WORLD METEOROLOGICAL ORGANIZATION

TECHNICAL NOTE No. 134

REVIEW OF URBAN CLIMATOLOGY 1968-1973

by

Dr. T. R. Oke
CoSAMC Rapporteur on Urban Climatology

WMO - No. 383

Secretariat of the World Meteorological Organization - Geneva - Switzerland

THE WMO

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— To promote the establishment and maintenance of systems for the rapid exchange of meteorological information;

— To promote standardization of meteorological observations and ensure the uniform publication of observations and statistics;

— To further the application of meteorology to aviation, shipping, water problems, agriculture, and other human activities;

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CONTENTS

TABLE OF CONTENTS

FOREWORD

In view of the increasing requirements to study problems related to urban areas, especially in connexion with their climate conditions, the WMO Executive Committee, at its twenty-fourth session (Geneva, 1972), endorsed a proposal of the president of the WMO Commission for Special Applications of Meteorology and Climatology (CoSAMC) to appoint a rapporteur on urban climatology. The rapporteur was requested, *inter alia*, to review developments in this field since the WHO/WMO Symposium on Urban Climates and Building Climatology (Brussels 1968).

Dr. T. R. Oke (Canada) subsequently agreed to serve as rapporteur and submitted his report to the sixth session of CoSAMC (Bad Homburg, 8-19 October 1973). The Commission agreed that the report provided a valuable reference compendium for workers in urban climatology and therefore recommended that it should be published as a Technical Note.

It is with much pleasure that I express to Dr. T. R. Oke the appreciation of the World Meteorological Organization for the time and effort he has devoted to the preparation of this valuable report.

D. A. Davies

D. A. Davies
Secretary-General

SUMMARY

This publication reviews developments in the field of urban climatology since the Symposium on Urban Climates and Building Climatology held at Brussels in 1968 under the joint sponsorship of the World Health Organization and WMO. During this relatively short period of time the importance of understanding the atmospheric environment of human settlements has become increasingly apparent and the amount of research has greatly expanded.

The review is in two parts. The first part deals mainly with observational studies. This begins with a look at work aimed at elucidating the nature of the fundamental flows of energy and water in the city-atmosphere system, and ends with a summary of the continuing work of measuring the climatological effects of these flows. The second part is concerned with attempts to model the workings of urban climates using physico-mathematical techniques.

The radiation-balance section includes work on the short-wave (global and reflected), long-wave (counter and outgoing) and net all-wave fluxes, and considers both the atmospheric and the surface effects. The water-balance section reviews research into the moisture releases by combustion and the nature of urban precipitation and evaporation. The energy-balance section summarizes investigations into the release of heat by combustion, and the fluxes of sensible, latent and soil (building) heat. The climatological effects section collates the very large amount of literature which is concerned with describing the impact of urbanization on such meteorological elements as air temperature (often the urban heat island), humidity, and wind speed, and considers both the horizontal and the vertical dimensions.

The modelling section deals with both energy and air-circulation models. The similarities and differences among models are reviewed in terms of the objectives of the models and the assumptions and techniques involved in their formulation. Problem areas are examined, and future work is suggested.

In the review of each topic a short introduction is provided which places the recent findings in perspective. Significant new features and suggestions for areas of new research are briefly noted. Where possible, summary tables of typical results are presented. The references cited form a useful bibliography of recent work.

RESUME

La présente publication passe en revue les travaux qui ont été menés dans le domaine de la climatologie urbaine depuis le Colloque sur les climats urbains et la climatologie appliquée à la construction, qui a eu lieu à Bruxelles, en 1968, sous le copatronage de l'Organisation mondiale de la santé et de l'OMM. Pendant cette période relativement courte, il est devenu de plus en plus évident qu'il était important de comprendre comment se comporte le milieu atmosphérique qui entoure les établissements humains. Les recherches se sont donc multipliées de manière considérable.

Cet aperçu des activités se divise en deux parties. La première partie traite principalement des études fondées sur l'observation. Elle débute par un aperçu des travaux visant à élucider la nature de la circulation fondamentale de l'énergie et de l'eau au sein du système ville-atmosphère, et s'achève par une synthèse des travaux actuellement menés pour mesurer les effets climatologiques de cette double circulation. La deuxième partie porte sur les tentatives faites pour établir des modèles des mécanismes des climats urbains en appliquant des méthodes physico-mathématiques.

La section relative au bilan radiatif fait état de travaux sur les flux de courtes longueurs d'onde (rayonnement global et rayonnement réfléchi) et de grandes longueurs d'onde (rayonnement atmosphérique descendant et rayonnement terrestre ascendant) ainsi que sur le flux résultant de tous les rayonnements, et examine les effets de ces rayonnements aussi bien dans l'atmosphère qu'en surface. La section relative au bilan hydrique passe en revue les recherches effectuées sur le dégagement de vapeur d'eau par les combustions et sur la nature des précipitations et de l'évaporation dans les villes. La section relative au bilan énergétique résume les recherches effectuées sur le dégagement de chaleur par les combustions, sur les flux de chaleur sensible et latente, ainsi que sur la chaleur emmagasinée dans le sol (les bâtiments). La section qui traite des effets climatologiques se fonde sur les très nombreux ouvrages qui décrivent l'incidence de l'urbanisation sur des éléments météorologiques tels que la température de l'air (fréquemment sous l'angle de l'îlot de chaleur urbain), l'humidité et la vitesse du vent, en considérant les variations de ces éléments tant dans le plan horizontal que dans le plan vertical.

La section relative aux modèles théoriques traite à la fois des modèles représentant les échanges d'énergie et de ceux décrivant la circulation de l'air. Les similitudes et les différences que présentent les modèles sont examinées en fonction des objectifs visés par ceux-ci et des hypothèses et méthodes qui ont servi à leur élaboration. L'auteur examine enfin les domaines qui posent des problèmes et formule des suggestions quant aux travaux à entreprendre à l'avenir.

Chaque sujet abordé est précédé d'une introduction qui situe les acquisitions récentes dans une perspective appropriée. Il est brièvement fait mention des faits nouveaux significatifs, ainsi que des domaines dans lesquels il serait judicieux d'entamer des recherches. Chaque fois que cela a été possible, les résultats particulièrement typiques ont été présentés sous forme de tableaux récapitulatifs. Les références citées constituent une précieuse bibliographie d'ouvrages récents.

РЕЗЮМЕ

В настоящей публикации делается обзор достижений в области городской климатологии со времени симпозиума по городскому климату и строительной климатологии, состоявшегося в Брюсселе в 1968 г. и совместно субсидируемого Всемирной организацией здравоохранения и ВМО. За этот относительно короткий период времени стала все более очевидной важная роль понимания атмосферной среды, окружающей поселения человека, и значительно расширилось количество научных исследований.

Обзор состоит из двух частей. В первой части рассматриваются, главным образом, исследования результатов наблюдений. Она начинается с обзора работ, цель которых состоит в выяснении природы основных потоков энергии и воды в системе город-атмосфера, и заканчивается резюме ведущихся работ по измерению климатологического влияния этих потоков. Вторая часть посвящена попыткам смоделировать ход городского климата, используя физико-математические методы.

Раздел радиационного баланса включает работы по коротковолновым (глобальные и отраженные), длинноволновым (встречные и уходящие) и эффективным всеволновым потокам, и в нем рассматриваются как атмосферные, так и приземные эффекты. В разделе, посвященном водному балансу, рассматриваются научные исследования в области влаги, выделяемой в результате горения, а также исследования характера городских осадков и испарения. Раздел энергетического баланса содержит резюме исследований выделения тепла при горении и потоков тепла: активного, скрытого и почвы (здания). В разделе, посвященном климатологическому влиянию, сопоставляется большое количество литературы, касающейся описания влияния урбанизации на такие метеорологические элементы, как температура воздуха (часто - остров городского тепла), влажность и скорость ветра, рассматриваемые как в горизонтальной, так и вертикальной плоскостях.

Раздел моделирования касается как энергетических моделей, так и моделей циркуляции воздуха. Аналогия и различия между моделями рассматриваются с точки зрения целей моделирования и предположений и методов, применяемых при составлении моделей. Подробно рассматриваются проблемные области, и делаются предположения о работах на будущее.

Рассмотрению каждого вопроса предшествует краткое введение, в котором указываются перспективы последних достижений. Кратко отмечены важные новые характеристики и предполагаемые области новых исследований. Там, где это возможно, представлены сводные таблицы типичных результатов. Приводимые ссылки составляют полезную библиографию последних работ.

RESUMEN

En esta publicación se examinan los progresos realizados en materia de climatología urbana desde que se celebró en Bruselas, en 1968, bajo el patrocinio conjunto de la Organización Mundial de la Salud y de la Organización Meteorológica Mundial, el Coloquio sobre climas urbanos y climatología de la construcción. Durante este período de tiempo relativamente corto se ha puesto de manifiesto, de manera cada vez más evidente, la importancia de llegar a una comprensión del medio ambiente atmosférico que rodea a los asentamientos humanos, y se han desarrollado de manera notable las investigaciones a ese respecto.

La presente reseña comprende dos partes. La primera versa principalmente sobre los estudios de observación. Estos empiezan por analizar los trabajos destinados a explicar la naturaleza de los flujos fundamentales de la energía y del agua en el sistema atmosférico urbano y terminan por un resumen de los continuos trabajos que se llevan a cabo para medir los efectos climatológicos de esos flujos. La segunda parte se refiere a los esfuerzos desplegados para elaborar modelos de trabajo de los climas urbanos, recurriendo a técnicas físico-matemáticas.

En la sección del balance de la radiación figuran trabajos sobre flujos de onda corta (globales y reflejados), de onda larga (contrarios y salientes) y flujos netos de toda clase de ondas, examinándose tanto los efectos atmosféricos como los de superficie. En la sección del balance hídrico se examinan las investigaciones efectuadas en relación con la humedad liberada por medio de la combustión, la naturaleza de las precipitaciones urbanas y la evaporación. En la sección del balance energético se resumen las investigaciones relativas al calor liberado por la combustión y los flujos del calor sensible, latente y del suelo (construcción). En la sección de los efectos climatológicos se recopila la enorme cantidad de literatura técnica en la que se describen las repercusiones de la urbanización en elementos meteorológicos tales como la temperatura del aire (a menudo, el calor de los islotes urbanos), la humedad y la velocidad del viento, y se estudian asimismo tanto las dimensiones horizontales como las verticales de esos elementos.

La sección de preparación de modelos trata de los modelos energéticos y de los modelos de la circulación del aire. Las analogías y diferencias entre los modelos se examinan en función de los objetivos que se persiguen con esos modelos, y basándose en las hipótesis y técnicas utilizadas para la elaboración de los mismos. Se examinan también los problemas zonales y se sugieren nuevos trabajos para el futuro.

Al estudiar cada uno de esos temas se facilita una breve introducción en la que se da cuenta de los resultados más recientes. También se indican de manera sucinta las nuevas características más significativas así como sugerencias en lo que respecta a los nuevos sectores de investigación. Cada vez que ello ha sido posible, se han facilitado tablas en las que se resumen los resultados más significativos. Las referencias citadas constituyen una bibliografía muy útil de los trabajos más recientes efectuados al respecto.

LIST OF TABLES

LIST OF SYMBOLS

C	water released by combustion
D	diffuse beam short-wave radiation
E	rate of evapotranspiration
I	piped urban water supply
K_H	eddy conductivity
K_M	eddy viscosity
K_W	eddy diffusivity
$K\downarrow$	global radiation (incoming short-wave)
$K\uparrow$	reflected short-wave radiation
L	latent heat of vapourization
$L\downarrow$	counter radiation (incoming long-wave)
$L\uparrow$	long-wave radiation emitted by surface
P	precipitation
Q	surface heat input (total all sources)
Q^*	net surface all-wave radiation flux density
Q_E	latent heat flux density
Q_F	heat flux released by combustion
Q_G	soil (sub-surface) heat flux density
Q_H	sensible heat flux density
R	gas constant

S	direct beam short-wave radiation
T	temperature
V_G	geostrophic wind speed
$\vec{V}$	horizontal vector wind
X	distance downwind of urban/rural boundary
c_p	specific heat of air at constant pressure
Δc	net horizontal atmospheric moisture flux
d	city width (section D(a)); zero plane displacement (section D(c))
e	terrain elevation
f	Coriolis parameter
Δf	net soil water storage
g	acceleration due to gravity
h	height of urban boundary layer
k	von Kármán constant
k_s	soil thermal diffusivity
ℓ	mixing length
p	pressure
q	specific humidity
Δr	net surface run-off
Δs	net atmospheric moisture change
t	time
$\bar{u}$	average horizontal wind speed
u_*	friction velocity

v	vertical wind speed
x	along wind co-ordinate
y	across wind co-ordinate
z	height co-ordinate
z_0	roughness length
α	surface albedo
β	Bowen's ratio
ρ	air density
σ	Stefan-Boltzmann constant
θ	potential temperature
Ω	Earth's angular velocity
ε	surface emissivity
ϕ	population (section D(a)); latitude (section D(c)); non-dimensional wind shear (section E(c))
ζ	vorticity

<u>Subscripts</u>

r	rural
s	surface
u	urban

REVIEW OF URBAN CLIMATOLOGY, 1968-1973

by

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The WMO/WHO Symposium on Urban Climates and Building Climatology held at Brussels in 1968 (WMO, 1970) marked a very important point in the development of urban climatology. Since that time the literature in the field has expanded at an increasingly rapid rate, and in accord with the remarks of Chandler (1970a) in concluding that symposium, the fundamental methodology of the subject appears also to have advanced. To a large extent this expansion and development has parallelled the concern of the general public about the state of the physical environment and man's relation to it.

During this period there have been conferences wholly or partly concerned with features of the urban atmosphere. For example the Symposium on Multiple-Source Urban Diffusion Models held at Research Triangle Park, N.C., U.S.A. in 1969 (Stern, 1970); Conference on Air Pollution

Meteorology held at Raleigh, N.C., U.S.A. in 1971 (American Meteorological Society (AMS, 1971); the WMO-CIB Symposium entitled 'Teaching the Teachers on Building Climatology,' held in Stockholm, Sweden in 1972; the Conference on Urban Environment held at Philadelphia Pa., U.S.A. in 1972 (AMS, 1972); and of course the UN Conference on the Human Environment held in Stockholm, Sweden in 1972. Relevant aspects of these meetings will be cited in the review which follows. Similarly there have been other reviews of a part or the whole field of urban climatology. Most of these reviews have built upon the excellent bibliography compiled by Chandler (1970b) for the WMO Commission for Climatology. The most important of these reviews are those by Peterson (1969), Landsberg (1970a, 1972), Frisken (1972) and Terjung (1973). These surveys were helpful in compiling the present review and this is acknowledged.

