The Urban Climate

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6 The Urban Wind Field

The changes brought about by urbanization in the local atmospheric boundary layer have a notable effect on the low-level wind. This is caused by the heat island and the change in surface roughness.

6.1 EMPIRICAL EVIDENCE FOR WIND ALTERATIONS

An early account of Kremser (1909) called attention to a decrease of wind speed in urban areas. An anemometer on top of a high school building in the suburbs of Berlin, at 32 m above the surface showed in the first decade a mean wind of 5.1 m sec^{-1}. In the next decade the formerly free terrain turned into apartment housing and the anemometer was now only 7 m above the average roof level. The mean wind speed had dropped to 3.9 m sec^{-1}, a 24 percent reduction. This is almost exactly the difference observed for the anemometers at Central Park in New York City and at La Guardia Airport. Nearly all studies show an increase in the number of calms observed in town compared with the rural areas and a notable 10–20 percent reduction in the speed of maximum winds.

In Columbia, Maryland, during the early period of urbanization, there was a gradual increase in weak winds and a decrease of strong winds. During a 6-yr period of observations 3-hourly wind-speed observations were compared with simultaneous observations at the nearby Baltimore–Washington International Airport. The Columbia wind speed was expressed as a percentage of the airport wind speed. These observations were grouped in three classes:

(1) Columbia wind speed was less than 70 percent of that observed at the airport;
(2) 70–99 percent of airport speed; and
(3) the speed was as high or higher in Columbia than at the airport.

The results are shown in Fig. 6.1. In 1969 wind speeds in Columbia were higher than at the airport in 25 percent of the cases, but by 1974 this class had dropped to 14 percent of the observations. In the same interval the wind speeds, which were less than 70 percent in the newly urbanized area of those recorded at the airport, had increased from 43 to 65 percent.

A very interesting case of a long-term trend in wind speed caused by urbanization has been published by Rubinshtein (1979). It showed for the growing town of Gantsevitchi in White Russia a monotonous drop from mean annual velocities of 3.9 m sec^{-1} in 1945 to 2.5 m sec^{-1} in 1971, as shown in Fig. 6.2. This is a reduction of about 36 percent.

Another excellent example of the urban effect on wind speed has been reported by Zanella (1976) for Parma located in the Po plain of northern Italy. With a progressively expanding urban area the annual number of calms is 55 percent of the observations, but only 48 percent at the airport. In the winter season these percentages go up to 82 percent in the city versus 64 percent at the airport. In that season and in spring, wind speeds are at all hours of the day lower in the city than at the airport. In summer and autumn the difference vanishes only in the early evening hours. Most revealing is the change from decade to decade shown in Table 6.1. The total speed

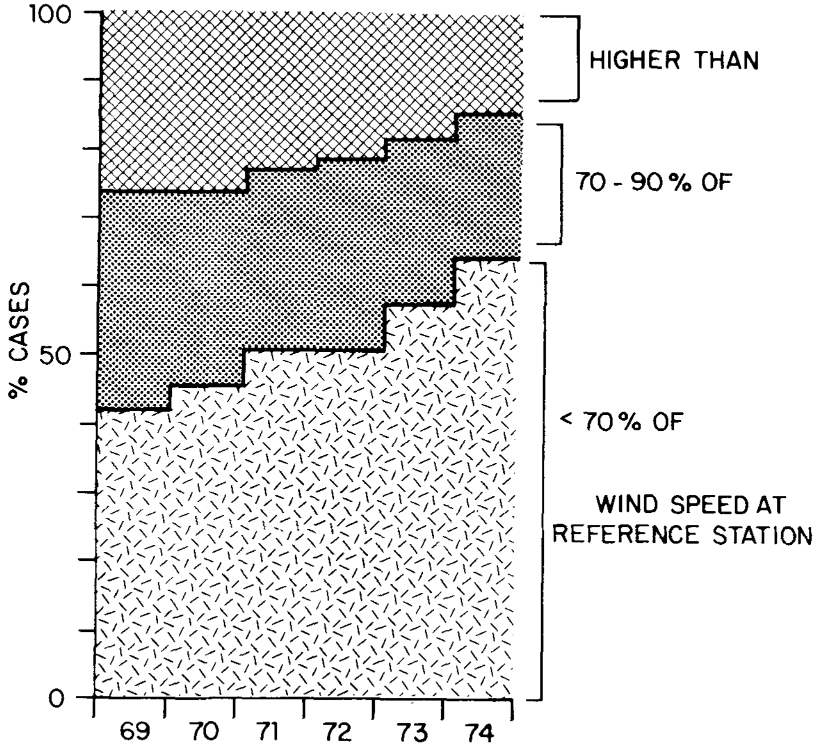


Fig. 6.1 Change in wind speed classes during rapid growth era of Columbia, Maryland.

reduction in a quarter century has been 39 percent. All directions appeared to be equally reduced.

The low winter wind speeds in Parma are attributable to the regional climate. In other regions the urban wind speeds are materially affected by the vegetation. Dirmhirn and Sauberer (1959) indicated that in Vienna when the deciduous trees were in leaf, urban wind speeds dropped 20–30 percent. The same effect was observed by Frederick (1961) in Nashville, Tennessee. There 24-hr wind movement was measured by totaling anemometers on 32 utility posts, about 10-m above the surface. The ratio of the wind movement at these urban stations was compared to that at the airport. This analysis showed that during summer in an area of numerous de-

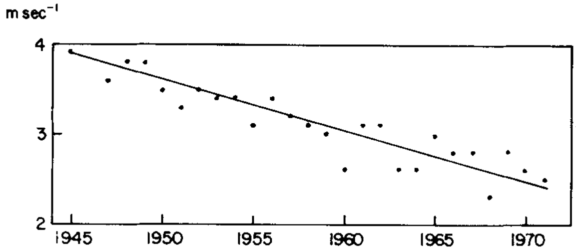


Fig. 6.2 Time series of mean annual wind speed in the growing city of Gansevitchi, U. S. S. R. (based on data by Rubinshtein, 1979).

ciduous trees, spread throughout the city, wind speeds were reduced 20–30 percent compared to sites not affected by trees. After defoliation wind speeds at the sites with trees increased by 25–40 percent. On the whole, reductions of 30–60 percent in wind speed at the various urban sites were noted.

