The Urban Climate

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Urban Energy Fluxes

4.1 SOLAR RADIATION

Little imagination is needed to gauge the effect of urban pollutants on the incoming solar radiation, especially the particulates, which scatter and absorb the sun's rays. It has long been known that urban areas have less sunshine than their surroundings. This is an urban effect that can not be doubted because sunshine duration is governed by the general weather situation so that differences in small areas are induced by local effects. Only in mountainous regions are similar mesoclimatic effects noted. In industrial cities the loss in sunshine duration can be between 10 and 20 percent. Similar losses are observed in terms of the energy received at the surface below the urban dust shield. Hufty (1970) noted that in Liège, Belgium, on days with high pollution the city lost 55 minutes of sunshine per day compared with surroundings. In London the inner city has 16 percent less sunshine duration, and the suburbs of the city have 5 percent less than the countryside (Chandler, 1965).

The urban reduction in energy received at the ground is greatest at low solar elevations when the relative thickness of the turbid layers is greatest. The losses are less at high solar elevations. As early as 1934, Steinhauser published some representative data for a few Central European cities, as shown in Table 4.1.

The losses are greatest at the low solar elevations, i.e., in the early morning and late afternoon hours. In winter and autumn the frequent low-level inversions contribute to the accumulation of pollutants and hence to the radiation loss. In spring the generally higher wind velocities and, in summer, the greater convection contribute to the dispersal of pollutants and thus the relatively smaller radiation loss.

Similar results were obtained by Terpitz (1965), who compared the solar energy received on a horizontal surface, the so-called global radiation, in the rather clean air at Trier with the polluted air at Cologne in West Germany. Although the two places are 130 km (81 miles) apart they have otherwise very comparable weather conditions. The radiative reduction was expressed in terms of the percentage of the energy that would be expected in a pure atmosphere where only molecular scattering (Rayleigh scatter) takes place. The mean monthly values of reduction are shown in Table 4.2. In this case, too, the cold season is the time when the metropolis receives much less radiation than the small town. In summer the difference nearly vanishes, but in December it reaches 25 percent of theoretically expected radiation in a clean atmosphere.

There are numerous comparable measurements from other localities. The observations show an annual value of 18 percent for Boston that is nearly identical to that for Cologne (Hand, 1949). For Canadian stations the loss is less: 9 percent for Montreal and 7 per-

TABLE 4.1

Percent Loss of Solar Radiation at Urban Sites Compared to Surrounding, Less-Polluted Countryside[a]

Solar elevation (deg)	Equivalent optical air mass (90° = 1)	Season			
		Winter	Spring	Summer	Autumn
10	5.4	36	29	29	34
20	2.9	26	20	21	23
30	2.0	21	15	18	19
40	1.6	—	15	14	16

[a] Adapted from Steinhauser (1934).

TABLE 4.2

Percent Reduction of Radiation on a Horizontal Surface at Trier and Cologne Compared with Alteration by Rayleigh Scattering[a]

Month	Trier	Cologne	Month	Trier	Cologne
I	70	50	VII	70	67
II	67	44	VIII	74	68
III	71	57	IX	65	61
IV	76	65	X	72	66
V	74	69	XI	60	40
VI	70	69	XII	71	46
Year:	70	59			

[a] From Terpitz (1965).

cent for Toronto (East, 1968). Japanese industrialized cities show rather high radiation losses. Nishizawa and Yamashita (1967) cite a range of 12–30 percent loss for Tokyo compared to a neighboring site.

In contrast, for a 19-month interval in St. Louis, Peterson and Stoffel (1980) found considerably less solar radiation depletion for all wavelengths. Their findings are shown in Table 4.3.

In an intensive survey of conditions in the British Isles, Unsworth and Monteith (1972) concluded that on an average, the direct beam of the sun is depleted by 38 percent. However, the diffuse radiation flux from the sky is greatly increased. When compared with diffuse radiation at the cleanest site in Britain in polar air, the average urban conditions show a 235 percent increase. The total radiation flux in the urban areas is only 82 percent of the lowest value found in the rural areas.

TABLE 4.3

Irradiation Depletion in St. Louis[a]

Site	Summer	Winter	Annual
Urban	2	4.5	3
Suburban	1	2	1.5

[a] In percent of rural area for cloudless sky conditions. (After Peterson and Stoffel, 1980.)

The attenuation is not evenly distributed over the solar spectrum. The shorter wavelengths are particularly affected. The ultraviolet suffers far more loss than the infrared. In many industrial cities in winter nearly all energy in wavelengths below λ = 400 nm is completely absorbed. There are not nearly as many spectral observations available as for total radiation, but an apparently representative example is available for Paris (Maurain, 1947) as shown in Table 4.4. The reduction of the ultraviolet by a factor of 10 is the most notable result. In the visible and the infrared the reduction is only 7 and 4 percent, respectively.

Measurements made in Leipzig and surroundings showed that in the urban area the heat portion of the spectrum has a considerably larger share of the radiation that reaches the ground than in the countryside (Bielich, 1933). For low solar elevations (<25°) the spectral distribution showed 6 percent more in the infrared of the total energy than in the nearby rural area. At high solar elevation (>40°) the difference diminished, but the relative share of the long wavelengths was still 3 percent higher in the city than in the country.

In the visible part of the spectrum, often referred to as illumination, the reduction is quite different in various cities. Steinhauser *et al.* (1955) pointed out that there is considerable compensation for the loss in the direct radiation beam by diffuse sky light. Nonetheless, these authors estimate that for Vienna, Austria, the illumination loss in summer is 10 percent and about 18 percent in winter, compared with the countryside.