This document reviews contributions to the field of urban climatology only, since the WMO Symposium in 1968, and does not attempt to provide a full historical perspective. It also does not substantively apply itself to the fields of urban air pollution or noise pollution; this is because the former is an enormous field largely covered by other WMO Commissions, and because information on the latter is sadly lacking in the meteorological literature.

The review is sequentially organized in the order in which one normally hopes to approach analysis of a boundary layer environment, namely first to assess the information concerning the cascades of energy and mass through the system, and then to evaluate the morphological response that these flows and stores evoke. In the case of the urban atmosphere this means we will look at components of the radiation and water balances, their combination in the surface energy balance, and finally the temperatures, humidities, windspeeds, etc. which are produced. This is then followed by a review of attempts to model the workings of urban climate.

A. RADIATION BALANCE

The surface radiation balance may be represented by:

$$Q^{*} = K\downarrow - K\uparrow + L\downarrow - L\uparrow \qquad (1)$$

where, Q^{*} - net all-wave radiation; $K\downarrow$ - global radiation (where $K\downarrow = S + D$, S,D being direct and diffuse beam short-wave radiation respectively; $K\uparrow$ - reflected short-wave radiation (where $K\uparrow = (K\downarrow)\alpha$ and α is surface albedo); $L\downarrow$ - long-wave radiation emitted by the atmosphere to the surface (counter radiation); $L\uparrow$ - long-wave emmitted by the surface. Of these terms urban/rural differences in K↓ and L↓ are determined by atmospheric properties, whereas K↑ and L↑ are dependent upon surface characteristics. [$L\uparrow = \varepsilon\sigma T_s^4$, where ε - surface emissivity; σ - Stefan-Boltzmann constant; T_s - surface temperature. Clearly ε and T_s are surface properties.] At the time of the WMO Symposium our understanding of these radiation fluxes was almost entirely restricted to short-wave effects including attenuation by pollutants, change in spectral quality, height variations and some aircraft observations of urban albedos. Since that time research projects have started to give a more balanced view of the radiation components, but much remains to be explored.

1. Short-wave

(a) Global radiation (K↓)

The attenuation of short-wave radiation (0.3 - 4μ) by the polluted urban atmosphere has been documented for a number of cities. In general it is known that the depletion is on average 15-20% of that received at a clean rural site (Landsberg, 1970). The attenuation is greatest for heavily polluted atmospheres and when the path length is large. It is common for short-period values to be larger than those quoted above. In addition to gross depletion, the spectral and directional character of the solar beam changes. The tendency is to preferentially filter out the shorter wavelengths (especially ultraviolet), and to increase the proportion arriving as diffuse beam. These changes are important in determining the urban climate, because flux divergence in the atmosphere may infer some degree of heating due to absorption, and depletion reduces the radiation forcing function at the surface. In addition, these changes are important optically, biologically and photo-chemically. For example filtering and scattering are important in visibility, lighting, and the colour perception of distant objects (e.g., Waggoner and Charlson, 1971; Horvath, 1971). Similarly reduction of the shortest wavelengths has implications for tanning, vitamin D production, and skin cancer in humans (e.g., Urbach, 1969; Trosko *et al.*, 1970), and for photosynthesis (e.g., Randerson, 1970). The photochemical reactions occurring in Los Angeles - type smog also involve considerations regarding short-wave character and intensity. Finally it should be noted that the nature of the scattering, reflection and absorption of solar radiation by aerosols is imperfectly understood, and yet is of the utmost importance in global radiation budget considerations as well as those of the urban region.

Since the Brussels Symposium there have been four climatological studies of urban/rural $K\downarrow$ differences. In

Montreal, East (1968) showed an average urban attenuation of 9% with clear skies. He also noted a seasonal trend with monthly averages in winter as high as 15% and summer values as low as 4%, and that this cycle was in-phase with a seasonal air pollution cycle (as given by coefficient of haze). Daily depletions as high as 25% were noted. The addition of cloudy data gave some monthly averages for attenuation as high as 30%, but the reason(s) for this increase were not apparent. In Tokyo, Yamashita (1970) found a general decrease of 10% under clear skies. Additional work showed variations in depletion over Tokyo due to wind direction (and therefore the sector contributing pollutants) and day of the week (due to the cycle of human activities). In Paris, depletion is estimated to be between 6-20% (Dettwiller, 1970c). Most recently Probald (1972) has shown that the average annual depletion in Budapest is 7-8% with winter months again showing maximum values of >15%. All of the above studies confirm the previous estimates of depletion, and add to our statistical base. It should also be noted that Jenkins (1969) reports the frequency of bright sunshine in London has increased since the implementation of air pollution laws. Results of turbidity calculations for Mexico City, and Basle are discussed by Galindo and Muhlia (1970) and Krammer (1970) respectively.

In addition to these climatological studies, there have been short-period measurements of urban $K\downarrow$. Patterson

(1969), Bach and Patterson (1969), and Bach (1970) conducted work in and near Cincinnati. Their instrumentation was crude but indicated an average depletion of ~6% for 6-8 clear days in the summer and autumn. Terjung (1969, 1970a, 1970b, 1971) includes some K↓ measurements and calculations for a few days in Los Angeles. The results are insufficient to draw generalizations, but on one clear, spring day the daily totals showed a 6% spatial variation across the basin. Rouse and McCutcheon (1972) measured K↓ differences between a residential and an industrial site in Hamilton, Ontario with clear skies. Their results show an average difference of 9%. In agreement with other work the attenuation showed a diurnal pattern with proportionately greatest losses at high zenith angles.

The height variation of short-wave attenuation has been studied by Bach (1970-1971a) using Volz sunphotometers in Cincinnati. Observations from roof-level and via a helicopter showed the profile of transmissivity to vary with meteorological and pollution conditions in the manner described previously by McCormick and Baulch (1962) for the same city. Sprigg and Reifsnyder (1972) attempted to measure K↓ divergence over a 60 m layer in Hartford but failed due to instrumental limitations.

Randerson (1970) used a spectroradiometer in and near Houston to investigate attenuation of the solar beam at specific wavelengths. Individual 0.025μ bandwidths could be isolated over the range 0.45-0.7μ. Results for a polluted atmosphere on November 12, 1968 showed an urban attenuation of 23% integrated over the complete range. Individual bandwidth data however, show a steady decrease in attenuation at longer wavelengths (from 31% at 0.45-0.475μ to 15% at 0.675-0.70μ). These results seem to agree with those of Stair (1966) and Nader (1967) for Los Angeles at even shorter wavelengths. In the ultraviolet these investigators showed losses averaging at least 38% and often in excess of 50%, to a maximum of 90%. Shettle (1972) also shows that smog aerosols absorb as well as reflect ultraviolet radiation. Sprigg and Reifsnyder (1972) give attenuation coefficients for Hartford in various bandwidths but feel their data are not able to accurately show spectral differences. Their results do however show total short-wave spectrum attenuation coefficients between 0.05 and 0.40 km^{-1}, and that direct beam (S) losses averaged 20-25% with polluted air. They also note that 75-80% of this direct beam depletion is regained at the surface as diffuse radiation (D). The proportion regained remained constant through the day, whereas the direct beam losses varied with sun angle being greatest with long path lengths in the morning and evening. Idso (1971) further suggests that measurements of D may provide a means of describing particulate air pollution when concentrations of dust are light to moderate.

(b) Short-wave reflection coefficient (α)

The surface reflectivity (here used as synonymous with albedo) is a most important determinant of the surface radiation balance. The modification of natural surface covers through urbanization may be expected to result in albedo changes. The amount of detailed work in evaluating urban albedos is however sparse. Most of the available experimental evidence is presented in Table 1. From Table 1 we note that a wide range of cities show albedos from about 0.10 to 0.30 with a strong preference for values of about 0.15. The highest values are typical of those for deserts and crops, the lowest approach values for water and coniferous forest, and the mean value is close to values for deciduous forests in summer (see SMIC, 1971 pp. 157, 171). The higher values appear to be related to cities in very hot, dry conditions where building materials are commonly lighter in appearance, the only data for a tropical forest city is that for Ibadan (Table 1) and shows a similarity to temperate values. Further comment should await more information. Seasonal effects of foliage change and snow cover are anticipated for higher latitude cities. No information is available on the former, but the studies of Kung *et al.* (1964) and Lindqvist (1968) for Madison, Wisc. and Lund, Sweden respectively, indicate marked urban/rural albedo differences with snow. The enhanced melting due to the heat island and

Table I

Summary of Measured Urban Albedos*

Author	Location	Albedo(α)	Comments
Kung *et al.* (1964)	22 N. American cities (downtown)	0.12-0.23 av. 0.16	Aircraft survey, (July and Sept.)
	22 N. American cities (suburbs)	0.11-0.27 av. 0.16	"
Barry and Chambers (1966)	Southampton, Portsmouth, U.K.	av. 0.17	Aircraft survey, (June and Sept.)
Bray *et al.* (1966)	Bethel, St. Paul, Minn.	0.06-0.09	Helicopter, and photometer
Oguntoyinbo (1970)	Ibadan, Nig.	0.12-0.15	
Bach and Patterson (1969)	Cincinnati, Ohio	0.15-0.16	spot value
Terjung (1971)	Los Angeles, Calif.	0.15-0.17	spatial average
Fuggle (1971)	Montreal, P.Q.	0.15	spot value
Approximate Mean		0.15	

*Snow-free conditions only.

soiling, and the removal of snow quickly cause urban albedos to drop below those of the surrounding countryside. In absolute energy terms this albedo difference may not lead to large differences of absorbed radiation, mainly because snow is commonly associated with seasonally low sun angles.

Clearly there is a need for accurate information concerning urban albedos. A number of features of the urban environment can be invoked to explain <u>either</u> higher or lower albedo values than for rural surfaces (e.g., trapping between building elements, or shading, can be used to explain a decrease; removal of vegetation and use of light coloured building materials to explain an increase). Care should be taken not to oversimplify the problem by only considering one effect in isolation (e.g., Terjung 1970a only quotes the factors tending to increase albedo). Controlled experiments and realistic models are required if progress is to be made. Craig and Lowry (1972) have provided a useful start in our consideration of trapping within the urban 'canopy.' Their numerical model results indicate multiple reflection will tend to decrease albedos, whether this is the dominant control only further careful work will reveal.

2. Long-wave

(a) Counter radiation (L↓)

It is widely held that urban areas experience increased L↓ due to atmospheric pollution. It is hypothesized that the aerosol blanket over the city absorbs L↑ from the surface by night, and L↑ and K↓ by day, and then re-radiates this heat with a large proportion arriving as L↓. This process was often quoted as one of the agents in the production of the urban heat island. However, on closer inspection it is impossible to find observational verification of this effect prior to the Brussels Symposium. Since that time a few pertinent studies have been published. In general they confirm that urban areas receive increased L↓, but the magnitude of the energy exchange involved appears to be small. In this regard care should be exercised in comparing urban/rural long-wave radiation differences. Use of percentages can be misleading unless accompanied by absolute energy values. For example in winter and at night with low temperatures, a high percentage difference in radiation may well represent a smaller energy difference than a smaller percentage by day or in summer.

In their Cincinnati study Patterson (1969) and Bach and Patterson (1969) calculated urban/rural L↓ differences. The differences were small and within the instrumental accuracy and hence little confidence can be placed upon them. Terjung (1970a, 1971) measured radiant sky temperatures, which by assuming a value for sky emissivity can be used to approximate L↓. Small intraurban variations are evident but no rural values are available for comparison and much of the data is complicated by cloud. Oke and Fuggle (1972) conducted an investigation of L↓ in Montreal at night. An infra-red pyrgeometer was traversed across the city utilizing an elevated freeway. Results showed an increase of L↓ in the urban area, which in cross-section closely parallelled the form of the urban heat island. The rural/urban energy difference was usually < 5% and in absolute units always <40 W m^{-2} (usually <20 W m^{-2}). In the absence of simultaneous temperature, water vapour and pollution concentration profiles it was impossible to isolate the exact cause of the increase, but from simple considerations it appeared that urban warmth from other processes was sufficient to account for the greater atmospheric emission. Hence the >L↓ may be an effect rather than a cause of the heat island. Observations were able to identify the contributions to L↓ from an oil refinery complex, and from the elevated urban 'heat plume' (see section D) in the downwind rural area. Rouse and McCutcheon (1972) report a similarly small nocturnal increase in L↓ at their industrial site in Hamilton. By day however, they found the increase in urban L↓ to be much larger, peaking at midday when the average increase compared to the residential site was 33%. They note that total incoming radiation ($K\downarrow + L\downarrow$) at the sites are almost identical through the day, thus the depletion of $K\downarrow$ is exactly balanced by the increase in $L\downarrow$. The authors ascribe the greater $L\downarrow$ to heating of the polluted atmosphere by absorption of $K\downarrow$. No additional measurements were available to check this. Short-period observations by Landsberg and Maisel (1972) and Maisel (1972), in and near Columbia, Md. also show virtually zero urban/rural $L\downarrow$ differences. Finally Probald (1972) shows that measurements of $L\downarrow$ in- and out-side Budapest give the city < 1% extra on an annual basis.

Jacobs (1971) took traverse measurements of incident long-wave at street-level in Minneapolis-St. Paul. The data therefore refer mainly to emission of long-wave radiation by buildings and other obstructions, and to a lesser degree from the sky. Comparison with air temperatures measured by car traverses showed agreement between warm areas and increased building density.

(b) Surface long-wave emission ($L\uparrow$) and remote sensing

Direct measurements of $L\uparrow$ from the urban surface are almost completely absent. Indirect assessments are based on measurements of effective surface radiant temperature (assuming emissivity = 1), or some other calculation procedure, or leaving this term as a residual having measured or calculated the other terms in equation (1). Simple considerations concerning typical urban/rural ε values, and typical urban heat island magnitudes suggest that L↑ losses should be greater from the urban area. Such observations as are available to date generally confirm this analysis.