A comparative set of observations above roof level was made in Vienna (Steinhauser *et al.*, 1959). One anemometer was installed on the roof of the Technical University in the densely built-up center of the city, the other at the edge of the city, in a parklike setting, on an observatory. Ratios of wind speeds at the two sites for various wind directions and speed classes were compiled. The results are shown in Table 6.2. The table shows a number of effects. For both listed wind directions the greater speeds are less affected by the urban area than the weaker ones. In summer the suburban site shows for west winds a higher ratio than in winter. The cause is the already discussed leaf effect, with many deciduous trees near the observatory. For the SE winds the ratios for the stronger winter winds and all speed classes in summer are reversed: the city wind speeds are

TABLE 6.1

Changes in Mean Wind Speeds in Parma, Italy, in Three Consecutive Decades[a]

Interval	January	April	July	October	Year
1938–1949	0.5	1.8	1.8	1.0	1.3
1950–1961	0.5	1.4	1.4	0.7	1.0
1962–1973	0.3	1.0	1.3	0.6	0.8

[a] Wind speeds (m sec^{-1}). Adapted from Zanella (1976).

TABLE 6.2

Wind Speed Ratios, Center–Suburb, in Vienna, Austria[a]

		Wind speed (km hr^{-1})			
Wind direction	Season	5 (1.4)[b]	15 (4.2)	25 (6.9)	35 (9.7)
W	Winter	0.5	0.6	0.7	0.77
	Summer	0.75	0.82	0.9	0.85
SE	Winter	0.5	0.8	1.05	
	Summer	1.0	1.25	1.25	

[a] After Steinhauser *et al.* (1959).
[b] Values in parentheses in m sec^{-1}.

higher than suburban values. This is produced by higher roughness in the wind-fetch area on the approach to the observatory.

An early observation has been low-level convergence of air flow into urban areas. This is a logical consequence of the heat island, which creates an unstable vertical lapse rate of temperature and thus induces a rising air current. More details on that will be presented in Section 6.2. Observations of inflow of air to the urban center are easiest at night and an example of such flow is shown in Fig. 6.3 (Stummer, 1939). Usually, even with a well-developed nocturnal heat island, concentric convergence is a rare case. The irregularity of urban structure and surrounding terrain dissimilarities prevent that. There are a few well-documented cases of centripetal urban air flow in the literature. One deserves to be singled out because of the ingenious method of observation. This has been reported by Okita (1960) from the city of Asahikawa (Hokkaido, Japan). This place of about 190,000 inhabitants has in winter frequent radiation fog at temperatures much below freezing. On some of these occasions heavy rime is formed on the windward side of trees by impinging of supercooled droplets. From these Okita was able to determine the wind direction at many localities around the city. On days with a minimum temperature difference (urban–rural) of about 4°C, implying a maximum heat-island difference during the night of 8°C, the convergence was almost perfect.

The nocturnal convergence into the urban areas is the reason for the observation that wind speeds at night in cities are not as much weakened as in daytime and occasionally can be stronger than in the

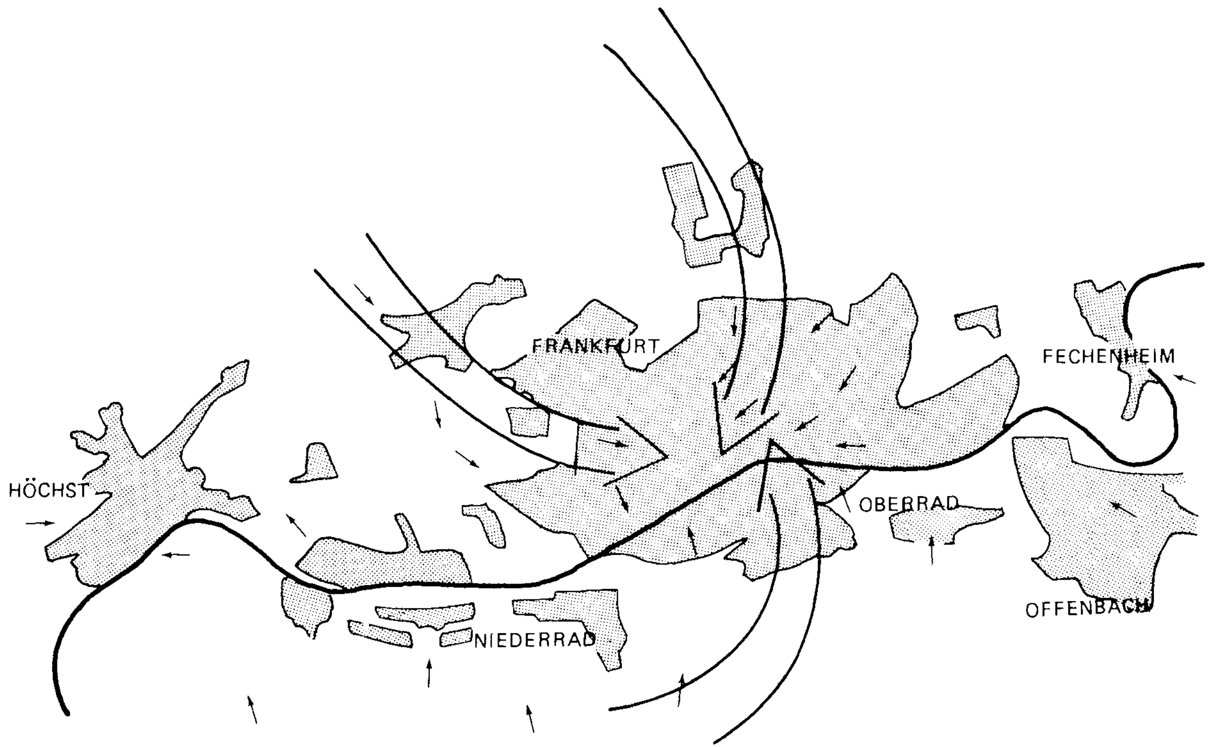


Fig. 6.3 Nocturnal wind convergence on calm nights into Frankfurt am Main, Germany (adapted from Stummer, 1939).