The formidable loss of ultraviolet radiation in the direct beam has been frequently observed. In Leicester, England, a town of 210,000, winter values in the 300-nm band showed 30 percent less than the

TABLE 4.4

Spectral Partition[a] of Solar Radiation In and Near Paris[b]

Spectral region	Paris center	Outskirts
Ultraviolet	0.3	3.0
Extreme violet	2.5	5.0
Visible	43	40
Infrared	54	52

[a] In percent of total intensity.
[b] After Maurain (1947).

countryside (Department of Scientific and Industrial Research, 1947). In the Los Angeles Basin, measurements on individual days indicated ultraviolet irradiance losses of 50 percent, and over a period of sustained measurements from August to November, 1973, an average reduction of 11 percent at the International Airport and 20 percent at El Monte (Peterson *et al.*, 1978). At the same time total radiation was only attenuated by 6 percent at the airport and 8 percent at El Monte. The passing of pollutant plumes can be distinctly noted in ultraviolet radiation at Riverside. When such a plume from Los Angeles moves to the far west side of the Basin, reductions in ultraviolet irradiance of 25 percent can be observed at midday (Pitts *et al.*, 1968).

In growing urban areas very definite trends in solar radiation are now documented. An interesting case is the Tel-Aviv, Israel, area, which grew from 350,000 inhabitants in 1946 to 1.2 million in 1974. A series of global radiation records for the decade 1964–1973 at Bet-Dagan, 10 km to the southeast, in the suburbs, showed for summer an overall decrease of 3 percent and a 7 percent decrease for cloudless days (Fig. 4.1). This is an indication of formation of photochemical smog (Manes *et al.*, 1975).

There are some welcome indications that these effects are reversible. Pollution abatement is showing some results. An example is the central part of London, where the creation of a smoke-free zone after the air pollution disaster of December, 1952, is showing gratifying changes. Open fireplaces and the use of coal were replaced by central heating and gas. A comparison of the "clean" decade 1958–1967 with the "dirty" era (from 1931 to 1957) showed a no-

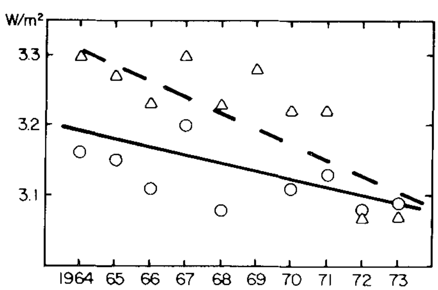

Fig. 4.1 Time series of global radiation at Bet-Dagan, Israel: dashed line (triangles) refers to clear days only; solid line refers to daily averages (adapted from Manes *et al.*, 1975).

table improvement in sunshine duration, especially in the heating season, as shown in Table 4.5 (Jenkins, 1970).

A very useful measure of atmospheric suspensions is the turbidity factor, which can be expressed as

$$T = \frac{\ln I_0 - \ln I - \ln S}{mE_R \ln e} \tag{4.1}$$

where

- I_0 solar radiation extraterrestrially (solar constant)
- I solar radiation at the earth's surface
- S correction factor of the seasonably changing distance of sun–earth
- m optical air mass
- E_R extinction factor for pure dry air (Rayleigh)
- e water vapor

In abbreviated terms one can describe this measure as a ratio of the prevailing attenuation of solar radiation to that produced by a clean atmosphere where only the molecular extinction (Rayleigh scatter and absorption) occurs. A correction for water vapor must be made because of its absorption in the infrared. For this reason an analogous turbidity factor T_K for the shorter wavelengths <625 nm is more useful in urban areas (Dogniaux, 1970).

Again, some interesting data have emerged from Tel-Aviv, where an early period of data from 1930 to 1934 is available. At that time T_K was 1.414; it had increased during 1961–1968 to 1.537 or about an 8 percent urbanization increase in turbidity (Joseph and Manes, 1971). A particularly striking example of the distribution of turbidity over a metropolitan area was published by Steinhauser (1934), Fig. 4.2. He depicted the total turbidity factor in Vienna, Austria, on a typical winter stagnation day. During that season water vapor was not as much of a contributing factor as in summer. A light SE wind

TABLE 4.5

Percentage Increase in Cold-Season Sunshine in London for the Interval 1958–1967 over the Prior Long-Term Average[a]

Month	IX	X	XI	XII	I	II	III
Percent increase	15	26	40	72	55	16	17

[a] After Jenkins (1970).

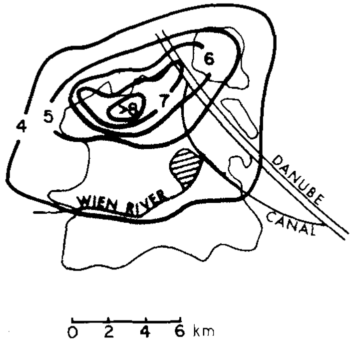

Fig. 4.2 Distribution of the turbidity factor on a stagnation day in Vienna, Austria (after Steinhauser, 1934).

had blown the pollution plume to the NW of the most densely built-up area of the city. The lowest values of T (=3.5) were found upwind at the edge of urban area. The highest value was T = 8.9, a 250 percent increase. In general, values over 4 are considered as indicating notably polluted air.

4.2 OTHER FLUX PARAMETERS

The interaction of solar, atmospheric, and terrestrial radiation at the earth's surface without any complicating anthropogenic factors is itself a very complex phenomenon. Add the man-made changes and it becomes a formidable problem. The basic theory looks deceptively simple:

$$\begin{aligned} \pm Q_N &= Q_I(1 - A) + Q_{L\downarrow} - Q_{L\uparrow} \\ &= \pm Q_S \pm Q_H \pm Q_E + Q_P \end{aligned} \qquad (4.2)$$

[at night the term $Q_I(1 - A)$ vanishes], where

- Q_N net energy balance
- Q_I incoming short-wave radiation received at the surface (both direct and diffuse)
- A albedo of the surface (reflectivity)
- $Q_{L\downarrow}$ long-wave atmospheric radiation downward

$Q_{L\uparrow}$ long-wave radiation upward emitted by the surface
Q_S heat flux into and out of the ground or other surfaces
Q_H sensible heat transfer between atmosphere and ground
Q_E heat loss by evaporation from surface (or plant cover) or gain by condensation (dew or frost formation)
Q_P heat production or heat rejection from man-made sources, including human and animal metabolism

Each one of these factors is different in urban areas from in the countryside and deserves special discussion. Q_I was already reviewed in the previous section. It is the easiest to deal with because it is readily amenable to measurement by pyranometers, and although adequate data sources in urban areas are as yet scarce, the trend toward solar energy use will encourage more of these measurements in the future. Q_N, the net energy gain or loss can also be directly measured, although here, too, the available data are generally restricted to very short intervals of observation.

