] J. Xu, Q. Wei, X. Huang, X. Zhu, G. Li, Evaluation of human thermal comfort near urban waterbody during summer, *Building and Environment*. 45 (2010) 1072–1080.

[170] K.D. Brosofske, J. Chen, R.J. Naiman, J.F. Franklin, Harvesting effects on microclimatic gradients from small streams to uplands in western Washington, *Ecological Applications*. 7 (1997) 1188–1200.

[171] P.M. Groffman, D.J. Bain, L.E. Band, K.T. Belt, G.S. Brush, J.M. Grove, R. V Pouyat, I.C. Yesilonis, W.C. Zipperer, Down by the riverside: urban riparian ecology, *Frontiers in Ecology and the Environment*. 1 (2003) 315–321.

[172] W.R. Herb, H.G. Stefan, Dynamics of vertical mixing in a shallow lake with submersed macrophytes, *Water Resources Research*. 41 (2005).

[173] A.M. Folkard, A.J. Sherborne, M.J. Coates, Turbulence and stratification in Priest Pot, a productive pond in a sheltered environment, *Limnology*. 8 (2007) 113–120.