country. That happens in particular when under nearly calm conditions a strong rural inversion builds up while the lower layer in the city remains unstably stratified. Chandler (1965) shows this well for London in a comparison of winds at 0100 hr GMT and 1300 hr GMT in the city center and the airport, as given in Table 6.3.

Lee (1979) clearly showed for London that the small nocturnal wind speed difference in and outside the city is a direct function of the frequency and intensity of the rural inversions. And although in daytime wind speeds are reduced on an average by 30 percent in the city, the average nighttime reduction is only 20 percent. In smaller towns the day–night differential is not nearly as pronounced.

It is also essential to understand that the wind flow at night into the urban area is not a steady one. In large urban areas the isotherms of the heat island, if it has indeed a single core, are not equidistantly spaced. They are usually crowded near the edge of the built-up area. Meteorologically this is analogous to a cold front where temperatures change abruptly over short distances. This characteristic of the heat island can lead to sharp pulses of cooler country air invading the city at night, often with a notable wind gust. As early as 1925 Schmauss presented a case of that type from Munich and called attention to these miniature cold fronts. Figure 6.4 shows two such invasions in one night. In daytime the wind flow into the urban area is not as readily discernible and in many cases can only be noted as a deformation of the general wind field. Occasionally on hot summer

TABLE 6.3

Average Seasonal Wind Speeds[a] at the Airport and the City of London (1961–1962) at Different Hours

Season	0100 hours Airport	0100 hours City difference (higher)	1300 hours Airport	1300 hours City difference (lower)
Winter	2.5	0.4	3.1	−0.4
Spring	2.2	0.1	3.1	−1.2
Summer	2.0	0.6	2.7	−0.7
Autumn	2.1	0.2	2.6	−0.6
Year	2.2	0.3	2.9	−0.7

[a] Values in m sec^{-1}.
[b] After Chandler (1965).

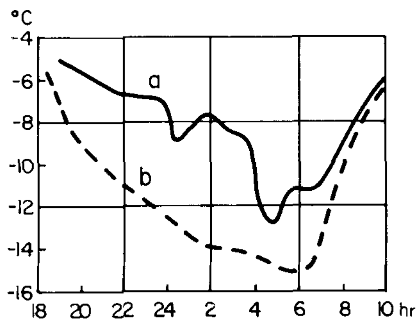


Fig. 6.4 Nocturnal march of temperature in Munich, Germany (solid line, a) and at a suburban station (dashed line, b) during winter night, showing in "a" two frontlike invasions into the city (after Schmauss, (1925).

afternoons with weak gradient winds the convergence can be directly observed (Fig. 6.5).

Berg (1947) made the attempt to use the elementary relations between temperature, pressure, and wind to estimate the speed of the country breeze.[1] He figured that there is a 5°C ΔT_{u-r}, decreasing to 1°C at 500 m and vanishing at 1000 m. Assuming a 10-km distance for this temperature difference leads to a pressure difference of 0.07 mbar (7 Pa). Such a small pressure difference is not measurable with ordinary equipment and has indeed never been observed in urban climate studies, but the calculated resulting surface wind is about 3 m sec^{-1}, a value entirely compatible with the observations.

One additional fact about the surface near winds in urban areas is readily noted in anemograms. The wind turbulence shows a distinct increase in the gusts with higher frequencies compared to open areas. A study of the power spectra of wind speeds at a large apartment complex and a nearby airport (Badger, 1973) showed that at the airport, where the wind had free and rather unobstructed fetch, most of the power is in the low frequencies (>1 min) and the spectrum is very close to a red-noise spectrum. This means that the turbulent eddies are large. In the apartment complex there is a notable shift in the power spectrum to higher frequencies. At that end of the spectrum there are notable departures from red noise and small eddies clearly dominate in that rough urban environment.

[1] We use the term "country breeze" in the same sense as those of other secondary circulations, such as mountain breeze or sea breeze, each indicating, as in general meteorology, the direction the wind is coming from.

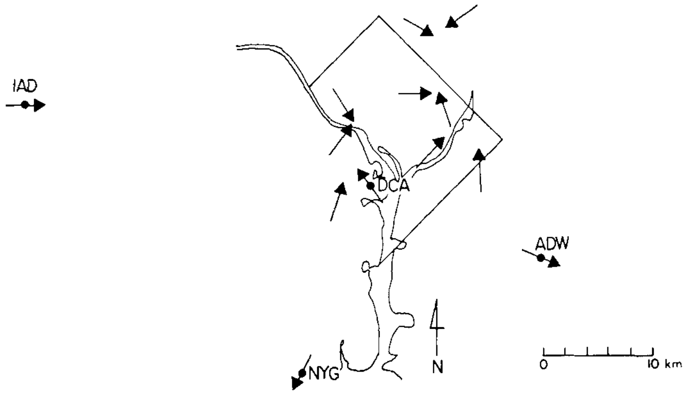


Fig. 6.5 Wind convergence on a hot summer afternoon in Washington D.C. IAD, Dulles Intl. Airport; DCA, Washington Natl. Airport; ADW, Andrews Airbase; NYG, Quantico Marine Base.