The value of A, the albedo, has been measured on a number of occasions from aircraft, usually covering both urban and rural areas. Kung *et al.* (1964) first showed the differences that existed in the various environments and the notable seasonal variations induced in higher latitudes by snow cover. These early values indicated 10–30 percent lower values for urban than for rural albedos. More recently, in an air pollution study in the St. Louis area, Dabberdt and Davis (1974) gave the values shown in Table 4.6 for various land uses, obtained during a summertime flight at an altitude of 160 m.

There is about a 4 percent difference between distinctly rural and urban areas. Even though this may seem quite small, it is a significant factor when one deals with the large short-wave energy income from sun and sky. These summer values found close corroboration from another summer study in the St. Louis METROMEX project by White *et al.* (1978), who found albedos of 16.5 percent for the agricultural rural areas and 11.5 percent for the compact residential sectors of the city. Measurements on individual building materials, road surfaces, and parking lots showed albedo values as low as 5 percent. On the other hand, parkland in full leaf may reflect as much as 20 or 25 percent of the incoming short-wave radiation.

Differences in the component Q_L for urban and rural areas have not been exactly determined by observations. In the few cases where Q_L has been separately determined during the daytime simul-

TABLE 4.6

Albedo Values Measured 160 m above the Surface in the St. Louis Area on a Flight of August 9, 1979[a]

Land use	Average albedo (%)
Farmland	14.7
road, some trees	15.4
Woods	
some fields, roads	16.6
fields, some farmland	16.5
Mostly woods, some fields	16.2
New suburban housing tract	16.6
Old urban residential	12.2
commercial	12.1
Commercial, industrial, old housing	13.8

[a] After Dabberdt and Davis (1974).

taneously, the values in these different environments seem to be closely alike. Theoretically, the difference in radiation-absorbing aerosols and gases, such as particles and increased concentrations of CO_2 and other absorbing gases, leads one to expect a somewhat higher $Q_{L\downarrow}$ component in the urban than in the rural area (Ackerman, 1977). Calculations lead to a value of not more than 1–1.5 percent difference in this flux in the two environments. One set of measurements in the Tokyo area (Aida and Yaji, 1979) yielded, however, for nocturnal values an average of 5.7 percent difference, with a range of about 1–10 percent (for 14 measured values). The cause for the increase is the higher temperature of the aerosol layer above the city.

The absorption and scattering of radiation by haze layers had been observed in the London, England, area two decades ago by Roach (1961). Measurements were made from an aircraft in the spectral range between 300 and 3000 nm. There was considerable attenuation of the solar radiation. This included a 5 percent backscatter of the total radiation, 5 percent absorption in the visible, 15 percent in the long waves, and the remainder was a strong forward scatter. This attenuation could lead to heating rates of 5°C/day in the lowest haze layer. Such warming would lead to a different vertical temperature profile over the city than in the country, with an isothermal layer.

Using an empirical model of Nakagawa (1977), Aida and Yaji (1979) calculated the downward flux from aerological observations:

$$Q_{\downarrow} = \sigma T^4[0.127 + (-0.114\Gamma^3 - 0.168\Gamma^2 - 0.173\Gamma + 0.603) \times (0.0000438e^3 - 0.001e^2 + 0.123e + 0.05)^{0.107}] \quad (4.3)$$

where

$Q_{\downarrow}$ downward flux (W m^{-2})
σ Stefan–Boltzmann constant [$5.6696K \times 10^{-8}$ W m^{-2} K^{-4}]
T surface temperature (K)
e surface water vapor pressure (mbar)
Γ vertical temperature gradient (K $mbar^{-1}$)

The calculations gave a 4 percent increase in downward flux over the urban area based on temperature structure alone. These different temperature profiles over the rural and urban areas have been variously simulated, and the role of the polluted layer has been explored (Atwater, 1971; Bergstrom and Viskanta, 1973; Viskanta *et al.*, 1977; Viskanta and Daniel, 1980). All these studies confirm that the pollutant layer plays a major role in the urban energy budget.

The factor $Q_{L\uparrow}$ is more amenable to measurement as a separate flux. A pyrgeometer will directly measure this element, but most of the time the value of this component is determined from the surface temperature

$$Q_{L\uparrow} = \epsilon\sigma T^4 \quad (4.4)$$

where

ϵ emissivity
σ Stefan–Boltzmann constant
T absolute temperature (K) of the surface

There are a large number of surface-temperature measurements available. In earlier years this was determined by thermocouples. Presently these observations are made by infrared thermometry either from aircraft near the surface or from satellites. The first helicopter observations of this element were made by Lorenz (1962). He clearly documented the fact that in daytime with sunshine the surface temperatures of water, forest, and farmland stayed cool, and even small settlements showed higher values. Paved areas, such as runways, were always warmer than the surroundings. In midday an asphalt street was 17.9°C hotter than the air, a hangar roof +17.4°C, a taxistrip +14.1°C, and a small village +3°C. Measurements of Kessler (1971) in Bonn, West Germany, yielded for an asphalt street the maximum temperature 23.5°C above air temperature and minimum temperature +2.6°C. Over a grass surface the corresponding values were: maximum, +9.4°C, minimum, −2.9°C.

It must be emphasized that in clear weather, the surface temperature is nearly always different from the air temperature, measured in a meteorological shelter, both day and night. In overcast, windy conditions the two temperatures may coincide. On a cloudless day in summer the contrasts show up at an early hour. Table 4.7 shows a set of measurements from a survey of the new town of Columbia, Maryland. These data were taken from a helicopter at elevations of 50–100 m above the surface, with an air (shelter) temperature of 29–30°C during the flight, which took place about $1\frac{1}{2}$ hr before solar noon on a summer day (Landsberg, 1969). From the same study of changes brought about by urbanization, two typical cases of day and night conditions under clear skys are shown in Table 4.8.