Bach and Patterson (1969) show small urban/rural L↑ differences in the early morning in Cincinnati, increasing to ~15% greater loss by the city at midday. This represented an average extra loss of ~70 W m^{-2} over six days in the period July to October, 1968. Landsberg and Maisel (1972) also show the urban L↑ excess to be greatest (21%) at midday (representing a relative energy loss of ~130 W m^{-2} on a near-equinoctial day). Their results are for a weed field (rural) and parking lot (urban). At night differences were 11-12% (~50 W m^{-2}). The results of Oke and Fuggle (1972) are restricted to conditions around midnight in Montreal. Assuming both urban and rural surface emissivities equalled unity, urban/rural L↑ differences always showed the urban area to exhibit a greater loss, which averaged ~5% (~20 W m^{-2}) with a maximum deficit of ~40 W m^{-2}. Daytime observations by Terjung (1971) under complete overcast do not allow isolation of urban/rural influences, but intra-urban spatial patterns were noted to show greatest L↑ in areas where D attenuation was least, and hence absorption was expected to be greatest.

The application of the Stefan-Boltzmann equation to the measurement of surface temperature via the infra-red radiation thermometer or scanner has great promise for remote sensing in urban climatology. Examples of ground-based uses of this technique are provided by the work of Landsberg and Maisel (1972), Terjung (1971), and Myrup and Morgan (1972). Aircraft studies include those of Fujita, Baralt and Tsuchiya (1968), Fagerlund *et al.* (1970), Lindqvist (1970), RPU (1972), and Holmes and Wright (1972). Most recently the use of infra-red radiometers from satellites has been shown capable of identifying urban heat islands (Rao, 1972). The applications promise well for urban climatology where the problems of physical size, and spatial integration of a myriad of individual elements are most formidable. Before the technique is fully utilized however, it will be necessary to examine the patterns of surface emissivity variation that are characteristic of urban areas. In rural areas Holmes (1972) estimates the error of failing to account for ε variations to be to ± 1.5°K. Initial considerations based on available tables of ε would suggest urban errors to be greater than this.

3. Radiation balances

A definitive study of the surface net all-wave radiation balance (Q^*) of an urban area is not at present available. However, on the basis of the material presented in the foregoing discussion it is clear that the component short- and long-wave radiation fluxes, and their related surface radiative properties, all experience a change when natural surfaces are converted to urban land-uses. From a climatic standpoint we are most concerned with the net result of these changes since that determines the available energy for heating the atmosphere, the urban fabric and evaporating water. Without wishing to enter into too much speculation we may however summarize the observed tendencies in the previous sections and move toward a qualitative statement concerning urban/rural Q^* differences.

It is clear that $K\downarrow$ input to the city is lower than to the surrounding countryside. Depending upon such factors as season, and the nature of the rural vegetative region, it appears that urban albedos are generally decreased. These changes therefore operate against each other, so that urban/rural net short-wave differences may be reasonably small. For example, an urban decrease of 10% in $K\downarrow$, and of 5% in α results in only about a 4% urban decrease of ($K\downarrow - K\uparrow$). This seems to be in agreement with the results of Probald (1972) for the annual balance in Budapest. Similarly we note that the long-wave changes tend to offset each other since both the input ($L\downarrow$) and output ($L\uparrow$) are increased by the city. From the work of Bach and Patterson (1969), Oke and Fuggle (1972), Landsberg and Maisel (1972) and Probald (1972) it appears that the increase in $L\uparrow$ is the greater, and hence ($L\downarrow - L\uparrow$) may be expected to be a larger loss for the city as compared to the country. Detailed comparison of urban/rural emissivity differences are needed to verify this conclusion.

Combining the net short- and long-wave terms we might therefore expect the city to generally show lower Q^* gains by day, and slightly greater losses by night. Again this conclusion is largely verified by available observations or calculations given by Golubova (1969), Oke and Fuggle (1972), Landsberg and Maisel (1972) and Probald (1972). One discrepancy is the early morning value of Bach and Patterson (1969) which shows a greater urban Q^* which they attribute to greater absorption by vertical structures at low sun angles.

The previous discussion has referred almost entirely to elements of the surface radiative balance. Fuggle (1971, 1972) however has attempted to monitor the nocturnal atmospheric radiative balance by a flux divergence experiment in Montreal modelled after the rural work of Funk (1960). He noted that radiative cooling rates were commonly three times as great as actual cooling, from which it was concluded that nocturnal cooling of the lower urban atmosphere is driven by radiation, with turbulent heat transfer acting as a retarding influence. These results are also important in showing that flux divergence must be expected in urban atmospheres, a conclusion which considerably complicates theoretical and observational approaches.

B. WATER BALANCE

The surface water balance of an urban area may be represented by:

$$P + I + C = E + \Delta r + \Delta f \tag{2}$$

where, P - precipitation; I - imported water supply; C - water released by anthropogenic activities; E - evapotranspiration or condensation; Δr - net run-off; and Δf - net storage. The process of urbanization changes the relative roles of all of the components in equation (2) vis á vis an undisturbed rural situation. Some of these changes cause the city to become a relative moisture source, and some a sink. A very generalized scheme of measured or anticipated hydrologic changes are shown in Table 2. The degree of certainty to be accorded some of the indications in this table is not great. This is again due either to the lack of research, or to genuine uncertainties in natural phenomena.

Table 2

Measured or Anticipated Surface Hydrologic Changes

Due to Urbanization

Element	Comparison with Rural Environment	Remarks
P	more?	Thermal and mechanical uplift, nuclei, combustion
I	more	Piped water supply
C	more	No rural counterpart
E	less?	Reduction of evapotranspiring surfaces
Δr	more	Lower permeability and channelling
Δf	less	Poor interception and infiltration

From the climatological point of view the most important terms are P and E. The precipitation as the main input (usually) to the hydrologic cascade is the forcing function which largely characterizes the hydrologic activity of the system. Evaporation is especially important because the energy release or uptake associated with water phase change is often large. This term also links the hydrologic and solar energy cascades (and thereby the water and energy balances), and helps determine the level of atmospheric moisture (humidity). It is most unfortunate therefore to note from Table 2 (and as will emerge from the review which follows) that the exact response of these two terms remains in some doubt.

The terms I and Δr and to a lesser extent Δf in equation (2) are well covered in the engineering and hydrological literature, including useful summaries by Leopold (1968), Lull and Sopper (1969), Moore and Morgan (1969) and Schaake (1972). This literature deals almost exclusively with the surface and sub-surface hydrological aspects of urban water. It begins to document the observed increases in total discharge, peak flow, overland flow and sediment load, and the decreases in flood response time, and water quality which accompany urbanization of a watershed. This information is of value to city engineers, architects, planners, etc. but does little to satisfy the climatologists' interests in city/atmosphere interaction. Here we will consider information available concerning the effects of cities upon precipitation (type and amount), interception and storage of precipitation, and release of water vapour through combustion. Implications for urban evapotranspiration will be mentioned but taken up in more detail in the energy balance section.

(a) Precipitation (P)

At the time of the Brussels Symposium there was some difference of opinion regarding the reality of urban-induced precipitation. Since then carefully conducted experiments have dispelled some of this scepticism. In particular one must single out Project METROMEX (Changnon, Huff and Semonin, 1971; Changnon, Semonin and Lowry, 1972) in St. Louis as having made important contributions to our knowledge. Considerable research remains to be done especially regarding the physical mechanisms underlying the climatological effects.

Prior to METROMEX Changnon (1968) reported results of a study concerning the effects of the Chicago-Gary urban-industrial complex on precipitation in the downwind region, especially at LaPorte, Ind. The anomalously high precipitation, thunderstorm and hail occurrences sparked some controversy as to their validity (Ogden, 1969; Holzman and Thom, 1970; Changnon, 1970; Changnon, 1971; Holzman, 1971a, 1971b; Hidore, 1971; Ashby and Fritts, 1972). Following this Landsberg (1972) reviewed the evidence of these and other studies and concludes that the case is a reasonable one but remains not completely proven. Further research, preparatory to METROMEX, was undertaken in eight U.S. cities chosen to give a variety of city sizes, functions and in different climates. Results showed increases in rain and convective storm activity in six of the eight cities (Huff, Changnon and Lewis, 1971; Huff and Changnon, 1972a; Changnon, 1972). Table 3 gives some representative values of general increases in precipitation derived from climatological analyses. Most of the increases agree with the average estimates of Landsberg (1970a) but La Porte clearly is in a class by itself. Thunder and hail frequencies were noted to require a critical size of urban area (>0.9 million population) before modification was noted, and that at a population of three million excess thunder frequencies are likely to be about 30% (Changnon, 1972). Maximum effects of thunder were usually found in the city, but maximum hail was found 24-40 km downwind. Changnon believes this to be a logical convective sequence for an urban area. The sequence would be increased convection over the warmer rougher city leading to thunderstorms, and the evolution of these storms into hailstorms downwind. Important urban-related increases were noted in cities whose industrial-activity is very different (e.g., Washington, D.C., non-industrial; Chicago, industrial). Where diurnal thunderstorm frequencies were noted, the increases were at night when the heat island is usually best developed and when rural areas would be expected to show minimum convective activity. Frequency increases were also noted to show a better relationship to population and heat emission in the city than to particulate emissions. These three points

Table 3

Summary of Percentage Increases in Urban Precipitation and Related Phenomena from Climatological Analyses*

	Total Rainfall			Thunderstorm Days		Hail Days	
City	Annual	Warmer half year	Colder half year	Annual	Warmer half year	Annual	Warmer half year
Champaign-Urbana	5	4	8	7	17	-	-
Chicago	5	4	6	6	13	-	-
La Porte	31	30	33	38	63	246	-
New York	16	12	20	-	-	-	-
St. Louis	7	10	-	11	21	-	35
Tulsa	8	5	11	-	-	-	-
Washington	7	6	9	-	-	-	-

*Source: Changnon (1969, 1971), Huff and Changnon (1972b).

(lack of correlation with industrial function; nocturnal frequency increase; relation to population and heat emissions) lead Changnon (1972) to conclude that thermal and mechanical turbulence induced by the city are much more important than the extra provision of condensation and freezing nuclei by urban aerosols.

On the basis of the above studies, St. Louis was chosen for project METROMEX, and soon will also be the 'model' city used in the 5-year Regional Air Pollution Study to be mounted by the U.S. Environmental Protection Agency. The METROMEX results from the studies conducted by a number of organizationsare just starting to be published and only a few of the findings can be reported here. In general it appears that the previous findings are being substantiated in St. Louis. Changnon (1972) reports greater summer hail and rain intensity in the downwind region. Beebe and Morgan (1972) conducted synoptic analyses on the 1971 St. Louis rainfall data, confirming the previous urban enhancement results, and showing that this is most likely to occur in the warm sectors ahead of frontal zones. This is related to maximization of downwind rain increases at times of instability with large mixing depths and free availability of heat, water vàpour and aerosols from the urban boundary. There was also evidence of local energy sources leading to intensification. This uptake of effluent by convective storms was also studied by Semonin (1972) utilizing the release of lithium as a tracer chemical. The study revealed rapid ingestion and rainout by the storm. Schickedanz (1972) has defined a raincell as a precipitation entity whose morphology can be described. Analysis of the raincell's characteristics can be used to show the effect of the urban area. Raincell analysis in St. Louis showed urban influences on the intensity, duration, velocity and area of precipitating clouds. In a study of pollutant scavenging in St. Louis, Gatz (1972) finds no difference in washout ratios across the metropolitan area, but concentrations in both air and water were higher in the downwind region.

In studies other than METROMEX the picture is less clear. An analysis of precipitation records for Central Park in New York showed a decrease over the period 1927-1965 (Spar and Ronberg, 1968). But to illustrate the meso-scale complexity of the problem of assessing precipitation trends, one should note that the nearest station to Central Park showed a slight increase over the same period. Barr *et al.* (1972)and Eagleman, Huckaby and Lin (1972) working in the Kansas City-Topeka Corridor find increases in monthly rainfall, especially in the winter months, and in the largest cities. In the summer there are indications of a reversal with the rural stations receiving more rain. A very recent study from an urban-industrial complex in Bombay, India clearly shows an increase in precipitation of 15% in the downwind region.

In a series of papers Atkinson (1968, 1969, 1971) has examined case studies of thunderstorms as they move across London, U.K. He argues the importance of this approach compared to purely climatological, or purely process-oriented approaches employed in other studies of urban precipitation enhancement. Atkinson (1971) feels that the urban effect can be important if conditions are favourable. One case study shows intensification of a thunderstorm and subsequent increase in rainfall, because of incorporation by the storm of warm humid air lying over central London.

Relatively little work has been conducted on the effects of urban areas on snow. Changnon (1969) attributes this to the difficulties of obtaining accurate long-term measurements. Changnon reports a small study for Champaign-Urbana showing a 12% increase in amount, and an 8% increase in days with snow. Recently Agee (1971) presents evidence that in one snowstorm in West Lafayette, Ind. there were areas of locally-induced snow. The areas were found to lie downwind of power plants and factories whose effluent was postulated to provide inadvertent seeding of a super-cooled cloud. These results are, however, at variance with previous experience (Potter, 1961; Woollum and Canfield, 1968), which suggest the heat island influence may reduce measured snowfall.

A meso-scale study by Grillo and Spar (1971) tends to support the urban decrease effect.

Lawrence (1971a) reviews previous information in the United Kingdom concerning the relationship between day of the week and rainfall. He notes that recent support for a pattern of weekdays being wetter than weekends comes from Craddock (1968), Gold (1968) and Scorer (1969), and his own analysis of data from the London area lends further evidence. Similar information is now available for Paris (Dettwiller, 1970a), for St. Louis (Changnon *et al.*, 1971; Huff and Changnon, 1972b), and for the Kansas-Topeka Corridor (Eagleman *et al.*, 1972). As Lawrence (1971a) points out the establishment of the validity of a weekly cycle links local rainfall anomalies in urban areas to man-induced influences, rather than physiographic or water influences, which would operate every day. Lawrence further postulates that the weekly cycle of human activities causes weekly cycles of pollution and temperature which lead to a weekly cycle of rainfall. Hence he suggests the coexistence of a heat island and a 'rain island.'

(b) Moisture release through combustion (C)

As Hanna and Swisher (1971) point out, nearly every activity of man involves the generation of some heat and moisture as by-products. The burning of fossil fuels such as natural gas, gasoline and fuel oil involves the simultaneous release of considerable amounts of water vapour. Similarly the use of water to absorb "waste" thermal energy from power plants or industrial processes inevitably results in greatly enhanced vapourization from cooling towers, cooling ponds, rivers, lakes and oceans. Since these domestic and industrial moisture sources are often located in or near urban areas they become of importance in urban energy, humidity, cloud, precipitation and visibility considerations.