6.2 VERTICAL WIND STRUCTURE IN URBAN AREAS

Several techniques have been complementarily used to gauge wind conditions over cities. The simplest are instrumented towers, which have increased in number in cities for broadcast and telecommunication purposes. But many years of observations are available from Paris where Alexandre Gustave Eiffel started wind observations on the 300-m-high tower designed by him in 1890. Television towers are usually less than 200 m in height but if equipped with anemometers at various levels they can contribute useful information. Interpretation of observations thus obtained may require corrections if the wind flows through the tower structure before striking the anemometer (Izumi and Barad, 1970).

Two procedures using balloons have been employed. The vertical profile is obtained by ordinary pilot balloon observations. If simultaneous values are obtained at various points around a city, convergence and divergence can be determined. Use of constant-volume balloons, so-called tetroons, has been made to obtain trajectories of flow over urban areas. These tetroons fly at a level of constant density. They carry a transponder and are tracked by radar. From their track and flight level the vertical component of air motion can be determined. In addition, observations from low-flying aircraft can contribute to the wind information over urban areas. Yet, all these airborne observations have the drawback of being essentially sporadic. But they can reach into higher air layers than the towers, which furnish continuous data.

The vertical profiles in low levels show one phenomenon clearly. The wind, which is notably retarded below the level of the buildings, shows usually a minor speed maximum just above the mean roof level, that can be designated as a roof-top jetlet. The increase is only about 1 m sec^{-1} on an average. However, in very strong winds, for example in typhoons, gust speeds may be much higher than surface winds and can dominate the total low-level wind profile. Observations on a 333-m-high tower in Tokyo by Arakawa and Tsutsumi (1967) showed for winds ≥ 30 m sec^{-1} that a level above the roof level of 30-m, gust speeds were commonly 10 m sec^{-1} higher than just below the roof niveau.

Tower observations have long been used to characterize air flow in the planetary boundary layer. Obviously the massive obstacles introduce major perturbations, but Munn (1970) has cogently pointed out how important it is to know this urban flow for engineering applications and for pollutant-dispersal estimates. Over smooth surfaces meteorologists have adopted the theory of flow in fluids, first developed by Ekman (1905) in a classical treatise. By analogy this was transferred to the atmosphere. The theory states that under ideal conditions, the surface influence causes the vectors of motion in successive layers, up to the layer where friction ceases, to have an effect on the wind: to form an equiangular (logarithmic) spiral. The wind, unimpeded by friction is called the geostrophic wind, represented by

$$V_G = -\frac{1}{\rho f}\frac{\partial p}{\partial x} \tag{6.1}$$

where

- V_G geostrophic wind
- ρ atmospheric density
- f Coriolis parameter ($2\Omega \sin \phi$; Ω, angular velocity; ϕ, latitude)
- $\partial p/\partial x$ horizontal pressure gradient from higher to lower pressure

The geostrophic wind blows parallel to the isobars at that level. The nongeostrophic wind components below, in the ideal case, make an angle of 45° with the tangents to the spiral. One can express for averaged wind vectors the relation between the geostrophic wind and the wind in the layers below as

$$\bar{v}_z - \bar{V}_G = \frac{1}{f}\frac{d}{dz}\left(\frac{\tau_z}{\rho}\right) \tag{6.2}$$

where

- $\bar{v}_z$ average wind vector at any level z below the geostrophic wind level
- τ_z shear stress
- ρ atmospheric density

The shear stress, also termed Reynolds stress, can be determined from the turbulent components of wind speed, i.e., the departures from the average in the three components u', v', w'. For convenience a parameter called friction velocity u_* has been introduced. It equals $(\tau/\rho)^{1/2}$ and is a function of the mean horizontal and vertical wind fluctuations

$$\frac{\tau}{\rho} = u_*^2 = \langle -u'w' \rangle \tag{6.3}$$

These wind fluctuations can be measured on towers.

In a neutrally stratified atmosphere, i.e., with an adiabatic vertical temperature lapse rate of 1°C/100 m, aerodynamic theory then permits us to derive a vertical wind profile (Taylor, 1952)

$$\bar{u}_z = \frac{u_*}{k}\ln\left(\frac{z}{z_0}\right) \tag{6.4}$$

where

- k von Kármán's constant (~ 0.4)
- z height
- z_0 friction height or roughness parameter

This is usually referred to as the logarithmic wind profile. In any environment the value of z_0 can be determined by wind observations at two heights:

$$\ln z_0 = \frac{\bar{u}_2 \ln z_1 - \bar{u}_1 \ln z_2}{\bar{u}_2 - \bar{u}_1} \tag{6.5}$$

where the subscripts refer to the values at the two heights.

Table 6.4 shows a number of values reported in the literature for urban areas. As is readily seen, the values scatter widely, by almost an order of magnitude. This is the result of varying roughness upwind of the measuring tower, the length of fetch, and the degree to which the neutral stability condition was fulfilled. On the whole, however, there is no doubt that the urban roughness parameter is generally an order of magnitude larger than over agriculturally used land. From measurements made in the Paris, France area, including the values from the Eiffel tower (Dettwiller, 1969), one can derive values ranging from 2 to 5 m.