Other nocturnal observations were reported by Vilkner (1962) who noted a 12°C difference between moist pastures outside the town of Greifswald, East Germany, and the densely built-up center on a clear night. Hence nocturnal differences are of great influence on the vegetation, but all accounts agree on even greater daytime differences. In Columbia, Maryland, a difference of as much as 26°C between the surface temperature on a parking lot and the air temperature was observed. In that growing town it was possible to relate the maximum midday surface temperature differences between the

TABLE 4.7

Infrared Measurements of Surface Temperatures on a Sunny Morning in Columbia, Maryland

Land use	Temperature (°C)
Lake	27.5
Forest	27.5
Farmland	30.8
Parkland	31.0
Open housing areas	32.2
Built-up spaces	34.7
Parking lots and shopping center	36.0

TABLE 4.8

Surface Temperatures and Difference in Air Temperature in the Developing Urban Area of Columbia, Maryland

	Day		Night	
Type of surface	Surface temp. (°C)	Air temp. difference (°C)	Surface temp. (°C)	Air temp. difference (°C)
Lake	26	−1	12	0
Bare soil	35	+4	6	−7
Grass	30	+3	2	−10
Asphalt	41	+14	12	−2

urban and rural areas to the building density, as shown in Fig. 4.3. The extremely high temperatures that can be reached in the afternoon hours of a sunny day have been noted in Vienna, Austria, by Steinhauser *et al.* (1959). For an August day with air temperatures ranging in the city from 22 to 32°C, these authors reported street surface temperatures of 25°C in the shade and up to 51°C in the sun. Metal-roof temperatures of 60°C under such conditions are not uncommon.

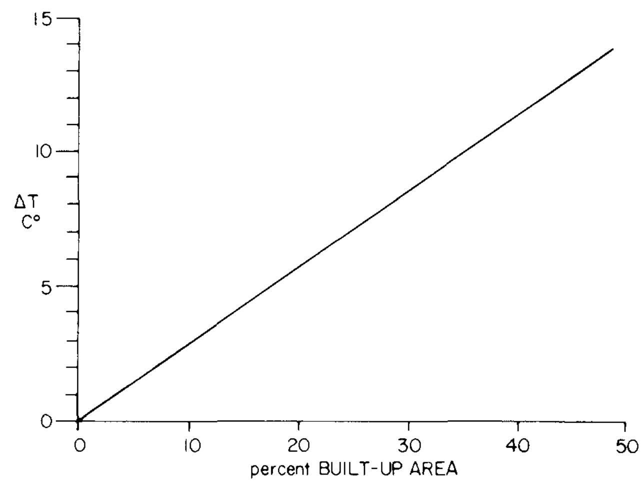

Fig. 4.3 Midday surface temperature excess (urban−rural, °C) on sunny days as a function of the percentage of built-up area.

A great deal of information on surface temperatures can be obtained from infrared scans of satellites. Only in recent times has it been possible to move from qualitative information to quantitative data (Matson and Legeckis, 1980). The resolution is still rather coarse but the pictures obtained are impressive, especially at night (Fig. 4.4). Some analyses have been performed for daytime values. An example is shown in Fig. 4.5 for the Baltimore–Washington corridor.

Such observations enable one to calculate the outgoing radiative energy flux ($Q_{L\uparrow} = \epsilon \sigma T^4$). The emissivity of urban surfaces can be set at 0.96, although this is often overlooked and the factor is

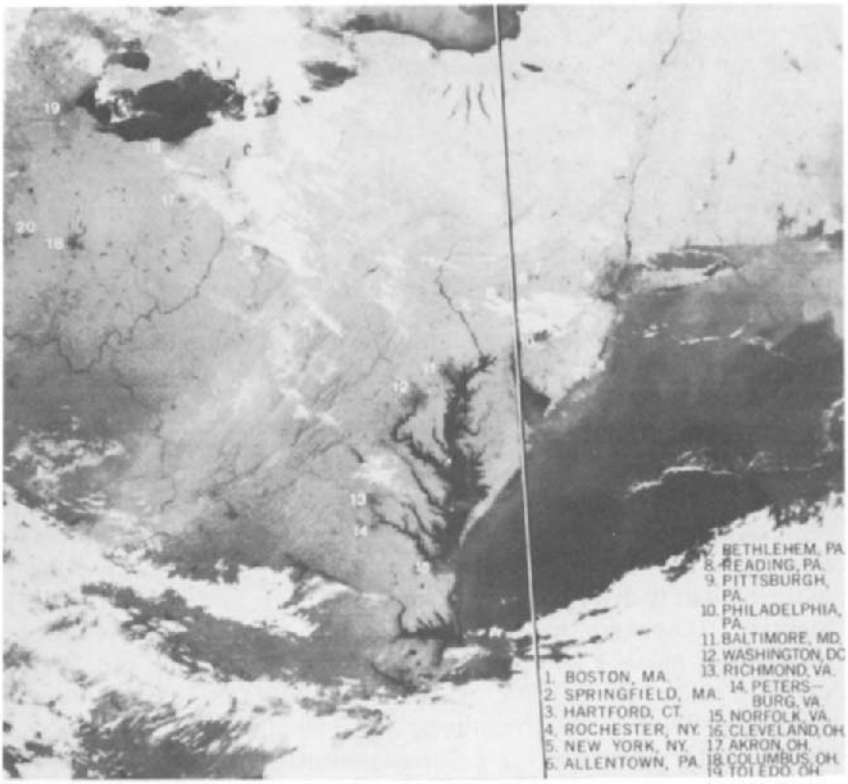

Fig. 4.4 Satellite picture of urban nocturnal heat islands by infrared sensing; dark areas are warm. (Courtesy of National Environmental Satellite Service, National Oceanographic and Atmospheric Administration.)