Most of the detailed work to date has been directed towards the local effects of cooling towers in producing fogging, precipitation, and possible plume interactions (e.g., Visbisky, Bierman and Bitting, 1970; Aynsley, 1970; Hewson, 1970), but relatively little is available concerning meso-scale effects. One exception is the interesting study of Hage (1972a) into the effects of vapour released by combustion, upon the production of low-temperature fog in Edmonton. Contributions to the urban vapour content were computed from fuel consumption figures. At an ambient temperature of -35°C the total daily water output was 2.8×10^{10}g, of which 17.6×19^{9}g was from natural gas, 9.8×10^{9}g from cooling towers and 0.7×10^{9}g from motor vehicles. This represented a seven-fold increase in <20 years, during which time the city's population more than doubled. Conservation equations were used to estimate fog layer depths and excess water contents as air moved across the city following a basic framework proposed by Summers (1964) for heat (see section D(a)).

(c) Evaporation (E)

The water balance (equation 2) potentially provides a means of obtaining evaporation from the city. Unfortunately it appears that for other than gross annual figures this is not practical. The many reasons for this include the inability to monitor or calculate the other terms in equation (2) with sufficient precision in space and/or time, problems of ungauged flows, building interception storage, complication of discharge measurements by suspended sediments, deep basin inflow, etc.

Attempts to estimate approximate evapotranspiration for urban areas have been made by Muller (1968), Lull and Sopper (1969) and Lacy (1972, personal communication). Muller estimated that if half of a basin in New Jersey were urbanized, E would decrease by 50%. Lull and Sopper estimated annual potential evapotranspiration would be reduced by 19, 38 and 59% if a forested watershed were converted to 25, 50 and 75% impervious cover respectively. Lacy feels that interception and storage of precipitation by building materials may be an important source of vapour, in support he quotes a study by Watkins (1963) which includes the fact that in an area of London which was 95% paved, only 50% of the rainfall was recorded as run-off. Givoni (1969) also suggests that building materials absorb "appreciable quantities of water." Further discussion of evaporation is left until section C(b). We may conclude that the urban water balance is imperfectly understood and requires increased attention.

C. ENERGY BALANCE

The surface energy balance of an urban area may be represented in its simplest form by:

$$Q^{*} + Q_F = Q_H + Q_E + Q_G \qquad (3)$$

where, Q_F - artificial or anthropogenic heat released due to combustion and metabolism; Q_H - sensible heat flux density; Q_E - latent heat flux density, (where, $Q_E = L(E)$, and L - latent heat of vapourization, and E - rate of evapotranspiration); Q_G - sub-surface (soil, building, road, etc.) heat flux density. This treatment ignores the effects of advection, energy transfer by precipitation, friction, photosynthesis, and snow-melt.

The importance of employing the energy balance framework in urban climatology was first noted by Munn (1966). At the time of the Brussels Symposium the truth of this approach was evident but practical problems hindered its implementation (Fuggle and Oke, 1970). Since then the first pioneering urban energy balance observations have begun to appear. Some of the studies have been crudely instrumented, and gross assumptions employed, but the approximate size and probable direction of the fluxes in equation (3) are emerging. The review which follows attempts to draw some of the scattered pieces of evidence together, and to assess its value in increasing our knowledge of the energetic basis of urban climate. The lack of this information has greatly hindered the development of a rational understanding of urban climates, because our interpretation of climatic effects has been based on pure speculation as to the causative processes. For example, the heat island has been one of the best described but least understood climatic phenomena. Similarly accurate observational data is required to provide realistic input and test criteria for modelling (Oke and Fuggle, 1972).

We may reasonably expect that urbanization will markedly alter some, or all, of the terms in equation (3). The 'natural' (rural) partitioning of energy receipts and losses is disturbed by changes in the thermal (including radiative), moisture and roughness characteristics of the urban interface.

Without wishing to pre-judge their validity, we will note here the changes in energy exchange that have been traditionally anticipated. From previous considerations (section A 3) Q^* is expected to show a slight deficit by comparison to the rural area. The radiative energy source is however supplemented by Q_F in the city due to anthropogenic releases. It is anticipated that the net effect of changes in the urban water balance (section B) is to reduce the amount of water freely available for Q_E. This is then expected to proportionately increase the energy available to Q_H and Q_G. The former being enhanced by the increased roughness and mixing, and the latter by the increased thermal capacity of building materials. The preceding generalities are assumed to apply to 'the city surface' often taken to be some datum at about roof-level. Below this level, in between the roughness elements where most activities take place, the situation may be expected to be so complex as to be almost undefined. The radiative trapping, multiple reflection and emission, and the storage in buildings may possibly be amenable to analysis, but the turbulent exchanges may be beyond detailed treatment. The coupling between the micro-scale street-level environment, and the meso-scale urban boundary layer, remains an area for research.

(a) Anthropogenic heat production (Q_F)

Recent reviews of this term have been written by Weinberg and Hammond (1970), Hanna and Swisher (1971) and SMIC (1971). Table 4 provides a summary of estimates of Q_F for some European and North American cities. No readily available estimates for low latitude cities could be found. Clearly in some high latitude (Fairbanks), densely populated (Manhattan) and industrialised (Moscow) cities the annual Q_F value can be ≥ annual Q^*. In the cold season the Q_F value may well dominate as the energy source, and must be accounted for in energy balance considerations. In summer it may be insignificant by comparison with solar energy. Computations based on theoretical models and empirical data confirm that during the winter in high latitudes a major part of the urban heat island can be accounted for on the basis of anthropogenic releases (Summers, 1964; Leahey, 1969; Leahey and Friend, 1971; Pasquill, 1970a; Oke and East, 1971; McElroy, 1971; Hage, 1972a). On an annual basis however East (1971) suggests that artificial heat is of much lesser importance. Ewing (1972) argues that the increase in Q_F with increasing energy consumption, will raise the mixing depth over major cities, and hence improve air quality. In the reviewer's opinion, the problem requires a more multivariate analysis than this provides.

Table 4

Anthropogenic Heat Production (Q_F) in European and North American Cities

City	Area (km^2)	Population ($\times 10^6$)	Averaging Period	Q_F ($W\ m^{-2}$)	Average Q^* ($W\ m^{-2}$)	Author
Budapest (1970)	113*	>1.0	Year Summer Winter	43 32 51		Probald (1972)
Hamburg	747	1.8	Year	13*	55	SMIC (1971)
Moscow	878	6.4	Year	127	42	SMIC (1971)
Sheffield (1952)	48	0.5	Year	19	56	Garnett & Bach (1965)
West Berlin	234*	2.3	Year	21	57	SMIC (1971)
Cincinnati (1963)	200*	0.5	Summer	26		Bach (1970)
Fairbanks	37*	0.03	Year	19	18	SMIC (1971)

CONTINUED

*Building area only.

NOTE: Some of the SMIC (1971) Table 4.2 values appear to be in error and have been corrected here. These include values for Q^* in Sheffield, Q_F for Manhattan, and the averaging period for Cincinnati.

Table 4 (Continued)

City	Area (km^2)	Population ($\times 10^6$)	Averaging Period	Q_F ($W\ m^{-2}$)	Average Q^* ($W\ m^{-2}$)	Author
Los Angeles	3,500*	7.0	Year	21	108	SMIC (1971)
Manhattan, N.Y.	59	1.7	Summer	40		Bornstein (1968)
			Winter	198		
Montreal (1961)	78*		Year	99		Summers (1964) and Oke (1969)
			Summer	57		
			Winter	153		
Vancouver (1970)	112*	0.6	Year	19		Yap (1973)
			Summer	15		
			Winter	23		

For detailed intra-urban studies better space and time resolution is required. Certain locations may possess anomalously high Q_F fluxes [e.g., due to steelmills (Oke and Hannell, 1970) or air conditioners (Federer, 1971)], and this must be recognized. Attempts to estimate diurnal variations of Q_F have been described by Turner (1968) and Bach (1970). It is often difficult to know the pathway by which Q_F escapes to the atmosphere. The energy may be radiated, conducted or convected outwards from the urban fabric. Direct measurement of the other terms in equation (3) will therefore include this component.

(b) Sensible and latent heat (Q_H, Q_E)

The nature of the urban/atmosphere interface is a major obstacle in providing observational evidence of the nature of the turbulent fluxes Q_H and Q_E. The effective surface or urban 'canopy' consists of an exceptionally rough, and multifaceted set of climatically dissimilar surface types. These exchange surfaces provide sources and sinks for energy and mass that are non-uniformly distributed in the horizontal and vertical. Hence, within the urban 'canopy' flux divergence is likely to be the norm, and micro-scale advective effects will be continually generated and dissipated. The roughness also causes the gradients of such elements as temperature and humidity to be very small and hence sophistication of measurement is required. These conditions pose such massive problems to the implementation of most flux determination techniques that it is not surprising that urban energy balance modelling (see section E) is more popular than energy balance measurement. Nevertheless, three main classes of approach have been used with limited success and will be reviewed here.

Firstly, there have been climatological approaches. Probald (1972) estimated monthly values of Q_H and Q_E for Budapest. Full details of the analysis are not available but it appears that some arbitrary apportionment of the sum $(Q_H + Q_E)$ was made to Q_E on the basis of the percentage of central Budapest having evaporative surfaces, and handbook values of Q_E for rural conditions. On this basis the monthly urban/rural Q_H differences show the city value to be greater all year long. This excess sensible heat was postulated to be the result of a smaller urban Q_E in the summer, and the greater release of Q_F in the winter. Hence it is suggested that Q_H is directed into the atmosphere throughout the year, unlike the rural surroundings where Q_H is towards the surface in winter. In Topeka and Kansas City Eagleman *et al.* (1972) computed values of potential evapotranspiration via climatological methods although their data were short-period. Not unexpectedly results indicated increased rates in the city centre. The exact physical meaning of such values from street-level traverses is not totally clear however.

Secondly, there are studies, mainly by Terjung and collaborators concerning short-period spatial variations and averages of derived fluxes in Los Angeles. In an initial study an area of downtown Los Angeles was surveyed (Terjung, 1970a). Some radiant fluxes were measured but the other fluxes were estimated on the assumption that $Q_E = 0$ over a 'dry' city. Later metropolitan Los Angeles was studied during one September day with complete stratus overcast (Terjung *et al.*, 1970b). Q^* and Q_G were measured, but again it was assumed $Q_E = 0$ because of a long dry period preceding the observations. This may be true for paved areas but unlikely to hold for the city as a whole. In each of these studies Q_H was estimated to be ~0.8 Q^*, but lack of measurements to apportion the turbulent fluxes leaves room for doubt. In another study one cloudy (frontal passage), and one clear day's data were compared. The Los Angeles basin was sampled from seventeen paved-surface stations visited by two cars. Q^* and Q_G were measured and for the cloudy day (wet surface) a Bowen ratio approach to partitioning Q_H and Q_E was used. Surface temperature and vapour pressure were computed from radiometer measurements (assuming $\varepsilon = 1$, and that the surface vapour pressure equalled the saturation vapour pressure at the surface temperature). Atmospheric temperature and vapour pressure were measured by Assmann psychrometer at 1 m. Myrup and Morgan (1972) have found difficulties using almost the same approach in Sacramento. This is probably due to the accuracy required in estimating gradients (see Fuchs and Tanner, 1970) which is not attainable with these instruments, and compounded by the use of different instruments for the two levels in Terjung's case. For the clear day $Q_E = 0$ was again assumed. The resulting fluxes for the cloudy day showed very small values (within instrument error). With clear skies Q_H was estimated to be ~0.6 Q^* at midday.

Thirdly, there have been a few point studies attempting to estimate turbulent transfer. Such studies depend strongly upon the site chosen as to the representativeness of their values. Early work by Bach and Patterson (1969) and Bach (1970) in Cincinnati attempted to calculate the turbulent fluxes using the aerodynamic approach, but they concluded that their instrumentation was too crude to separate the individual fluxes. The aerodynamic, Bowen ratio, and eddy correlation approaches to determination of eddy fluxes have been tested and used in Montreal, P.Q. and Vancouver, B.C. The philosophy and limitations of these studies are given by Oke (1969), Fuggle and Oke (1970), Fuggle (1971) and Oke, Yap and Fuggle (1972), and results are presented in Fuggle (1971), Oke (1972a), Oke, Maxwell and Yap (1972) and Yap (1973). Measurements were conducted from 1 to 20 m above roof-level, in areas of relatively uniform land uses near the downtown area, mostly during the summer. The initial studies in Montreal showed the aerodynamic approach to be inapplicable to the urban area. The Bowen ratio ($\beta = Q_H/Q_E$) showed a recognizable diurnal pattern. All β values were positive reaching a peak of ~1.2 at midday. This indicates that for Montreal Q_E can be a significant energy sink. Pilot eddy correlation results showed promise, and subsequently yaw sphere-thermometer systems (Tanner and Thurtell, 1970) were constructed, tested (Yap *et al.*, 1972), and used at urban and rural sites in Vancouver. Measurement of Q^*, Q_H and Q_G directly, left Q_E as a residual in equation (3) [since Q_F was small in the summer (see Table 4)]. Results at 1 m above the roof showed urban/rural Q^* differences to be small, and Q_H and Q_G to be slightly larger by day in the city. The net result was to reduce Q_E in the city but this term remained a significant energy sink. Near the roof-top typical midday values were $Q^* = 600$ W m^{-2}, $Q_H = 300$ W m^{-2}, $Q_G = 100$ W m^{-2} leaving $Q_E = 200$ W m^{-2}. At 20 m above roof-top level (thought to more nearly approximate the surrounding urban area), results indicated almost equal partitioning of Q_H and Q_E giving $\beta \sim 1$ at midday, in agreement with the Montreal results. Following rain, Q_E was noted to become the main energy dissipation term, followed by a recovery of Q_H in a few days. At night Q_H was consistently observed to be directed upwards, i.e., the surface remains an energy source. In another study Landsberg and Maisel (1972) quote turbulent flux values for a rural weed field and a parking lot in Columbia, Md. Unfortunately no methods or assumptions are given. Planned studies in Sweden and Russia aim to contribute to this research area in the future (Miller, 1970; Högström, 1972).

(c) Heat Storage (Q_G)

The thermal properties of the urban fabric (buildings, pavements, etc.) are conducive to conduction and storage of absorbed energy. Especially in summer the nocturnal release of the energy gained by day is often cited as a major contributor to the urban heat island (e.g., Peterson, 1969; Landsberg and Maisel, 1972). Again, however, little observational evidence is available to substantiate or refute this proposition.