A theoretical approach for estimating the roughness parameter has been suggested by Lettau (1969, 1970) using the similarity principle. According to this scheme, z_0 can be predicted if the effective heights of obstacles are known, the subtended area encountered by the wind, and the area covered can be ascertained. In that case

$$z_0 = \frac{1}{2}\frac{Ha}{A} \tag{6.6}$$

where

H height of obstacle (m)
a silhouette area encountered (m^2)
A area covered by obstacles (m^2)

and the factor $\frac{1}{2}$ approximates the average drag coefficient of roughness elements. Estimates of roughness parameters using this

TABLE 6.4

Measured Roughness Parameters z_0 in Urban Areas

Locality	z_0 (m)	Source
Kiev, U.S.S.R.	4.5	Ariel and Kliuchnikova (1960)
Fort Wayne, Indiana	3.0	Csanady *et al.* (1968)
Minneapolis, Minnesota	2.0	Deland and Binkowski (1966)
Tokyo, Japan	1.7	Yamamoto and Shimanuti (1964)
Liverpool, England	1.2	Jones *et al.* (1971)
Austin, Texas	0.4–2.4	Peschier (1973)
Reading, England	0.7	Marsh (1969)
Cambridge, Massachusetts	0.5–2	Dobbins (1977)
Columbia, Maryland	0.7	Landsberg (1979)

TABLE 6.5

Aerodynamic Roughness Based on Building Characteristics in Urban Areas[a]

Parameter	Building type		
	Low	Medium	High
Height (m)	4	20	100
Silhouette (m^2)	50	560	4000
Built-up area (m^2)	2000	8000	20,000
Calculated z_0 (m)	0.5	0.7	10

[a] According to Lettau (1970).

scheme are given in Table 6.5. The values obtained this way seem to yield realistic numbers that are quite comparable to the measured figures. Lettau uses the logarithmic wind profile in the following form for the Ekman boundary layer, using logarithms to base 10:

$$\bar{u}_z = 5.5cV_G \log_{10}(1 + z/z_0) \tag{6.7}$$

where

$\bar{u}_z$ mean wind at height z
c the drag coefficient[2]

Lettau and others have observed and calculated Ekman spirals for various drag conditions. The observed values only approximate the theoretical model, which has been used to simulate urban conditions for wind tunnel experiments. Dobbins (1977) offers measured values from observations made in Cambridge, Massachusetts. He estimated from his data that the atmospheric boundary layer ranges from 10 to 40 times the roughness height z_0. A typical hodograph from his work is shown in Fig. 6.6. The hodograph is the envelope of the wind vectors at various heights above the surface.

In other representations of the vertical wind profile, a simple power law was used; this is especially notable in engineering litera-

[2] This is related to the nondimensional Rossby number Ro = $V_G/z_0 f$, where f is the Coriolis parameter. Csanady (1967) gave this relation: Ro = $V_G/z_0 = 0.4((1/c) - 115)^{1/2} + 1.15 \ln(1/c) - 1.52$.

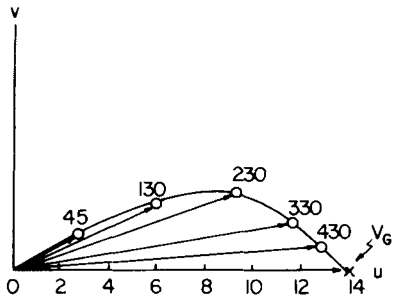


Fig. 6.6 Hodograph of wind field over urban area of Cambridge, Massachusetts. Numbers on wind-vector envelope are heights in meters; abscissa shows wind speed in meters per second (adapted from Dobbins, 1977).

ture. Two forms are common:

$$\bar{u}_z = u_1\left(\frac{z}{z_1}\right)^{\alpha} \tag{6.8}$$

and

$$\bar{u}_z = V_G\left(\frac{z}{z_G}\right)^{\alpha'} \tag{6.9}$$

where α is an exponent representing all the friction factors. The version using the geostrophic wind was introduced by Davenport (1965). This author gives the value of the exponent α' and the average level of the gradient wind as shown in Table 6.6.

The average vertical wind distribution in percent of the gradient wind is depicted in Fig. 6.7. Obviously such an idealized picture is only valid for a neutral stratification, and wide deviations, depending on the synoptic situation and time of day, are observed in practice.

TABLE 6.6

Measurements of the Vertical Wind Profile for Different Surface Roughnesses[a]

Terrain characteristic	Exponent of power law	Gradient wind level (m)
Open country, flat	0.16	270
Suburban settlement	0.28	390
Inner cities	0.40	420

[a] Davenport (1965).

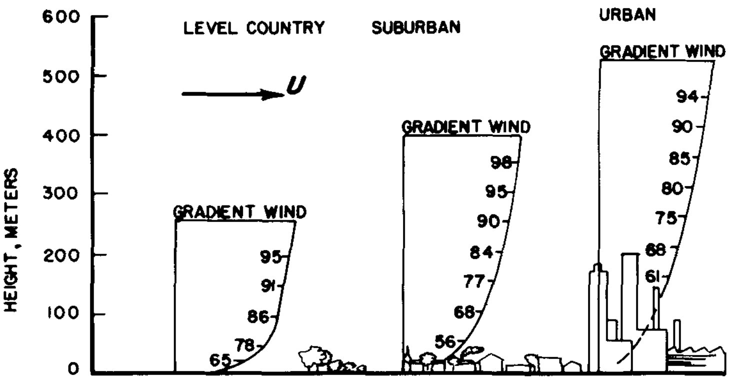


Fig. 6.7 Vertical wind speeds, in percent of the gradient wind at various heights over terrain of different roughness (after Davenport, 1965).