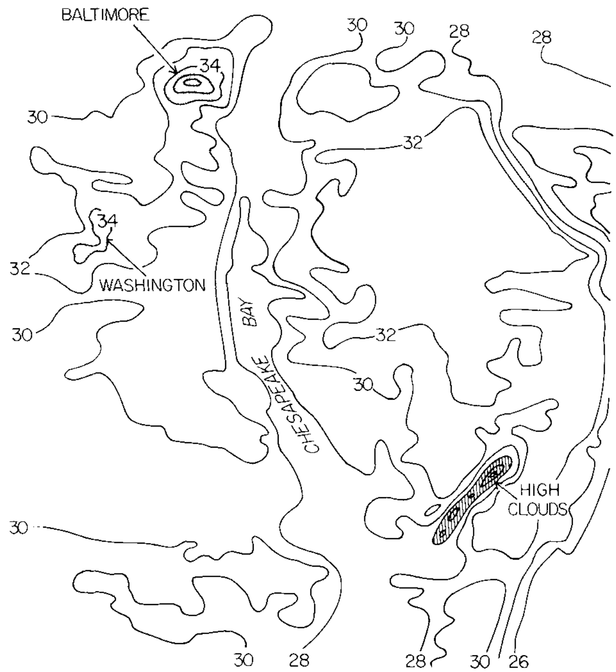

Fig. 4.5 Isotherms of surface temperatures (OC) interpreted from satellite infrared sensor transmissions. (Adapted from National Environmental Satellite Service, National Oceanographic and Atmospheric Administration interpretation.). Note the distinct heat islands of Washington, D. C., and Baltimore.

neglected. In actuality, this makes little difference in view of the fact that both in daytime and at night, as noted above, there is such a multiplicity of surface temperatures. These enter the equation as 4th powers of absolute values. Then a 10° nocturnal difference from, say, 278 to 288 K raises the outgoing energy by 14 percent. Thus it is quite difficult to arrive at a value for $Q_{L\uparrow}$. Fuggle and Oke (1970) remarked: "It appears that the major obstacle to the implementation of the heat balance approach to the urban heat island, and other urban climatic phenomena, is connected with the almost overwhelming diversity and irregularity of urban surfaces. Each one of these surfaces possesses a different albedo, emissivity and heat capacity."

Oke and collaborators have tried to overcome the difficulties by a series of studies investigating the infrared fluxes in urban areas (Oke and Fuggle, 1972; Fuggle and Oke, 1976; Nunez and Oke, 1976; Nunez and Oke, 1977). They introduced the very useful distinction of radiation above the roof level and from what they called the "urban canyon." This term refers to the streets and walls of houses and buildings. These are considered as the essential feature of an urban surface in contrast to the usually flat surfaces of the countryside. From measurements in Montreal these authors obtained as an average of 12 clear nights the values for the radiation balance shown in Table 4.9.

The differences are relatively small. Increased atmospheric long-wave radiation downward over the city is somewhat exceeded by outgoing radiation from the surface. The calculated urban radiative cooling rate above the roof level was determined by Fuggle and Oke (1976) by measuring Q_N and comparing it with actual temperature measurements. The radiative cooling was obtained from the relation

$$\left(\frac{\Delta T}{\Delta t}\right)_{\mathrm{rad}} = \frac{\operatorname{div} Q_N}{\rho c_p} \tag{4.5}$$

where

T air temperature
t time
Q_N net radiation
ρ air density
c_p specific heat at constant pressure

TABLE 4.9

Urban and Rural Long-Wave Radiation Components in the Montreal Area[a,b,c]

	$Q_{L\downarrow}$	$Q_{L\uparrow}$	Q_N
Urban	31.3	−40.1	−8.8
Rural	29.8	−38.2	−8.4
Δ_{u-r}	1.5	1.9	+0.4

[a] Oke and Fuggle (1972).
[b] All values in mW cm^{-2}.
[c] Negative sign indicates heat loss.

The results showed that the calculated radiative rate is always larger than the observed cooling rate. The difference between the two rates was about 3°C hr^{-1}. The explanation is that the radiative cooling is compensated by sensible heat flux convergence.

The complexity of the "urban canyon" energy fluxes are immediately obvious by a glance at the schematic representation of Fig. 4.6 (Nunez and Oke, 1977). A multiple-layer scheme for evaluation was adopted and actual measurements were made in an alley 7.54-m wide, a west wall 5.59-m high, and an east wall 7.31-m high. The energy balance of the *i*th canyon level is given by:

$$Q_{N_i} = Q_{H_i} + Q_{E_i} + Q_{S_i} \tag{4.6}$$

using the same designators as in Eq. (4.2).

The principal conclusions from the experiment are that in daytime Q_H is the principal element carrying heat from the canyon, representing 64 percent of the net radiation. Clearly, wind flow through or across the canyon must be important. Most of the remainder of the energy is stored in the ground and walls. At night Q_H is small because the data were gathered when winds are weak. The radiative flux divergence is large and the net loss is partially compensated for by heat stored in the walls and street during the day. Q_E under the circumstances is small.

In the Columbia, Maryland, experiment (Landsberg, 1973), a complete set of measurements of the elements of the radiation balance on a clear day near equinox, with very light winds (<3 msec^{-1}), was made. The results are shown graphically in Fig. 4.7. In the rural area, in daytime, the evaporative component Q_E is quite large and almost negligible in the urban area. The reflected short-wave radiation because of the high albedo in actively growing vegetation is also

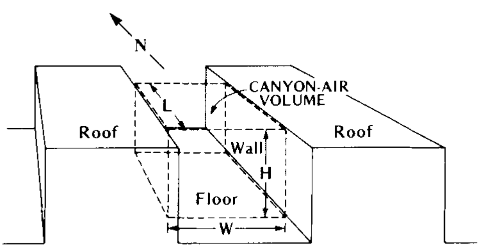

Fig. 4.6 The "urban canyon" in schematic representation (from Nunez and Oke, 1977).