A number of investigators have attempted to approximate the urban storage value by embedding heat flux plates in the ground or beneath the roof surface (Höglund, Mitalas and Stephenson, 1967; Terjung and collaborators, 1970b, 1971; Borgel, 1972; Myrup and Morgan, 1972; Oke, Yap and Fuggle, 1972; Oke, Maxwell and Yap, 1972; Yap, 1973). The surface types surveyed include concrete, tarmacadam, tarpaper, tar-and-gravel and cellular concrete. For clear weather conditions in summer, the largest daytime fluxes are ~170 W m^{-2} into the tarmacadam, and the lowest ~10 W m^{-2} into the cellular concrete. The tarmac Q_G values represented ~0.35 Q^* at midday in the study by Terjung and collaborators (1971) and ~0.28 Q^* in that by Yap (1973), and represent close to the upper limit to be anticipated. Surfaces with higher albedos, with unfavourable slope or aspect, and with vegetation are certain to produce a spatially-integrated 'urban storage' value well below this. The only nocturnal values are for the tar-and-gravel roof, and show a relatively constant upflux of 30-40 W m^{-2} on clear nights following clear days (Oke, Maxwell and Yap, 1972; Yap, 1973). Such a flow is capable of supporting a small heat island but is probably smaller than previously anticipated.

Indirect approaches to the estimation of Q_G have also been attempted. Probald's (1972) climatological analysis does not state how Q_G was derived, but he states that it is an insignificant term on a monthly and annual basis, but important diurnally. This is to be expected from simple conservation principles. In a different approach Oke, Maxwell and Yap (1972) sought to solve a theoretical cooling rate equation for the urban thermal admittance. But cooling rate data from Montréal and Vancouver did not conform to the rural model, and hence suggested the operation of unique urban process controls at night.

In summary, it appears that this important urban energy term is poorly researched. Future studies must provide information for other than isolated, horizontal surfaces, and seek means of providing spatially-integrated values. It would also be interesting to study the flow of energy from the building interior to exterior since this opposes the normal flow by day and enhances the upflux by night. Finally, it may be useful to calculate heat storage in the relatively stagnant air and vapour below roof-level. Analogies with forest canopy storage could be drawn.

D. CLIMATOLOGICAL EFFECTS

Traditionally studies of the effects of cities upon the weather elements form the largest portion of the urban climatological literature. As can be seen from the foregoing review, studies of process are presently gaining impetus, but in sheer volume morphological descriptions of the features of urban climatology remain the main mode of enquiry. Although often appearing somewhat repetitive many of these studies may provide information of considerable practical value. Fields which commonly benefit include weather forecasting (including specialized forecasts for interested user-groups), urban and architectural design, heating and air-conditioning engineering, construction, air pollution modelling, planning and forecasting, land-use management, human comfort and health care. Although the value of many of these studies is their specialized information about a specific city, there is a need for periodic review in which the similarities of urban climates rather than their differences are stressed. This is necessary if the subject is not to become swamped by a flood of apparent trivia.

(a) Air temperature

The above comments are especially pertinent in considering thermal modification of the city atmosphere. Both the nature of these studies and their number precludes a detailed review. Instead, a list of most of the recent studies is provided in Table 5, and a general overview is provided which attempts to emphasize some of the more significant advances. The content of Table 5 reveals a strong research thrust across most of the world, but even so does not claim to have included all works published in the period.

Probably the single most important development in the study of heat islands since the Brussels Symposium is the increase in our knowledge of the vertical temperature structure (see Table 5). Much remains to be done, but it is clear that the

Table 5

Summary of Urban Heat Island Studies Published in the Period 1968-March 1973 [Excluding those Published in WMO (1970) Tech. Note 108]

Author(s)	Locations
1968	
*Bornstein	New York, N.Y.
Lawrence	Manchester, U.K.
Lindqvist	Lund, Sweden
Ludwig and Kealoha	Albuquerque, N.M., Dallas, Denton, Tex., New Orleans, San José, California
Oke	Montreal, P.Q.
Riquelme de Rejón	Mexico City, Mex.
Tag	Denver, Colo., Buffalo, N.Y.
Tamiya	Tokyo, Japan
Woollum and Canfield	Washington, D.C.
1969	
Ashwell	Calgary, Alta.
*Baker *et al.*	Minneapolis-St. Paul, Minn.
Bell	Winnipeg, Man.
*Clarke	Cincinnati, Ohio

*Includes vertical temperature structure.

Table 5 (Continued)

Author(s)	Locations
Dmetriev	Moscow, U.S.S.R.
*East	Montréal, P.Q.
Findlay and Hirt	Toronto, Ont.
Goldreich	Johannesburg, S.A.
Hamm	Stuttgart, Germ.
Kingham	Christchurch, N.Z.
Lawrence	London, U.K.
Lentini	Sacramento, California
*Longley	Calgary, Edmonton, Alta.
Munn *et al.*	Toronto, Ont.
Oke	Montréal, P.Q.
*Yap *et al.*	Montréal, P.Q.
1970	
*Bowne and Ball	Ft. Wayne, Ind.
*Davis and Pearson	Ft. Wayne, Ind.
Dettwiller (b,c)	Paris, France
Goldreich	Johannesburg, S.A.
Kopec	Chapel Hill, N.C.
Lindqvist	Lund, Malmö, Sweden
O'Sullivan	Newcastle, U.K.

Table 5 (Continued)

Author(s)	Locations
Pool	Chichester, U.K.
Preston-Whyte	Durban, S.A.
*Roberts *et al.*	Chicago, Ill.
1971	
*Bach (a,b)	Cincinnati, Ohio
Chopra and Pritchard	Norfolk, Va.
Conrads & van der Hage	Utrecht, Neth.
*East	Montréal, P.Q.
Fonda *et al.*	Bellingham, Washington
Goldreich	Johannesburg, S.A.
*Hilst and Bowne	Ft. Wayne, Ind.
Jacobs	Minneapolis-St. Paul, Minn.
Lawrence (a)	London, U.K.
Lewis *et al.*	Washington, D.C.
Maisel	Columbia, Md.
Maxwell	Montréal, P.Q.
*McElroy	Columbus, Ohio.
Nicholas	Washington, D.C.
*Oke and East	Montréal, P.Q.
Sekiguti *et al.*	

CONTINUED

Table 5 (Continued)

Author(s)	Locations
Wilton	Pietermaritzburg, S.A.
Wood	Austin, Texas
1972	
Beuchley *et al.*	New York, N.Y.
*Bornstein *et al.*	New York, N.Y.
Clarke (a)	New York, N.Y., St. Louis, Mo
Chopra and Pritchard (a,b)	Norfolk, Va.
Eagleman *et al.*	Kansas City, Topeka Lawrence, Kans.
Emslie	Vancouver, B.C.
Fosberg *et al.*	Fort Collins, Colo.
*Hage (b)	Edmonton, Alta.
Holmes and Wright	Lethbridge, Alta.
Landsberg and Maisel	Columbia, Md.
*Ludwig and Dabberdt	St. Louis, Mo.
Moffitt	London, U.K.
Norwine	Schaumburg nr. Chicago, Ill.
Oke (b)	Montréal and 9 settlements, P.Q.
Oke and Fuggle, Oke, Maxwell and Yap	Montréal, P.Q.
*Raman, Raman and Kelkar	Bombay, India
Runnels *et al.*	Houston, Tex.

Table 5 (Continued)

Author(s)	Locations
Sekiguti (a), Sekiguti *et al.*	First West-Ageo, Takiyama, Hibrigaoka, Nagano, Shinonoi Yashiro, Matsushiro, Japan
*Shaffer and Cohen	Philadelphia, Pa.
Sharon and Koplowitz	Ashdod, Israel
Spangler, Spangler and Dirks	St. Louis, Mo.
*Tyson *et al.*	Johannesburg, S.A.
*Wagner and Peschier	Austin, Tex.
1973	
Clarke and Peterson	St. Louis, Mo.
Hirt and Shaw	Toronto, Ont.
Oujezdsky	Austin, Tex.
Padmanabhamurty and Hirt (unpubl.)	Toronto, Ont.
Peschier	Austin, Tex.

thermal influence of a large city commonly extends up to 200-300 m and even to 500 m and more. The decreases in frequency and intensity of inversions, due to increased thermal and mechanical convection, is now a well established fact (Bornstein, 1968; Baker *et al.*, 1969; Yap *et al.*, 1969). Summers (1964) suggested that this increased mixing would result in an adiabatic layer over the city, and that this urban boundary layer could be visualized to develop in a manner similar to flow of air over a heated plate. The boundary layer height would increase with height downwind of the urban/rural leading edge as a response to the urban heat input. For the critical nocturnal case, stable rural air advecting across a uniformly heated city would produce a modification proportional to the square root of the distance of travel, and hence the heat island size and depth will be related to the city size, the heat input, and the rural lapse rate.

Clarke (1969) was the first to verify observationally, the Summers urban boundary layer concept, using helicopter traverses across Columbus, Ohio. Further evidence is given by Oke and East (1971) for Montréal, P.Q. and by Tyson *et al.* (1972) for Johannesburg. Clarke (1969) also recognized the development of another rural boundary layer at the trailing edge of the city. This resulted in a warm 'urban plume' extending aloft and downwind of the city. The L↓ results of Oke and Fuggle (1972) also suggest a residual warm layer downwind of the city. The extent of this effect has yet to be mapped. With weak rural stability and moderate ventilation, the model appears to work less well (Clarke, 1969; Tyson *et al.*, 1972), and with concentrated or elevated heat sources a number of internal boundary layers may develop (Oke and East, 1971). With calm air a self-contained 'urban dome' rather than a 'plume' is envisaged (SMIC, 1971). This temperature structure is conducive to a well-mixed surface layer with a distinct cap. Air pollution measurements tend to confirm a rather uniform concentration with height above the city and a sudden decrease near the top of the boundary-layer (East, 1969; Leahey, 1969; Roberts *et al.*, 1970; Bach 1971a, 1971b; Oke and East, 1971). The pollution situation is however far from being simple. As Frisken (1972) points out, at night when the deeper day-time mixing layer shrinks in depth, pollutants can remain trapped aloft if the shrinking is not due to large-scale subsidence.

The processes by which the urban boundary layer is modified are not yet fully clear. Thermal convection requires instability to enable it to affect a deep layer. The helicopter and fixed-mast data indeed show the urban layer to be decreased in stability, but except for the surface zone (< 50 m), the urban temperature profile still exhibits weak stability rather than instability. For example, Yap *et al.* (1969) show the Montréal mixed layer lapse rate to be 6°C km^{-1}; approximate calculations from Bornstein (1968) and Roberts *et al.* (1970) show corresponding values to be 0.5°C km^{-1} and 0.6°C km^{-1} for New York and Chicago, respectively. Hanna (1969) also shows the urban lapse rate in Montréal to approach 5°C km^{-1} as rural stability increases using tower data. Only in St. Louis did the lapse rate approximate the 'ideal' neutral form. Upward diffusion of heat would then have to be in the form of buoyant plumes penetrating this weakly stable layer. Alternatively, the urban modification could be due purely to roughness causing a redistribution of heat, or it may be due to radiative exchanges within the polluted urban atmosphere. Models appear to be able to produce a heat island using any of these processes in isolation. It is likely therefore that all three operate to a greater or lesser degree depending on the city, topography, season and meteorological conditions.

There have been attempts to seek some degree of generalization concerning the nature of urban heat islands based on semi-empirical or wholly empirical grounds. Noting the importance of rural cooling rates in determining heat island intensity, ΔT_{u-r}, (maximum urban/rural difference), Ludwig and Kealoha (1968) showed a good relation to exist between ΔT_{u-r} and the rural lapse rate $\Delta\theta/\Delta z$ in the form:

$$\Delta T_{u-r} = a - b\ (\Delta\theta/\Delta z) \tag{4}$$

with the constants depending on the city size. Later Ludwig (1970a) using Summers' argument relating ΔT_{u-r} to the square root of the urban fetch, suggested that ΔT_{u-r} should be proportional to the fourth root of the city's population, ϕ. Comparison between observed and calculated heat islands showed good agreement in 18 cities using the equation:

$$\Delta T_{u-r} = \phi^{\frac{1}{4}} \left[0.0633 - 0.298 \left(\frac{\Delta T}{\Delta p} \right)_r \right] \tag{5}$$

where, $(\Delta T/\Delta p)_r$ is the rate of change of temperature with pressure (°C mb^{-1}) from a rural sounding. Using an independent approach Oke (1972b) empirically corroborated the $\phi^{\frac{1}{4}}$ relation from ΔT_{u-r} observation for settlements in Québec. For clear skies he found ΔT_{u-r} (°C) to be given by:

$$\Delta T_{u-r} = \frac{\phi^{\frac{1}{4}}}{4\ \bar{u}^{\frac{1}{2}}} \tag{6}$$

where, $\bar{u}$ is the regional wind speed (m s^{-1}). Using equation (5) and Summers' model for predicting the mixing depth (h), Ludwig (1970b) and Ludwig and Dabberdt (1972) arrive at:

$$h = 29.3\ \phi^{\frac{1}{4}}\ \overline{T} \left[\frac{0.298\ (\Delta T/\Delta p)_r - 0.0633}{\overline{p}\ (\Delta T/\Delta p)_r - 0.287\ \overline{T}} \right] \quad (7)$$

where, $\overline{T}$ and $\overline{p}$ are the average temperature and pressure in the mixing layer. Comparison of lidar observations of h (Uthe, 1972), and predictions from equation (7) for St. Louis were encouraging. In another extension of Summers' model, it can be shown that ΔT_{u-r} should be given by:

$$\Delta T_{u-r} = \left[\frac{Q\ d\ (\Delta\theta/\Delta z)_r}{c_p\ \rho\ \overline{u}} \right]^{\frac{1}{2}} \quad (8)$$

where, d - city width; $(\Delta\theta/\Delta z)_r$ - rural potential temperature gradient; c_p - specific heat of air at constant pressure; ρ - air density. Hanna (1969) showed ΔT_{u-r} to be proportional to $[(\Delta\theta/\Delta z)_r/\overline{u}]^{\frac{1}{2}}$ using data from Fort Wayne, Ind., and Munn (1972) quotes Padmanabhamurty and Hirt (unpublished) as having provided further support for equation (8) using Toronto data. Equation (8) is generally to be preferred to the previous expressions on the grounds that it is dimensionally correct. Further extension of the use of equation (8) is provided by Clarke and Peterson (1972). Holding Q, d, c_p and ρ constant they used climatological estimates of $(\Delta\theta/\Delta z)_r$ and $\bar{u}$ to obtain seasonal heat island intensities for the continental United States.