The exponent in Eq. (6.8) has been determined empirically to vary in open country between 0.07 and 0.25. A commonly used value is 0.14. Borisenko and Zavarina (1968) have determined the dependence of the value of α in the lowest 100 m of the atmosphere on the roughness parameter z_0. An excerpt from their table, applicable to suburban and urban environments is shown in Table 6.7. The relation between these two variables is fairly linear in this range and can be approximated by $\alpha = 0.12z_0 + 0.18$. The value for the 2-m

TABLE 6.7

Relation between Roughness Parameter z_0 and Exponent α in Power Law of Vertical Wind Profile in the Lowest 100 Meters[a]

z_0 (m)	α
0.1	0.18
0.2	0.21
0.5	0.25
1.0	0.31
2.0	0.42

[a] Adapted from Borisenko and Zavarina (1967).

roughness parameter compares well with an α value of 0.44 measured over the most densely built-up area of Columbia, Maryland. Measurements made by Jackson (1978) in the windy city of Wellington, New Zealand, between 10 and 70 m above ground, yield an average α of 0.5. For northerly winds z_0 was determined at 3.8 ± 0.6 m. The greatest turbulence was below the average roof height at the 10-m level. It rapidly decreased to the 70-m height.

A great deal of information not only on wind flow over urban areas but on the vertical component of wind motion has been obtained from tetroon flights. Much of this information has been obtained over New York City (Hass *et al.*, 1967; Angell *et al.*, 1968; Druyan, 1968). These flights were made at about 300 m and their main objective was the collection of facts about air trajectories for travel of air pollutants. Obviously, because of the coastal location one has to be careful not to generalize the results too much as characteristic for urban climates. Yet a few points stand out. Over the center of the urban area in Manhattan the tetroon drift was only about $\frac{2}{3}$ of the geostrophic wind. The vertical motion obviously was greatly influenced by the atmospheric stability. Early morning flights were far more level than midday flights when temperature lapse rates were large. In midday the heat-island effect caused notable vertical lift in the balloon flights over the city, but when crossing the Hudson River, where cool water stabilized the air, the balloons returned to lower levels. This is very well demonstrated in Fig. 6.8, which shows the vertical track of a flight near noon across Manhattan Island. The flight levels over water surfaces are low, but over the built-up areas they float several hundred meters higher. The jump

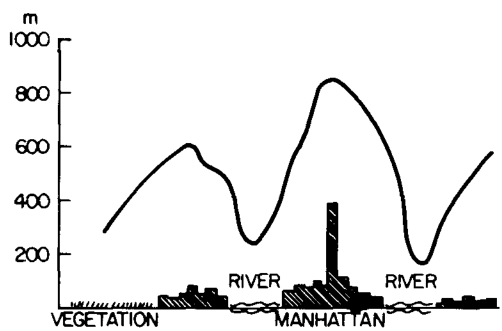


Fig. 6.8 Vertical track of a tetroon on a midday flight across Manhattan Island, New York (after Hass *et al.*, 1967).

over the Empire State Building (381 m) is particularly spectacular, where the balloon climbed above 800 m.

Vertical velocities determined from the tetroon flights over New York City were twice as large in midday than either in the early morning or evening. Between 0900 and 1500 hr local time, at a height of 300–500 m, it exceeded 1 m sec^{-1} in 15 percent of the observations and 0.4 m sec^{-1} for 50 percent of the observations. The convection currents over cities, resulting in "bumpiness," have been well-known since the early days of manned flight.

Another fact noted is that the tetroons have a tendency to move toward lower pressure on the days when the sea breeze did not interfere with the flights. The tetroon observations were confirmed and extended by the work of Ackerman (1972, 1977) and Auer (1975) over St. Louis in the METROMEX study. The information gathered there is likely to be more representative than the New York data because of the absence of topographic disturbances. The analysis by Ackerman was based on pilot balloon observations to which Auer added wind information obtained by aircraft tracked by Doppler radar.

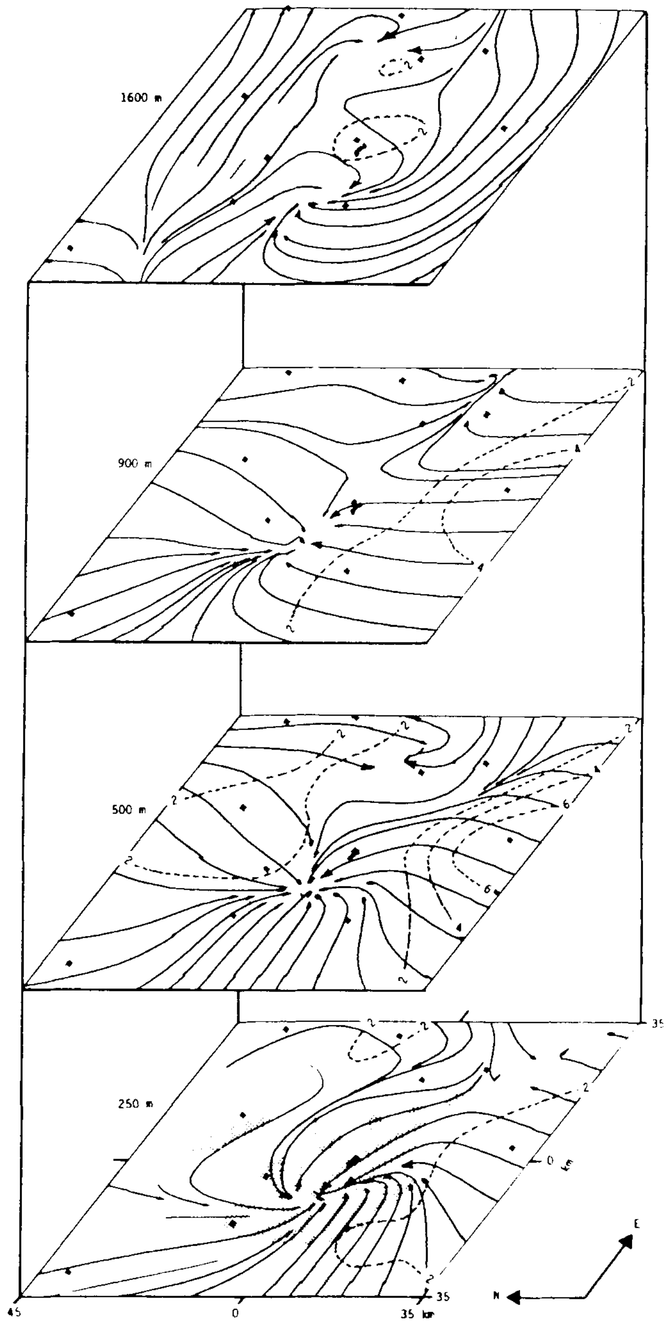
These studies clearly showed convergence over the urban area and slightly cyclonically curved trajectories. Obviously, the wind field distortion was most notable when regional winds were weak but still observable even in strong winds. The divergence is calculated from the wind field by measuring the increase in area of the horizontal winds subtended in their passage over the city during a given time interval