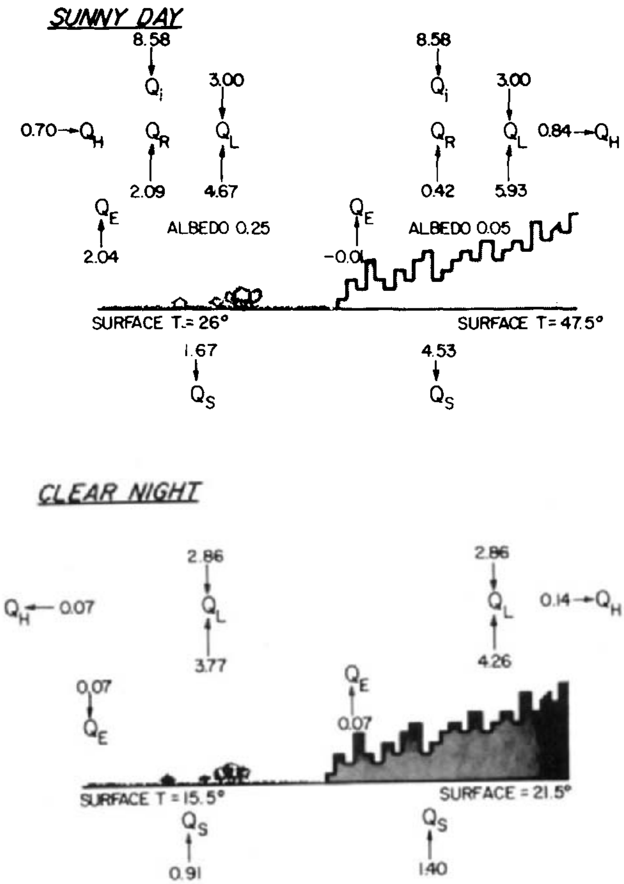

Fig. 4.7 Heat flux data from Columbia, Maryland, experiment. (All values given in W m^{-2}.)

large in the rural area and small in the urban area. The outgoing long-wave radiation is larger in the city than in the rural area, but the greatest difference between them is the stored heat Q_S, which is moderate in the vegetated rural area (19 percent) but to over 50 in the city of the incoming radiation. This is well reflected in the high surface temperature. The nocturnal values are not as startlingly different between the two environments.

Some new information on $Q_{L\downarrow}$ has become available from interpretations of data gathered by an Explorer satellite, launched as Heat

Capacity Mapping Mission (HCMM) by NASA's Goddard Space Flight Center. Surface temperatures were calculated from emissions in the infrared (10.5–12.5 μm). The HCMM satellite flew at 620-km altitudes and had a horizontal resolution of 500 m. Data obtained on a nearly clear summer day (June 6, 1978) at the time close to maximum solar elevation (1300 hr EST) over parts of the heavily urbanized northeastern United States are shown in Table 4.10 (Price, 1979). This table shows the radiative temperature excesses of the urban areas above the surrounding countryside and the calculated radiated power.

There is an obvious, if loose, relation of the excess temperatures to the population and extent of the urban areas. However, we deal with an isolated observation, and for very systematic linkages one would have to resolve the urban fabric into finer segments. In fact, in some of the cities isolated points with additional temperature excesses of 2–3°C were noted, possibly attributable to power plants

TABLE 4.10

Surface Temperature Differences and Excess Radiated Power[a] for Urban Areas in the Early Afternoon in Summer[b,c]

City	Metropolitan population ($\times 10^3$)	Area (km^2)	Temp. diff. (°C)	Radiated power (kW)
New York City	7895	547	17	40,000
Providence, RI	720	48.2	13.2	3190
Hartford, CT	817	26.2	15.0	1770
Schenectady, NY	78	5.3	15.0	368
Bridgeport, CT	157	14.2	12.1	954
Syracuse, NY	197	6.5	10.9	417
Binghampton, NY	64	6.0	12.4	394
New Haven, CT	745	5.8	11.2	372
Worcester, MA	637	5.1	11.5	327
Albany, NY	116	3.7	10.3	227
Stanford, CT	109	3.2	11.2	202
Waterbury, CT	108	1.2	10.9	74
Fitchburg, MA	43	1.2	11.2	131
Troy, NY	62	1.4	10.3	86
Pittsfield, MA	57	0.7	9.7	42

[a] 9°C was chosen as threshold for calculating the excess radiation. Some smaller localities from the original tables were omitted.

[b] Northeastern United States.

[c] From Price (1979).

and industrial agglomerations. Although "ground truth" was not ascertained the figures look entirely reasonable in the light of the measurements discussed earlier.

4.3 MAN-MADE FACTORS IN THE URBAN HEAT BALANCE

Radiation conditions, especially in the canyon streets of the inner cities, are greatly complicated by the change in horizon, which affects the duration of sunshine and illumination. There are also radiative interactions between building fronts in narrow streets and between the buildings and the street surface. This interaction is determined by the height of houses or buildings z_b, the width of the street w_s, and the azimuth of the street ϵ. One can then define an index of street narrowness N as

$$N = \frac{z_b}{w_s} \tag{4.7}$$

This ratio also defines an angle σ of the horizon created by the building line so that

$$\tan \sigma = N$$

The value of N for various values of z_b and w_s are shown in Table 4.11 (Kaempfert, 1949).

The relation between the azimuth angle of the street ϵ, N, and the horizon angle σ is shown in Table 4.12 adopting the convention that

TABLE 4.11

Index of Street Narrowness

	w_s (m)			
z_b (m)	5	10	15	20
5	1	0.5	0.33	0.25
10	2	1	0.67	0.5
15	3	1.5	1	0.75
20	4	2.0	1.33	1

TABLE 4.12

Dependence of Horizon Angle σ on Index of Street Narrowness N and Street Azimuth

	N				
$\epsilon°$	0.2	0.5	1.0	2.0	5.0
0	11.3	26.0	45.0	63.4	78.7
20	10.7	25.2	43.2	62.0	78.0
40	8.7	21.0	37.5	56.8	75.4
60	5.7	14.0	26.6	45.0	68.2
80	2.0	5.0	9.9	19.3	41.0

$\epsilon = 0$ points south and that westerly azimuths are positive and easterly azimuths are negative (only the positive values for σ are given in Table 4.12). This information enables one to superimpose on the theoretical march of the sun above the horizon the restrictions introduced by the building contours. For architectural purposes the horizon can also be constructed empirically by whole-sky cameras (Pleijel, 1954). The artificial horizon will retard sunrise and advance sunset for the street dwellers and thus reduce the available solar radiation and illumination to the inhabitants.