Table 5 also shows an interest in the heat island effects of small settlements. For example, small towns and villages have been surveyed by Kopec (1970), Lindqvist (1970), Pool (1970), Fonda *et al.* (1971), Sekiguti (1972a), Sekiguti et al. (1972), Sharon and Koplowitz (1972) and Oke (1972b). In all cases heat islands were noted. Kopec (1970) strongly urged further study of small settlements, and Munn (1972) notes the need for denser meteorological sampling densities in small cities. The trend to smaller features also includes work on isolated intra-urban land-uses such as shopping centres (Chopra and Pritchard, 1972a, 1972b; Norwine, 1972), parks and squares (Poltarus, 1966; Clarke and Bach, 1971; Conrads and van der Hage, 1971; Herrington *et al.*, 1972; Jauregui, 1972; Oke, 1972a) and even subway tunnels (Sekiguti, 1972b). Similarly there is considerable interest in human biometeorological responses in the urban environment. McBoyle (1972) considers human perception of urban climate; and comfort, stress and biotropism is dealt with by Adamenko (1970), Clarke and Bach (1971), Adamenko and Khairullin (1972), Buechley *et al.* (1972), Clarke (1972a, 1972b), Garnier and LaFleur (1972), and Padmanabhamurty (1972). Similarly Terjung (1970a, 1970b), and Myrup and Morgan (1972) are seeking to characterize the energy exchange between man and his complex urban surroundings through the construction of numerical models and their evaluation in the field.

In conclusion we may note a few other areas of current heat island research interest. The rather neglected area of daytime heat islands is gaining some attention. Ludwig and Kealoha (1968) showed the existence of a cool centre in densely built-up cities, probably due to shading by tall buildings. Daytime analysis of the heat islands in Toronto (Munn, *et al.*, 1969; Findlay and Hirt, 1969; Hirt and Shaw, 1973) and Durban (Preston-Whyte, 1970) show the horizontal displacement of the core in response to lake (sea) breeze circulations. Diurnal, weekly, seasonal and annual variations in ΔT_{u-r} and urban/rural cooling rates have been studied by Dettwiller (1970b, 1970c), Lawrence (1971a, 1971b), Maxwell (1971), Hage (1972b), Moffitt (1972), Munn (1972), Oke, Maxwell and Yap (1972), Runnels et al. (1972) and Oujezdsky (1973). Diurnal variations show that the maximum heat island is usually attained 3-5 hours after sunset, after which ΔT_{u-r} declines towards morning. The peak is mainly attributable to very strong rural cooling after sunset; during the rest of the night the urban cooling often slightly exceeds the rural rate. Weekly variations appear to relate to the cycle of human activities, seasonal variations to climatological, vegetation and anthropogenic changes, and long-term annual variations reflect increasing urbanization.

Finally we should note that low-latitude studies are beginning to appear (Nieuwolt, 1966; Nakamura, 1966; Goldreich, 1969, 1970, 1971; Preston-Whyte, 1970; Wilton, 1971; Raman, 1972; Raman and Kelkar, 1972; and Tyson *et al.*, 1972). Heat islands are clearly evident but salient comparisons with mid-latitude results must await a greater body of information.

(b) Atmospheric humidity

The effects of urbanization upon atmospheric humidity are hard to assess, and probably small in magnitude. Even during the recent surge in urban climate literature the available studies remain modest. The availability of water is summarized by the atmospheric water balance equation:

$$\Delta s = C + E - P - \Delta c \tag{9}$$

where, Δs - net atmospheric moisture change; and Δc - net horizontal moisture exchange, and the other symbols were defined in section B. The terms C, E and P were covered previously. For the atmosphere C and E act to increase, and P to decrease, the moisture status. (It should be noted however, that as defined in section B, E includes condensation on the surface as dewfall, and it is expected that the heat island effect reduces dewfall). The vapour flux divergence term Δc has never been accurately evaluated or observed for the city. Hence we are left with the indirect evidence available from urban/rural humidity measurements.

Interpretation of urban/rural humidity differences should be approached with caution. Some investigators have suggested the city to be 'drier' on the basis of lower urban relative humidity values. Failure to recognize the role of temperature, and hence the heat island, in the relative humidity expression may prove such conclusions to be misleading. Both Chandler (1967) and Bornstein *et al.* (1972) show cases where relative humidities were lower, but absolute humidities or vapour pressures were higher, in the city. In terms of real moisture differences it is more meaningful to use absolute, or specific humidity, mixing ratio, vapour pressure or dew-point.

It appears that the general consensus is that cities may have lower atmospheric moisture levels than their surrounding rural areas, but that such differences are small (Ludwig and Keoloha, 1968; Peterson, 1969; Clarke, 1972a; Landsberg, 1972; Landsberg and Maisel, 1972). Ackerman (1971) conducted a comparison of urban/rural dew-point data for Chicago. She finds a weak seasonal, but strong diurnal variability in urban/rural differences. At night the city was more humid and it was suggested that this was due to the lack of dewfall. By day in the summer the city was less humid and this was suggested to be due to the stronger vapour input by E in the country. In the winter the increase in C and decrease in rural E was postulated to explain the city being more humid by day and night. Bornstein *et al.* (1972) show New York City to possess a 'vapour dome.' Based on five test days, helicopter soundings showed the city vapour excess to be larger in the early morning than in the afternoon. Morning excesses were about 0.9 g m^{-3}, and afternoon excesses about 0.2 g m^{-3} in the urban boundary layer. The Chicago and New York studies are the two most recent advances in urban humidity research, and both present results contrary to the previously held view. More and careful work is clearly warranted.

The introductory remarks concerning humidity expressions should not be construed to denigrate relative humidity comparisons for physiological or other studies where degree of saturation is important. Recent studies continue to support the view that urban areas decrease relative humidities (e.g., Thomas, 1971; Bornstein *et al.*, 1972; Landsberg and Maisel, 1972).

(c) Wind

In concluding the Brussels Symposium, Chandler (1970a) stressed the need for much more information on airflow in cities. Considerable progress has been made in the intervening period but the nature of the surface continues to provide an important resistance to detailed understanding. Reviews of this field have been provided by Davenport (1968), Munn (1970) and Landsberg (1972).

In the lowest 40 m of a neutrally stratified atmosphere over an extensive and uniform surface, the form of the wind profile is commonly described by the logarithmic law:

$$\bar{u}_z = \frac{u_*}{k} \ln \frac{z}{z_0} \qquad (10)$$

or the power law:

$$\bar{u}_z = \bar{u}_1 \left(\frac{z}{z_1}\right)^a \qquad (11)$$

where, $\bar{u}_z, \bar{u}_1$ - mean horizontal wind speed at levels z and 1 respectively; u_* - friction velocity; k - von Kármán's constant; z_0 - roughness length, a - empirical power law exponent. Alternatively, for a deeper air layer Panofsky (1971) quotes a log + linear relation suggested by Blackadar (unpublished):

$$\overline{u}_z = \frac{u_*}{k} \ln \frac{z}{z_0} + 144\ fz \tag{12}$$

and Davenport (1965) suggests the power law form:

$$\overline{u} = V_G \left(\frac{z}{z_G}\right)^a \tag{13}$$

where, f - Coriolis parameter (= $2\Omega\sin\phi$ with Ω - Earth's angular velocity and ϕ - latitude); V_G - gradient wind speed at height z_G. Over very rough terrain such as a city, the heights z, z_1 and z_G in equations (10-13) should be modified to read (z-d), etc. where d is a zero-plane displacement representing the level of the effective momentum sink. Typical urban z_0 values computed from wind profiles on the basis of equation (10) are given in Table 6.

It must be noted that there are very real problems in the acquisition and interpretation of z_0 as given in Table 6. As noted, equation (10) requires an adiabatic atmosphere and an extensive uniform surface. The former is a relatively common state of the urban atmosphere (section D(a)), but the latter is unlikely to be properly fulfilled in an urban setting. In addition, Munn (1970) points out that the constants in equation (10) only have their connotations when there is a balance between the

Table 6

Summary of Urban Roughness Length (z_0) Values for Urban Areas

[Caution - values are computed without allowance for zero - displacement]

Author	Location	z_0(m)	Comments
Ariel and Kliuchnikova (1960)	Kiev, U.S.S.R.	4.5	
Shiotani (1962)	Tokyo, Japan	0.4	urban/rural fringe
Shiotani (1962)	Kokubunji, Japan	0.48	urban/rural fringe
Yamamoto and Shimanuki (1964)	Tokyo, Japan	1.65 ±0.2	urban centre
Deland and Binkowski (1966)	Minneapolis-St. Paul, Minn.	2.0	suburban
Csanady *et al.* (1968)	Ft. Wayne, Ind.	3.0	urban centre
Slade (1969)	Philadelphia, Pa.	2.6	suburban
Marsh (1969)	Reading, U.K.	0.7	urban centre, open
Jones *et al.* (1971)	Liverpool, U.K.	1.23	suburban
Helliwell (1971)	London, U.K.	0.78	urban centre
Helliwell (1971)	Kew, Coryton, Hampton, U.K.	0.43	suburban
Peschier (1973)	Austin, Tex.	0.44-2.4	suburban

production and dissipation of turbulent, kinetic energy, a condition unlikely to be fulfilled in the city.

Specific problems in interpreting values in Table 6 relate to the heights of measurement, fetch variability, and the omission of d in calculating z_0. A number of the studies have exceeded the height range for applicability of equation (10) (e.g., comments of Panofsky, 1971). Some of the average z_0 estimates mask a large internal variability, and Peschier's (1973) results are instructive here. In winter at a suburban site the lack of foliage caused marked z_0 differences between along- and across-street trajectories. In summer the foliage effectively 'seals off' the streets and z_0 was almost insensitive to wind direction. Haana (1969) shows the importance of neglecting d in urban z_0 computations. By assuming a reasonable value of d = 10 m he shows the Kiev result (Table 6) to be decreased from 4.5 m to 1.5 m. Pasquill (1970b) and Helliwell (1971) note considerable problems in assigning d values to the London, Post Office Tower data using a power law. Analysis resulted in d being greater than building height, which is physically nonsensical. Use of d in the logarithmic form gave more reasonable values and decreased the computed z_0 values from 0.78 m (Table 6) to 0.70 m. Considering the importance of z_0 as a modelling parameter, and surface descriptor, it seems imperative that the values in Table 6 should be re-assessed in the light of the above comments. Perpetuation of the present state of affairs can only cause confusion and may lead to erroneous usage.

Munn (1970) suggests that instead of the z_0 values based on wind profile analysis it may be preferable to simply have estimates of the average roughness element dimensions. Lettau (1970), Fuggle (1971) and Morgan and Rogers (1972) have used measures of this type to arrive at z_0 values for cities, based on the equation of Lettau (1969). The relevance of d to this analysis is not mentioned.

Pioneering and most important research into the characteristics of turbulence in the urban environment is now becoming available. The most comprehensive data are for Ft. Wayne, Indiana (Bowne *et al.*, 1968; Graham, 1968; Bowne and Ball, 1970; Hilst and Bowne, 1971). All the results suggest that the city acts to increase the level of turbulent exchange. Measurements of turbulent intensities, and observations of the diffusion of aerosol tracers, both showed the city to produce a 30-50% increase in turbulent mixing. Turbulent intensity decreased with height in both urban and rural locations, but more quickly in the urban area. Vertical turbulence profiles in the city have also been presented by Slade (1969), Brook (1972) and Peschier (1973). Energy spectra at the urban and rural locations in Ft. Wayne showed little change in structure, and the

Kolmogorov - 5/3 power law appeared to apply for high frequencies at both sites (see also Steenbergen, 1971 for Edmonton, Alta.). The shapes of the spectral curves were however different, with the urban peak being shifted toward higher frequencies. The vertical velocity spectra show that energy is spread more evenly across a wider frequency band in the rougher more unstable urban situation. One interesting finding was the suggestion that the atmosphere has a 'memory.' So that although intensity and peak shifts were noted, a persistence in the order was maintained. Further to the discussion of boundary layer process (section D(a)) the Fort Wayne data suggest that thermal and not mechanical turbulence is most important (Bowne and Ball, 1970). On the other hand Brook (1972), working in Melbourne, Australia concluded that his measurements were dominated by mechanical turbulence. The difference may be due to the heights of measurement (≤ 28.5 m for Brook; 53 m for Bowne and Ball) but the point deserves more attention.

Increased frictional drag, and decreased stability combine to produce a change in momentum exchange over the city. The effect on horizontal windspeed is, however, difficult to isolate, especially since the identical exposure of anemometers is almost impossible to achieve, and the urban 'surface' is ill-defined. Peterson (1969), Munn (1970) and Landsberg (1972) provide reviews of urban airflow and distinguish between conditions with strong and weak regional flow characteristics. In the former case the city tends to modify the flow, in the latter it may generate its own circulation system.

Recent analysis of urban anemometer networks tend to confirm the results of Chandler (1965) for London, who showed that with strong winds urban speeds are decreased, but with light winds urban speeds are higher. He explained that this might be due to frictional retardation dominating at high speeds, but the relatively enhanced urban turbulence at low speeds may transport greater momentum toward the surface. The critical windspeed determining the effect showed a seasonal range of 3.5 to 5.5 m s^{-1}. Bornstein *et al.* (1972) show similar results for New York. Their critical windspeed was about 3.8 m s^{-1} below which they suggest the heat island pressure gradient induces accelerated flow, and above which it is decelerated due to greater surface roughness. In a series of investigations the Toronto heat island-lake breeze interactions have been studied (Findlay and Hirt, 1969; Munn *et al.* 1969; Anderson, 1971; Hirt and Shaw, 1973). The existence of a heat island circulation was demonstrated, the heat island core was noted to be displaced when augmented by a lake breeze. The characteristics of the lake-breeze front were also studied. Differences in the slope and thickness of the front did not appear to be related to urban heating and roughness, but urban heating did cause the retreating front to be poorly defined.

Constant volume (tetroon) balloons have been used to study the wind fields over New York (Hass *et al.*, 1967; Angell *et al.*, 1968), Los Angeles (Angell *et al.*, 1971a; Angell *et al.*, 1972), Columbus, Ohio (Angell *et al.*, 1971b), and Oklahoma City (Angell and Pack, 1972). The horizontal and vertical trajectories indirectly exhibit important urban effects. The results consistently show a tendency to move towards anticyclonic turning over the central urban area (towards lower pressure), followed by a cyclonic recovery downwind of the city. Direction changes of up to 10-20° are common, and may extend 30-50 km downwind. With strong winds the midday trajectories commonly exhibit a wave-like appearance. This often disappears as evening stability sets in and then the city may act as an 'obstacle' to flow, with the wind 'bending' around it from both sides. At night the effects are usually confined below 200-300 m. It is hypothesized that frictional drag is the likely cause of the turning by day, and that the heat island is the cause by night. Horizontal wind speeds at 100-200 m were decreased by up to 20% and this deceleration was most marked with unstable conditions. The tetroons also indicate upward air motion over the city in both light and strong winds. For light winds upward motion was ~0.04 m s^{-1}, and for strong winds it increased to ~1 m s^{-1}. Downwind perturbations were also noted. Tetroon-derived Reynolds stress results always showed urban values to be greatest.