$$D = \frac{1}{A_0}\frac{\Delta A}{\Delta t} \qquad (6.10)$$

where

- D divergence (if negative: convergence)
- A_0 initial arc
- ΔA difference between A and distortion of winds in time lapse Δt

Of course, the procedures permit only limited samples, but in her work Ackerman was able to use data for 21 undisturbed summer afternoons in the lowest 1500 m above St. Louis. These data should reflect the urban contribution to the wind field quite plausibly. In the lowest layers there is mass convergence into the city with D values of -1×10^{-4} sec^{-1}, reaching values up to $\sim 1.7 \times 10^{-4}$ sec^{-1}. At



1600 m
900 m
500 m
250 m
45
0
35 km
35
0 km
35
N
E

about 300 m the convergence changes to divergence. Calculations of vertical motion yielded mean values of $\bar{w} = 0.07$ m sec^{-1} to maximally 0.3 m sec^{-1}. The afternoon values showed convergence into the city in about 60 percent of all observations, with the greatest disturbance shown in the early afternoon at the time of the maximum temperature. Figure 6.9 shows the urban wind perturbation over St. Louis for a typical case at midday.

6.3 URBAN WIND FIELD AND OTHER MESOSCALE CIRCULATIONS

The urban wind field is rarely simple. Even when the synoptic situation is least complicated, with clear skies and weak winds in the center of an extended high-pressure area, small differences in local topography will cause irregular air flows. Thus the broad generalizations that have been introduced into the literature are at best guideposts to the phenomena. Even in the case of St. Louis, where the terrain has relatively less influence than in case of other cities, one can only take the data presented by Dannevik, *et al.* (1974) as order-of-magnitude estimates, shown in Table 6.8.

Similarly, the values obtained for the equally simple terrain of Oklahoma City by Angell *et al.* (1973) in strong winds (≥ 13 m sec^{-1}) still raise notable questions. The daytime values when the heat-island effect is small, measured by tetroon and on a 460-m-tall television tower, suggest that the city acts simply as an obstacle to the air flow. Thus there is a tendency for streaming around the city and a vertical component introduced by the barrier effect. The rising current in daytime is about 0.3–0.4 m sec^{-1} close to values obtained in other experiments. A 5° turning toward the low-pressure side was noted and the surface stress estimated at 2 dynes cm^{-2}.

But when we turn from these very simple settings, many complexities arise. It should be emphasized here that cities usually have

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Fig. 6.9 Wind perturbation field at noon on a summer day over St. Louis, showing the convergence field by streamlines (solid curves), perturbation isotachs in meters per second (dashed curves) at various levels above the city (heights are given in meters above mean sea level) (from Ackerman, 1972).

TABLE 6.8

Estimates of the St. Louis, Missouri, Heat-Island Circulation Elements[a]

Element	General magnitude
ΔT_{u-r}	≥2°C
V_G (900 cPa)	≥5 m sec^{-1}
$\bar{u}$	2 m sec^{-1}
$\bar{w}$	0.3 m sec^{-1}
Diameter of:	
surface inflow	30 km
updraft	7 km
Depth of circulation	1 km

[a] From Dannevik *et al.* (1974).

arisen because of some terrain features favorable to one or more human activities. Most common are locations favoring various forms of traffic. Thus, river valleys, bays, and shore locations on lakes and on the sea are preferred spots for urban areas. These landscapes develop invariably their own secondary circulations, such as land and sea breezes, and mountain and valley breezes. In some instances both of these circulations can be present. They interact with the urban-induced circulation and may, at times, dominate it.

A prize example of the latter variety is the Los Angeles basin, where land and sea breezes reinforce mountain and valley breezes. They cause inland flow in daytime and seaward flow at night. Angell *et al.* (1966) were the first to apply their tetroon techniques to the air flow in that area. Flights at 300 m clearly showed the reversal of flow from land to sea and sea to land as a function of time of day. Vertical velocities were again of the order of magnitude of 0.5 m sec^{-1} over land and, in the stable air over water, about 0.2 m sec^{-1} at sea. The land and sea breeze seemed not to exceed 2 m sec^{-1}. Although this study did not specifically refer to any mountain-induced circulation, its existence in that area is well known. As far as one can tell, the urban influence is completely masked. It is regrettable that, at least in the air pollution literature, the completely uncharacteristic patterns of the Los Angeles basin dominated urban meteorology.

As work on other metropolitan areas progressed, better information on urban effects on the secondary circulations became available. It is not too surprising that one finds that each city has its own patterns. The magnitude of urban influence can be minor or major. In some cases notable microclimatic differences due to cold-air drainage are clearly marked. Patterson and Hage (1979) recorded an October observation in Edmonton (Alberta) in the North Saskatchewan River valley. A steep bank, with a 50-m elevation difference caused a slope flow of 0.4 m sec^{-1} on a clear night to develop an intense inversion, with a vertical gradient of ~12°C/100 m. Less than a kilometer away in the city center the temperature inversion had only a gradient of ~2°C/100 m.