The attenuation of sky illumination and radiation by buildings on the opposite side of the street can be gauged from Table 4.13 where the percentage loss is related to the depth below the roof line in

TABLE 4.13

Illumination Loss Caused by Urban Street Width and Building Height

Ratio of depth below roof/w_s	Illumination loss (%)
0	50
0.25	58
0.5	65
1	75
2	85
3	90
4	92

TABLE 4.14

Air and Surface Temperatures (°C) in a Courtyard on a Sunny Day

			Surface					
					Walls facing			
Weather conditions	Time	Air	Grass	Courtyard	N	E	S	W
3/10 clouds, full sunshine, wind 3 m sec^{-1}	1620	30.6	33	50	32	35	35	50
Sunset, wind 1 m sec^{-1}	1934	28.3	29	33	31	31	32	34
Clear, calm	2115	25.6	23	31	28	28	30	30

terms of the width of the street. It might be noted here in passing that illumination in cities is always less than in the rural environs because of the interception of light by pollutants. Only a few systematic sets of measurements are available. For the steel-manufacturing town of Zaporozhe in the Ukraine, Fedorov (1958) showed that at the highest solar elevations in June illumination in town was reduced by 5 percent and in December when the sun was lowest by 13 percent.

Measurements of infrared temperatures on walls and surfaces in courtyards show the multiplicity of micrometeorological interactions that must be expected in the urban radiative processes. Table 4.14 shows a set of measurements on a bright summer day in a courtyard (32 × 42 m) of buildings 18-m high, with grass surfaces on the outside (Landsberg, 1970). The startling contrasts that remain in a small space, even after the direct radiation ceases, show that intricate internal heat fluxes will rule the energy balances over large portions of a city.

The wall and ground temperatures also govern the energy exchange with the interior of buildings. This is a reciprocal flux that is of great importance for the design of buildings for optimal energy efficiency. Sagara and Horie (1978) have explored this problem in Senri New Town in Japan. The radiative model employed follows the classical radiation equations for wall surfaces. The daytime solar flux and sky radiation affect the walls, as follows (Kondratyev, 1977):

$$\text{Vertical surface facing}\begin{cases}\text{South,} & Q_{I_S} = Q_\perp \sin h \cos \psi \\ \text{East or west,} & Q_{I_{E(W)}} = Q_\perp \cos h \sin \psi \\ \text{North,} & Q_{I_N} = Q_\perp[\sin \delta \cos \phi \\ & \qquad - \cos \delta \cos \phi \cos \Omega]\end{cases} \tag{4.8}$$

where

Q_I direct solar beam energy hitting wall
$Q_\perp$ solar energy received at earth's surface normal to sun's rays
h solar elevation above horizon
δ solar declination
ϕ latitude
ψ solar azimuth
Ω solar hour angle

For the sky radiation we have:

$$Q_{SK} = I_0 \frac{\sin h}{2}\left[\frac{1 - \tau \operatorname{cosec} h}{1 - 1.4 \ln \tau}\right] \tag{4.9}$$

where

Q_{SK} diffuse sky radiation
I_0 solar radiation at boundary of atmosphere
τ transparency

The energy loss from the surface of the walls is, according to Sagara and Horie (1978), governed by

$$Q_{wo} = \sum_{i=1}^{n} \epsilon_w \epsilon_i \sigma\left[\left(\frac{T_w}{100}\right)^4 - \left(\frac{T_i}{100}\right)^4\right]\eta + \epsilon_w \sigma\left[\left(\frac{T_w}{100}\right)^4 - \left(\frac{T_a}{100}\right)(a + b\sqrt{f})\right]\left(1 - k\frac{M}{10}\right)\eta \tag{4.10}$$

where

Q_{wo} outgoing long-wave radiation from the walls
T_w, T_i, T_a absolute temperature of: outer wall, inner wall, and air near wall, respectively
ϵ_w, ϵ_i wall emissivities, outer and inner wall
σ Stefan–Boltzmann constant
M cloud amount

k constant depending on cloud height
a, b constants
η sky aspect factor of wall

Some measured values for summer and winter in the Japanese town are shown in Fig. 4.8. In summer the notable gains of the east wall in the early morning hours are remarkable. The nocturnal losses are not markedly different for the various wall exposures. In winter east and south gain most heat in the morning hours. Nocturnal heat losses are highest for the east wall and lowest for the south wall, but the short measuring interval of only three days permits no categorical statements. The presence of balconies on the building complex produces further complications.

An important quantity in the urban heat balance is obviously the energy used in the city and its ultimate release into the atmosphere.

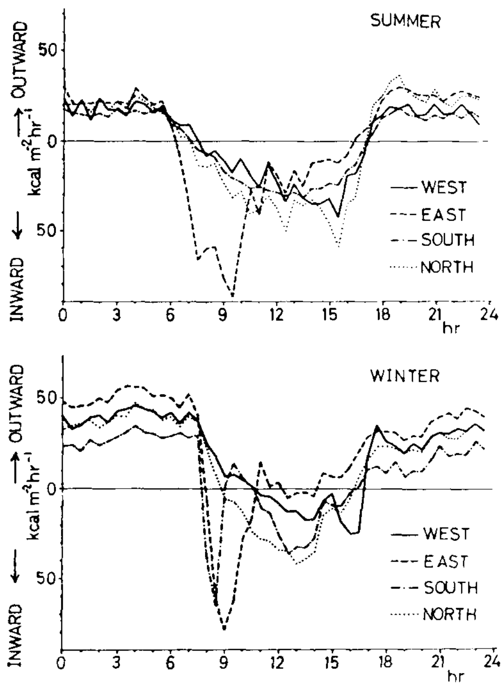

Fig. 4.8 Heat fluxes through external wall surfaces in summer and winter (after Sagara and Horic, 1978).

Only a limited store of information on this flux is available; estimates have been made for a few urban areas. The problem lies with the statistics on fuel use. Not all fuel delivered to a given city is used there, such as gasoline sold there but burned by driving in the countryside. Often the statistical districts encompass substantial amounts of rural area economically linked to a particular city.

For a few cities estimates can give an idea of what orders of magnitude are dealt with. Particularly notable are the differences of heat rejection by unit areas of various metropolitan localities as shown in Table 4.15.