Urban wind profiles using conventional balloons are reported by Clarke (1969), McElroy (1971), Ackermann (1972) and Wuerch (1972). Results of night profiles in St. Louis revealed the presence of several 'jets' (Ackermann, 1972). And in broad agreement with the tetroon studies, anticyclonic turning of the wind, urban deceleration, and restriction of effects to levels below 300 m were all noted.

In concluding this section the increasing use of wind tunnels and scale models to simulate urban wind fields must be noted. Jones and Wilson (1968) show comparisons between full-scale and a 1:500 scale model of a limited urban area, and Davis (1968) and Davis and Pearson (1970) tested a 1:1000 model of Fort Wayne, Indiana at an open site and compared with full-scale tower measurements. Antonia and Luxton (1969) and Luxton (1970) studied the development of boundary-layers over upstanding step changes in surface roughness which are very similar to the urban situation. Similar work has been undertaken by Cermak (1970), Chaudhry and Cermak (1971) and Sadeh *et al.* (1971).

In recent years meteorologists have formulated a number of numerical and other models to describe the workings of the urban boundary layer. Some have been designed to simulate the energy exchange and thermal characteristics of the surface, and others for the surface/atmosphere system. Although there is necessarily some overlap it is broadly possible to distinguish between those concerned primarily with the thermal characteristics of the city system, and those concerned with the circulation of air in and over cities. When viewed in perspective the ultimate goal of both is to predict the operation of the boundary layer mechanisms and their effects, and hence, aid in prediction of pollutant distributions and other features of use in land-use planning. To a greater or lesser degree both suffer from lack of quantitative information concerning the observed structure of the urban boundary layer (see sections A to D). But modelling studies are helping to pinpoint the key elements of the system which require most urgent research.

E. ENERGY AND CIRCULATION MODELS

(a) History of boundary layer modelling

Before the development of the two-dimensional finite-difference analogs to the dynamical equations, simpler models of the urban boundary layer were tried, such as the model of Gold (1956). In this model, circulations similar to a sea breeze developed into an urban area in otherwise calm conditions due to prescribed temperature differences. These differences decreased with distance from the urban-rural boundary, and were assumed to be constant to a height of 91 m. It was further assumed that there were no obstructions to the path of the moving air.

The model was used by Findlay and Hirt (1969) to compare computed urban breezes to observed circulations in Toronto. Results showed that the observed values of the urban breeze normal to the heat island boundary were less than the predicted values. This is reasonable, as Gold's model does not include the surface roughness factor.

A pioneering steady-state mixing layer model of the urban boundary layer was formulated by Summers (1964). As noted in section D(a) this purely thermodynamic model postulated the development of an adiabatic mixed layer of increasing depth as stable nocturnal air of rural origin traverses over a city and is altered by thermal turbulence. The growth of the mixed layer h, was related to a constant rural lapse rate $(\Delta\theta/\Delta z)_r$, constant heat input Q, and constant wind speed $\overline{u}$, by:

$$h = \left[\frac{2\ Q\ X}{c_p\ \rho\ \overline{u}\ (\Delta\theta/\Delta z)_r} \right]^{\frac{1}{2}} \tag{14}$$

where, X is the distance downwind of the urban/rural boundary. The relation for ΔT_{u-r} for the same conditions is given by equation (8). Summers also gives formulae for the case of a linearly increasing heat source. Summers lacked sufficiently accurate data to fully test the model but gained reasonable agreement with urban tower data. Pasquill (1970a) modified the model to allow for general wind and temperature profiles, but this necessitates the evaluation of additional empirical constants.

The addition of heat sinks to Summers' model by Leahey (1969) extended its use to areas downwind of the city. A simple energy balance equation was applied to both the reference rural site, and the given urban site. Thus Q in equations (8) and (14) became:

$$Q = Q_F + Q_{H_r} - \sigma\ (T_u^4 - T_r^4)\ . \tag{15}$$

In the original model Summers assumed $Q = Q_F$. Using fuel consumption data for New York City, Leahey and Friend (1971) used the model to predict the spatial variation of the mixing layer depth and surface temperature distribution. An excellent correlation was obtained between observed and predicted h values for five non-summer mornings.

Another early approach to modelling the urban boundary layer was the outdoor 1:1000 scale model of Davis (1968) and Davis and Pearson (1970), which was used to simulate the flow over Fort Wayne, Indiana. Hot wires and roughness elements were embedded in the ground, and measured temperatures at heights up to 1.5 m within the model were compared, by scaling with the Monin-Obukhov parameter; to observations at heights up to 300 m obtained from a tower in Fort Wayne. Moderate success was claimed in reproducing the stability variation across the urban complex.

A wind tunnel model of an urban area has been used by Yamada and Meroney (1971), Meroney and Yamada (1971, 1972) and Yamada (1972), to study the flow over an urban area. The depth of the model atmosphere was about 15 cm, and its horizontal extent was about 50 cm. Heat was supplied to the flow by electric heaters at the entrance to the tunnel and along its ceiling. Results of flow over heated obstacles showed changes in the flow pattern, frequent elevated inversions, less frequent surface inversions, and cross-over layers forming over the "city" during nighttime hours.

A simple statistical-harmonic model has been used by Preston-Whyte (1970) to summarize temperature distributions from 28 midday motor traverse in Durban, South Africa. Results showed that for a symmetrical heat island, the mean spatial temperature wave closely corresponded to a single sine wave. The extent of the influence of other factors, such as winds which persist from one quarter, or a steep temperature gradient at the urban edge, were also reflected in the amplitude of succeeding harmonics. Very recently Clarke and Peterson (1973) have used an eigenvector analysis to study the relation between the heat island in St. Louis and land-use and meteorological parameters. The first four eigenvectors explained 85% of the variance in the observed data. The analysis provided the basis for an empirical model.

Various attempts have been made to use the dynamic equations to simulate the structure of the urban boundary layer. The most complete models consist of a sub-surface layer, an analytical constant flux surface boundary layer, and a numerical transition layer, the top of which is assumed to be the top of the planetary boundary layer. In most cases the following assumptions are made for the upper layer: 1) the fluid is hydrostatic; 2) the flow is incompressible; and 3) horizontal advection is more important than horizontal diffusion. The equations for each layer resulting from these assumptions are given in Table 7, where k_s - soil thermal diffusivity; q - specific humidity; K_M, K_H, K_W - the eddy mixing coefficients for momentum, heat and moisture,

Table 7

Set of Dynamic Equations Used for Modelling the Flow Over an Urban Area

I. Soil layer equations

$$\frac{\partial T}{\partial t} = k_s \frac{\partial^2 T}{\partial z^2}$$

II. Constant flux surface boundary layer

$$\frac{\partial}{\partial z} (K_M \frac{\partial u}{\partial z}) = 0 \quad \text{(momentum flux)}$$

$$\frac{\partial}{\partial z} (K_H \frac{\partial \theta}{\partial z}) = 0 \quad \text{(heat flux)}$$

$$\frac{\partial}{\partial z} (K_W \frac{\partial q}{\partial z}) = 0 \quad \text{(moisture flux)}$$

III. Transition layer

A. Motion

$$\frac{\partial u}{\partial t} = - u\frac{\partial u}{\partial x} - v\frac{\partial u}{\partial y} - w\frac{\partial u}{\partial z} + fv - \frac{1}{\rho}\frac{\partial p}{\partial x} + \frac{\partial}{\partial z}(K_M\frac{\partial u}{\partial z})$$

$$\frac{\partial v}{\partial t} = - u\frac{\partial v}{\partial x} - v\frac{\partial v}{\partial y} - w\frac{\partial v}{\partial z} - fu - \frac{1}{\rho}\frac{\partial p}{\partial y} + \frac{\partial}{\partial z}(K_M\frac{\partial v}{\partial z})$$

$$\partial p = - \rho g \partial z$$

B. Thermodynamic energy

$$\frac{\partial \theta}{\partial t} = - u\frac{\partial \theta}{\partial x} - v\frac{\partial \theta}{\partial y} - w\frac{\partial \theta}{\partial z} + \frac{\partial}{\partial z}(K_H\frac{\partial \theta}{\partial z}) + \frac{1}{\rho c_p}\frac{\partial Q^*}{\partial z} + Q_E$$

C. Mass continuity

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0$$

D. Ideal gas law

$$p = \rho RT$$

E. Moisture continuity

$$\frac{\partial q}{\partial t} = - u\frac{\partial q}{\partial x} - v\frac{\partial q}{\partial y} - w\frac{\partial q}{\partial z} + \frac{\partial}{\partial z}(K_W\frac{\partial q}{\partial z}) - E$$

respectively; u, v and w are the three components of the wind in the x, y and z directions, respectively; and the other symbols have been defined (see also Table of symbols).

Boundary conditions at the bottom of the soil layer, the surface, the internal boundary between the surface boundary layer and the transition layer, the top of the transition layer, and the upwind and downwind boundaries must also be specified. Expressions for the eddy mixing coefficients must be specified, and a summary of the various approaches is included in the following sections. In addition, many studies utilize a surface energy equation to predict surface temperature, and this equation will also be discussed in detail in the following section. Finally, profiles within the constant flux layer for various convective regimes have to be specified.

Attempts at using finite difference analogs to the above equations over homogeneous <u>rural</u> surfaces include the non-steady, one-dimensional models of Stevens (1959), Estoque (1959, 1963), Pandolfo, *et al.* (1963, 1964, 1965a, 1965b), Wu (1965), Krishna (1968), Luther (1969), and Sasamori (1970), who included soil moisture prediction in his model. In these models, the advection terms of the equation in Table 7 were set to zero.

The first solution over a homogeneous <u>urban</u> surface was obtained by Tag (1969), but who did not include the effects of Q_F in his surface energy equation. The only result presented in his paper was the time variation of the surface heat island. This parameter was predicted to be negative during early morning hours, as the thermal properties of the urban surfaces overcame the lower urban albedo and reduced evaporation. The afternoon heat island, contrary to observations, showed a magnitude greater than that of the nighttime heat island, which arose solely from the release of stored solar energy. A similar study was carried out by Bergstrom and Viskanta (1972), except that the radiative effects of gaseous and solid pollutants were included in their model.

Circulations that develop in otherwise calm conditions due to horizontal differences in surface heating have been studied by extending the one-dimensional finite difference planetary boundary layer models, to two- or three-dimensions. In all such extensions, soil layers have been omitted, and thus surface temperatures had to be specified. In addition, constant flux layers had to be omitted, as the profile laws are not valid in near calm conditions. Examples of such two-dimensional models, in which advection in the y-direction is ignored, due to the assumption of slab symmetry, include the sea breeze models of Fisher (1961), Estoque (1961), and Neumann and Mahrer (1971), the valley breeze model of Shieh (1967, 1971), the lake breeze model of Moroz (1967), and the urban breeze model of Delage and Taylor (1970). In the last model, prescribed differential surface cooling beneath an initially stable atmosphere, produced a symmetric double cell urban circulation pattern, and an unstable layer capped by an elevated inversion over the city.

Flows of daytime planetary boundary layers over islands surrounded by cooler ocean waters, i.e., heated islands, were studied by Malkus and Stern (1953), Stern and Malkus (1953), and Smith (1955, 1957), who obtained analytical solutions for the two-dimensional, stationary, inviscid, linearized equations by Fourier analysis. Similar solutions for flow over an urban area have also been obtained. In the linearized urban circulation model of Vukovich (1971), the Coriolis effect was ignored, and a constant linear friction was used. Results showed a rather weak, two-cell circulation system in which the depth and intensity of the circulation depended on the assumed stability of the boundary layer. A mean wind produced a displacement of the cells downstream to a location which was a function of a prescribed urban heating rate, and of the advection by the mean wind.

A similar linear model developed by Olfe and Lee (1971) investigated the coupled velocity and temperature perturbations induced by an urban heat island on a uniform flow of constant stability. The resulting urban temperature profile from the basic model (no eddy viscosity, and no Coriolis force) was described as the superposition of an eddy conduction profile, and a gravity wave, which produced an elevated inversion and a crossover layer. Additional simulations, within the linearized framework, each included one of the following: a) a three-dimensional circular heat island; b) a two-layer atmosphere having a change in stability at a suitable altitude; c) a constant eddy viscosity for the perturbed flow; or d) the Coriolis force.

Finite difference numerical models have also been used to study the two-dimensional flow over surface discontinuities. Estoque (1962) used his two-layer model (i.e., surface boundary and upper transition), to study sea breezes under various non-calm synoptic conditions. Onishi (1966), and Taylor (1969) used one-layer models to investigate steady flows of neutral planetary boundary layers over single discontinuities in surface roughness, while time dependent solutions to the same problem were obtained by Wagner (1966). Similar two-dimensional numerical models were applied to the problem of the <u>heated island</u> by Tanouye (1966), the rough, heated island by Estoque and Bhumralkar (1968, 1969a), and the trade wind circulation by Pike (1968).

Four non-linear, finite difference, two-dimensional numerical models have been developed for the flow over an <u>urban</u> area. The first was the one-layer model of Yamada and

Meroney (1971) and Meroney and Yamada (1971, 1972), which was the numerical analog to their wind tunnel model discussed earlier. They assumed constant eddy mixing coefficients, and solved the following vorticity equation, derived from the momentum equation with the assumption that the boundary layer is Boussinesq:

$$\frac{\partial \zeta}{\partial t} = - u\frac{\partial \zeta}{\partial x} - w\frac{\partial \zeta}{\partial z} + \frac{g}{T}\frac{\partial T}{\partial x} + K\frac{\partial^2 \zeta}{\partial z^2} , \qquad (16)$$

where, g - acceleration due to gravity, and

$$\zeta \equiv \frac{\partial w}{\partial x} - \frac{\partial u}{\partial z} . \qquad (17)$$

The vorticity is related to the stream function ψ by:

$$\frac{\partial^2 \psi}{\partial x^2} + \frac{\partial^2 \psi}{\partial z^2} = \zeta , \qquad (18)$$

and the stream function is related to the velocity field by:

$$u = - \frac{\partial \psi}{\partial z} \qquad (19)$$

and

$$w = \frac{\partial \psi}{\partial x} . \qquad (20)$$

The model was used by Yamada (1972) to study the mutual interactions between two heated islands, and their effects on pollutant concentrations.