In the same area, Calgary (Alberta) on the flood plains of the Bow and Elbow rivers, 110 km east of the Rocky Mountains, developed similar conditions on a January night, as reported by Nkedirim (1980). A 200-m elevation difference in the town site led to cold-air drainage and to a reduction of the rural–urban temperature contrast. Although low temperatures of between −8 and −18°C were noted in the area, the difference from city center to city edge was only 4°C. Nkedirim commented that Summer's model (see p. 95) would predict a difference ΔT_{u-r} of about 6.6°C and he concludes that the cold-air drainage strengthens stability and reduces the effect of anthropogenic heat production on the urban heat island.

A rather opposite effect has been observed by Goldreich (1979a,b) in the Bezuidenhout Valley of South Africa. There nocturnal katabatic winds develop on clear winter nights, with site elevation differences of about 200 m. The slight breezes are generally below 1 m sec^{-1}. The city of Johannesburg apparently warms these up, and in the Bezuidenhout Valley temperatures are nocturnally 0.5–1°C higher than in air flowing down the tributary valleys. Unfortunately, no sustained observations of the effects of an urban area on a well-developed mountain and valley breeze system, occurring with high frequency is as yet available.

There is far more systematic information on the effect of urban areas on land and sea breezes, much of it due to Bornstein (1975) and collaborators for New York City, which shall be discussed later in greater detail. Auliciems (1979) has reported on some conditions for Brisbane (Queensland). There a weak land breeze from the southwest, of 2–3 m sec^{-1} develops in winter, probably weakened by a 5°C heat island. In summer a more vigorous sea breeze of 5–6 m sec^{-1} develops and is assumed, in conjunction with a weak nocturnal land breeze, to contribute to the recirculation of pollutants produced by the urban area.

Colacino and Dell'Osso (1978), made some studies of the development of the sea-breeze circulation in the region of Rome, Italy. Their observations include measurements of infrared surface temperatures on a day near solstice. At that time the sea-surface temperature was between 19 and 21°C. At 0600 hr, surface temperatures on land were 12°C in the country and 18°C in Rome. By 1330 hr the country surface temperatures had risen to 36°C and to 50°C in Rome. The maximum air temperatures were 21.2°C at sea, 26°C at a rural station, and 28°C in Rome. At night, from 2100 to 0600 hr, there was hardly any air motion. No well-developed land breeze existed, but at 0800 the sea breeze started to cross the coast at 4 m sec^{-1}. By 1500 hr it reached a peak velocity of 8 m sec^{-1}, a vigorous sea-breeze response to the 30° infrared temperature contrast between the sea and the city.

Bornstein and collaborators treat this penetration of the sea breeze in the New York area, by case studies, as a frontal phenomenon. Credit is properly given to Koschmieder (1935), who first recognized that the advance of the sea breeze resembles a miniature front. In some respect the work of Bornstein *et al.* (1978, 1979) yields results on the city influence that correspond to the obstacle effect observed by Angell *et al.* (1973) for Oklahoma City. Isochrone charts of the movement of the sea-breeze front reveal a frictional retardation by the built-up areas of New York City. Figure 6.10 shows very clearly the crowding of the isochrones over the built-up area (Fontana and Bornstein, 1979). From tetroon flights Anderson and Bornstein (1979) calculated the steepening of the sea-breeze front. They state: "Results show a dramatic steepening of frontal slope over the central urban area." In rural areas, the frontal slope $\Delta z/\Delta x$ was usually in the neighborhood of 1:100, a value not much different from that observed for many cold fronts. But urban values as low as 1:17 were noted in the city area. Six values in Anderson and Bornstein's data set were $\leq$1:71, which is ascribed to retardation at the surface and acceleration aloft. Vertical velocities of 0.65 m sec^{-1} were noted from these tetroon flights. The frictional retardation was also noted for the advance of the lake-breeze front in Chicago (Landsberg, 1958), where on many occasions this front showed a typical "nose", i.e., it advanced at 50 m, just above roof height, earlier over the built-up area than at the surface.

The effect of urban friction on fronts was first discussed by Belger (1940). For a small sample of four cases of migrating rain fronts, he

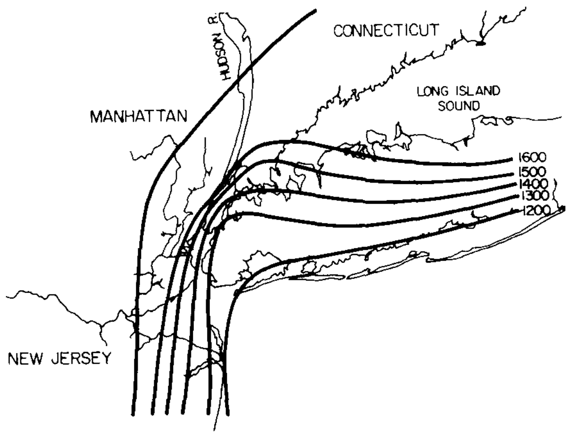

Fig. 6.10 Isochrones of a sea-breeze front in the New York City area, showing the retardation in the built-up area (after Fontana and Bornstein, 1979).

showed a retardation of 25 percent in speed over Berlin, Germany. This metropolis deformed the fronts and Belger speculated that this caused a local increase in rainfall by the lingering of the rain-producing discontinuity. We will pick up this theme again in Chapter 8. Loose and Bornstein (1977) also address the problem of frontal slowdown over New York City. From a number of cases with fronts moving from various directions they find an even more pronounced retardation of frontal movement in the urban area than Belger. Their value is about 50 percent, almost uniformly for fronts whose movement over rural areas ranged from 4 to 14 m sec^{-1}. They note that these observations agree qualitatively but not quantitatively with Bornstein's 1975 URBMET model which will be discussed in Chapter 7.

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