The extraordinary concentration of energy use in the crowded quarters of Manhattan is not equaled elsewhere. Most other places cover a greater surface area and have considerably less population density. Only for a few cities are there more detailed studies for various subdivisions of metropolitan areas: London (McGoldrick, 1980), Montreal (East, 1971), and Sydney, Australia (Kalma *et al.*, 1973) are pertinent examples. The Australian study showed that the ratio of annual energy use in the inner city to the outlying sector of the Sydney Statistical Division was about 360 to 1. Such differences

TABLE 4.15

Metropolitan Energy Use[a]

City	Population (millions)	Area (km^2)	Energy flux density ($W\ m^{-2}$)
Manhattan	1.7	234	630
Moscow	6.4	878	127
Sydney (city)	0.1	24	57
Chicago	3.5	1800	53
Budapest	2.0	525	43
Aruba	0.06	180	35
Brussels	1.0	400	28
Cincinnati	0.54	200	26
West Berlin	2.3	233	21
Los Angeles	7	3500	21
Sheffield	0.5	49	20
Fairbanks	0.03	36	19
St. Louis	0.75	250	16
Hongkong	4.4	92	3

[a] Sources: Bach, 1970; Borisenkov, 1977; Kalma *et al.*, 1973; SMIC, 1971.

must find expression in the micrometeorological fabric of the urban area. Kalma *et al.* (1973) estimated that the annual artificial heat generated in Sydney, in 1970, was 24.74 J (234.5×10^{12} Btu). For Montreal, East calculated the anthropogenic heat production at 39×10^{16} J (9.33×10^{16} cal) or 24 percent of the total energy transformation of the urban area.

A very detailed map of energy use in Greater London for the year 1971 was undertaken by McGoldrick (1980). He found that the average daily artificial heat release in the outer districts is about 0–5 W m^{-2}. In the city center there are several square kilometers where the average exceeds 100 W m^{-2}, with a maximum for 1 km^2 in the inner city of 234 W m^{-2}. The daily global solar radiation for the area averages 106 W m^{-2}. It is estimated that Greater London rejects from domestic, commercial, institutional, industrial, and traffic energy use 17.4 GW/day during the year and 21.8 GW/day in December when energy use reaches its annual peak.

For one specific sector of use of energy a very good relation to meteorological elements exists, namely space heating. Turner (1968) showed for St. Louis that a very high correlation exists between daily temperatures in winter and the demand for gas and steam for heating purposes, as Fig. 4.9 demonstrates. The linear regression between temperature and energy use on weekdays explains 82 percent of the variance. East (1971) performed a similar analysis in which he related the space heating urban energy flux in Montreal to both temperature deviation from a specific base 65°F (18°C), the well-known degree day value, and a wind function. The linear regression had the following form

$$Q_{\mathrm{Sh}} = a + b(1 + u)\,\mathrm{dd} \qquad (4.11)$$

where

- Q_{Sh} heat flux from space heating
- a, b constants
- u wind speed
- dd degree-day value

Under very simplified conditions one can set the temperature rise attributable to the anthropogenic heat Q_{p} in urban areas to be

$$\Delta T_{\mathrm{Gp}} = \left[Q_{\mathrm{p}} a \frac{d\Theta}{dz} \Big/ c_{\mathrm{p}} \phi \bar{u} \right]^{1/2} \qquad (4.12)$$

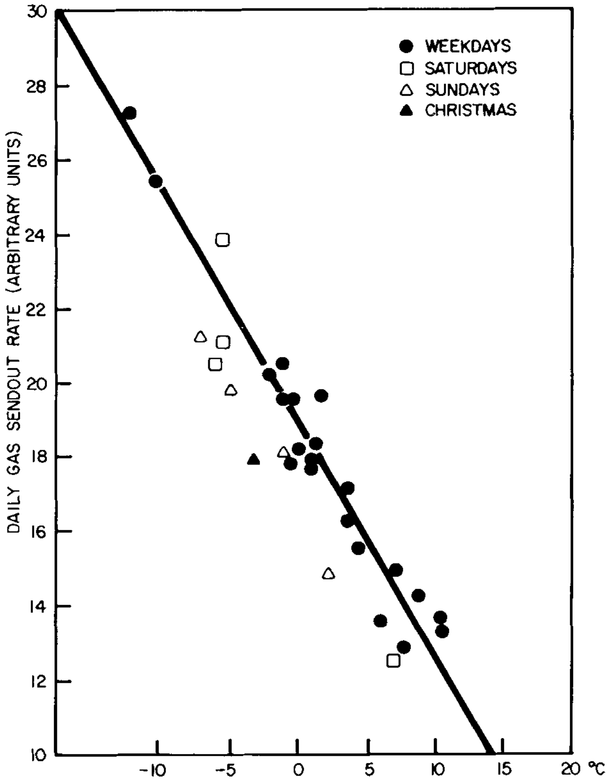

Fig. 4.9 Use of heating gas in St. Louis, Missouri, as a function of daily temperature (from Turner, 1968).

where

a	area of the city
$d\Theta/dz$	vertical lapse rate of potential temperature Θ with height z
c_p	specific heat of air at constant pressure
ϕ	density of air
$\bar{u}$	mean wind speed

The fact that Q_p and $d\Theta/dz$ vary considerably in an urban area and that the lapse rate is itself a function of ΔT, makes this only a first approximation.

There has also been some discussion of the significance of meta-

TABLE 4.16

Anthropogenic Heat Production during the Course of the Day[a]

Heat source	Weighted share (%) at various daytimes				
	Hour				
	0800	1300	2000	Night	Daily share
Stationary	71	64	71	45	66.6
Mobile	69	45	25	12	33.3
Metabolism of organisms	0.05	0.2	0.1	0.02	0.1
($W m^{-2}$) metabolic heat	0.36	0.29	0.26	0.14	0.26

[a] In relative values (Bach, 1970).

bolic heat of people and animals on the urban heat budget (Terjung, 1970). For a population of a million persons and a corresponding number of domestic animals one can estimate metabolic heat production at $\sim 5.3 \times 10^{15}$ J (5×10^{12} Btu). This, depending on the other anthropogenic heat production, is not more than maximally 3 or 4 percent of the total urban energy rejection and, in most cases, less than 1 percent. This component is usually neglected in urban energy budgets. Based on data collected in Cincinnati during summer, Bach (1970) estimated the contribution of various sources of anthropogenic heat production throughout the day, as shown in Table 4.16.

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