The model of McElroy (1971, 1972a, 1972b), for the flow over Columbus, Ohio, was a steady-state two-dimensional version of the model of Tag (1969). It did not include vertical motions, but did include the radiative effects of water vapour on the temperature structure, as well as anthropogenic heat via a surface energy equation. No results were presented for the wind field, but observed and simulated depths of the mixed layer were in good agreement. This model was used by McElroy (1972b) to estimate the effects of alternate land-use strategies on the nocturnal urban thermal structure.

A similar model by Wagner and Yu (1972) and Yu (1973) did include vertical motions, but did not include a constant flux surface boundary layer, nor radiative effects from water vapour. Surface temperature was predicted from a surface energy balance equation which did not include a combustion heating term. A variable grid was used in the vertical, while equal grid spacing was used in the horizontal. Simulations with geostrophic wind speeds of 8-15 m s^{-1} showed the heat island to be inversely proportional to speed. The results indicate a major portion of the heat island can be produced by increased roughness, and that the addition of anthropogenic releases can be significant. The solution would not however converge for winds < 8 m s^{-1}, and hence only urban deceleration was predicted.

The non-steady, two-dimensional model of Bornstein (1972a, b) also solved the vorticity and stream function equations, as given by equations (16) to (20), with the following exceptions. Eddy mixing coefficients were a function of height, the Coriolis force was included as was the v-equation of motion, and the vorticity was given by

$$\zeta = \partial u/\partial z \ , \tag{21}$$

which is in agreement with the assumption of hydrostatic equilibrium. Surface temperature was specified by use of the observed urban and rural cooling rates of Oke and East (1971), and the surface boundary layer formulation followed that of Pandolfo, *et al.* (1965a). Results for flows over rough cities showed reduced urban speeds at heights up to several hundred meters as compared to values at corresponding heights at the upwind rural boundary. In addition, the surface wind turned towards low pressure as the flow passed over the rough city, but returned to the original upwind value some distance downwind of the city. Increased urban wind speeds, as well as turning to high pressure of the surface winds, resulted from simulation of flows over warm cities.

Simulations by the model for flows over rough, warm cities reproduced most of the features usually associated with observed urban temperature fields. Predicted wind fields consisted of patterns showing areas in which urban speeds were increased, as well as decreased, over those at the upwind rural boundary. This is in agreement with the tetroon derived urban wind field over Columbus, Ohio, as determined by Angell *et al.* (1971b). In addition, the model predicted maximum wind speeds near the surface at the downwind edge of the city as observed by Bornstein *et al.* (1972). It also verified the existence of a critical wind speed, as observed by Chandler (1965) and Bornstein, *et al.* (1972) at the upwind rural boundary, below (above) which urban speeds near the surface were increased (decreased). However, predicted wind speed differences were only about 20% of observed values.

Three-dimensional models of well-mixed atmospheres, in which the hydro-thermodynamic equations were averaged with height, were used by Lavoie (1968) to study flows over lakes, and by Spelman (1969) for flows over rough heated islands with topographic features. Fully three-dimensional models, which start from calm conditions, have been used by Thyer (1962, 1966) for the study of valley winds, and by McPherson (1968, 1970) for sea breeze studies. Three-dimensional homogeneous neutral and unstable planetary boundary layers were simulated by Deardorff (1970, 1971), while three-dimensional flows of unstable atmospheres over rough, heated islands were studied by Estoque and Bhumralkar (1969b).

Two models of three-dimensional heat islands have been formulated. The finite difference model of Atwater (1971a, 1971b, 1972a, 1972b) was an expansion of the Pandolfo *et al.* (1971) model, and included the radiative effects of both gaseous and solid pollutants. The model had a 4 by 9 grid (i.e., 6 by 16 km area), but used large scale synoptic gradients, rather than internally computed gradients. The three-dimensional model of Black *et al.* (1971) was a steady state analytical model of the convective field over a heated area in the presence of an imposed wind. The Coriolis force was neglected, and the eddy exchange coefficients were equal and a function of height only. The study predicts the formation of buoyant vortex rings along axes parallel to the prevailing wind, which illustrates the importance of using a three-dimensional treatment in the study of convective phenomena.

(b) Surface energy balance equation

Some of the above-mentioned urban boundary-layer models utilize a surface energy balance equation for the prediction of surface temperature (e.g. Tag, McElroy, Wagner and

Yu, and Atwater). Others using such an equation, but with an analytical surface boundary layer and no upper transition layer, include Myrup (1969, 1970), Bach (1970), Outcalt (1972a, 1972b), Miller *et al.* (1972), Nappo (1972) and Myrup and Morgan (1972).

Clearly the energy balance at the urban/atmosphere interface is crucial to an energetically satisfactory system-specification, but the paucity of knowledge and test data (see section C), is a considerable hindrance to its accurate modelling. Ideally each of the terms in the energy balance (equation (3)) should be specified, but few models have in fact done so. For example, only a few studies have included Q_F (e.g. McElroy, Bach and Atwater), and none have included its diurnal variation. The spatial variability of the energy fluxes has not been accounted for, nor the coupling between the below-roof and above-roof boundary layers. Similarly the role of moisture is only crudely treated. Most assume a constant surface relative humidity, rather than the preferable treatment of Sasamori (1970) who used a surface moisture balance equation in his rural model. A few models have included prediction of moisture in the urban boundary layer (e.g. Tag, Miller *et al.*, Myrup and Morgan, McElroy, Nappo, and Atwater), but none has accounted for the releases via combustion (C).

(c) Problem areas

This section attempts to summarize some of the problems associated with the modelling of the urban boundary layer.

On the problem of whether to use the primitive equations, or their vorticity form, only Thyer (1962) has reported trying both approaches. He found that "better smoothing" was obtained with the vorticity approach, as "vorticities have to be integrated with respect to space to obtain velocities, and during that integration process the order of discontinuities is reduced by one." In addition, he mentioned that the "elimination of less easily visualized quantities, such as density and pressure, give a better intuitive insight into the processes involved." However, due to recent developments in finite differencing techniques, most of the recent two-dimensional models have used the primitive equations. In either case, because of the diurnal variation in the intensity of the urban heat island, more information about the urban boundary layer can be obtained by use of non-steady models than of steady models.

All of the methods mentioned in this review have assumed hydrostatic conditions, except for the sea breeze model of Neumann and Mahrer (1971). This approximation is valid when the horizontal size of the resulting circulation cells are larger than their vertical depths, a situation that is valid for most mesoscale convection cells, including urban convection cells.

All the models using the vorticity equation made the Boussinesq approximation, in which perturbations of density are ignored, except in the buoyancy term, where they are assumed to result only from perturbations of temperature, and not pressure. This approximation is valid when the depth of the circulations are less than the scale height of density [see Spiegel and Veronis (1960)]. Thus, this assumption is also valid for urban convection cells, and its use means that the flow can be considered incompressible, an assumption made in many of the primitive-equation boundary layer models.

When the radiative flux divergence from polluted layers is included in boundary layer models, it contributes to $L\downarrow$ in the surface radiation, and therefore energy, equation. It thus alters predicted values of urban surface temperatures, as shown by Pandolfo *et al.* (1971). In addition, Atwater (1972a) concluded that the radiative effects on the thermal structure of the urban boundary layer due to pollutants "are minor, but probably are large enough to change the pollutant concentration."

On the question of whether or not to use an analytical constant flux layer, Taylor and Delage (1971) have pointed out that "the use of finite differences right down to the ground can be a very inaccurate procedure when used in conjunction with an eddy viscosity or mixing length proportional to $(z + z_0)$ or z near the ground." Clarke (1970a) and Hanna (1971) also recommended use of a surface boundary layer. If the ratio of the horizontal grid spacing to the vertical grid spacing is at least 100, then according to the results of Shir (1972), the equilibrium boundary layer that results due to readjustment after flow over a surface discontinuity, will have grown to the top of the constant flux layer. Profiles for the constant flux layer should be formulated to include the inequality of the eddy heat and momentum transfer coefficients, such as those developed by Pandolfo (1966), and used by Atwater (1972a, 1972b) and Bornstein (1972a, 1972b).

Specification of the eddy transfer coefficients for the transition layer is perhaps the most difficult problem in boundary layer modelling. Early attempts at modelling these coefficients included prescribing their distribution [e.g., see Fisher (1961)]. However, the introduction of a constant flux layer into numerical models by Estoque (1961) made the transition layer coefficients dependent on the value of K at the top of the constant flux layer, i.e.,

$$K(z) = K(z') \frac{H - mz}{H - mz'} \quad , \tag{22}$$

where z' and H are the tops of the constant flux and transition layers respectively, and m is a constant. Selection of unity for m leads to a linear decrease of K from a maximum value of z' to zero at H. Many others have used equation (22), including Estoque (1963), Tanouye (1966), Moroz (1967), McPherson (1968, 1970), Tag (1969), Neumann and Mahrer (1971), and McElroy (1971).

While Stevens (1959), Fisher and Caplan (1963), had made K(z) in their transition layers proportional to the local gradient of potential temperature, Pandolfo *et al.* (1963) make K(z) dependent on the local Richardson number by use of the following

$$K(z) = \ell^2 \frac{\partial \vec{V}}{\partial z} \Phi \tag{23}$$

where ℓ is the mixing length of Blackadar (1962), $\vec{V}$ is the horizontal vector wind, and Φ is a nondimensional wind shear, or stability function, which is dependent on the local Richardson number. Various formulations of Φ have been used [e.g., the Monin-Obukhov (1954) form for forced convection, the Priestley (1959) form for free convection, and the general forms of Holzman (1943), Crawford (1965), Dyer (1967), Deardorff (1967), and Yamamoto and Shimanuki (1966)].

However, many investigators who incorporate K(z) dependent on the local Richardson number report rapid nighttime decreases in K aloft when wind speeds are low. Resulting impediment of the downward flow of heat produces lower computed temperatures near the surface than observed [e.g., see Zdunkowski *et al.* (1967), Pandolfo *et al.* (1964), Tag (1969), Luther (1969), Sasamori (1970), and Wu (1965)].

This "problem" is related to the formation of nocturnal jets at the tops of surface based radiation inversions, as predicted by Blackadar (1957). A layer of small vertical wind shear in the regions of highest speeds causes increased local Richardson number, and thus decreased computed values of K. However, Lile (1970) observed that the largest values of K within elevated west coast temperature inversions were in these very regions. His eddy viscosity was determined from observed values of $\overline{u'w'}$.

Thus, it appears that Richardson numbers give valid estimates of atmospheric stability only in layers of strong positive wind shear. In order to overcome problems produced in numerical planetary boundary layer models as Ri becomes too large, and hence K(z) too small, various methods have been employed, for example Pandolfo *et al.* (1964) introduced a fixed minimum value for K. Estoque and Bhumralkar (1969a), on the other hand, used a

Richardson number for the entire planetary boundary layer, which was averaged through the lowest 100 m, and not dependent on local stability at upper levels.

Sasamori (1970) overcame the problem by using a third order polynomial developed by O'Brien (1970), which increased his K values at 200 m from 10^3 to 5×10^4 cm^2 s^{-1}, and wiped out an erroneously predicted surface inversion. This formulation, also used by Bornstein (1972a, 1972b) yields parabolic curves, as predicted by Blackadar (1962), Lettau (1962), and Lettau and Dabberdt (1970), and as observed at night by Elliot (1964). However, there are several limiting factors, including the assumed fixed shape, and the dependency only on the stability of the constant flux layer. In an urban area for example, the stability aloft at night is quite different than the stability near the surface. Perhaps the solution is in relating K(z) to $\overline{u'w'}$, which can be evaluated from solutions to the turbulent energy equations, as was done by Shir (1972) for a neutral planetary boundary layer.

Controversy surrounds the question of whether or not a second boundary condition on w is allowable at the upper boundary. In some models, the derivative of the incompressible form of the continuity equation, i.e.

$$\frac{\partial^2 u}{\partial x \partial z} = - \frac{\partial^2 w}{\partial z^2} \qquad (24)$$

is used to justify setting w equal to zero at the upper boundary. However, it has been pointed out that the integration of equation (24) leads to a non-zero constant, and thus a violation of mass continuity. Estoque and Bhumralkar (1969b) adjusted the horizontal distribution of pressure at the upper boundary in order to assure that w at that level approached zero.

However, Delage and Taylor (1970) stated that "there is no physical necessity for having vertical velocities equal to zero at the top boundary," and Estoque and Bhumralkar (1970) showed that non-zero vertical velocities at an upper boundary have little effect on other distributions. Finally, the results of Bornstein (1972a) showed that it is not necessary to force a zero value at the top of the model, as when a physical mechanism exists, values will approach zero.

Finally, there is the problem of the inclusion of varying topography at the lower boundary, something that has not been included in any of the models described above. However, it has been included in several "large mesoscale" boundary layer models. In the model of Gerrity (1965, 1966, 1967), the flow was computed from analytical expressions involving the geostrophic and thermal winds, and the vertical co-ordinate was defined as the height above the varying surface. Terrain induced vertical motions $\hat{w}$ were computed from

$$w = u\frac{\partial e}{\partial x} + v\frac{\partial e}{\partial y} \quad , \tag{25}$$

where e is the terrain elevation. This model formed the basis of the AFGWC Boundary Layer Model, see Hadeen (1970) and Hadeen and Friend (1972).

Topography was handled in the NCAR Global Circulation model by the elimination of those grid points below ground level, see Oliger *et al.* (1970), while in the three-dimensional diffusion model of Hino (1968), a transformation of the vertical co-ordinate of the form

$$\zeta = z - e(x, y) \quad , \tag{26}$$

took into account the variable terrain.

Initial adiabatic conditions over both urban and rural regions for simulations beginning in the evening are confirmed by observations in urban and/or rural regions by DeMarrais (1961), Deland and Binkowski (1966), Kuo (1968), and Bornstein *et al.* (1972). In addition, the assumption of initially adiabatic conditions coincides with a neutral planetary boundary wind profile, which is a convenient initial wind profile. However, a neutral stratification does not represent the diurnal average state at the boundary layer. Thus, problems arise as, 1) the energy flux upwards during daytime hours is greater than the energy flux downwards during nighttime hours, and 2) K values are normally small at the top of the boundary layer. These lead to a convergence of energy at upper levels during extended simulations as shown by Bornstein (1972a). The daytime upward flux of heat could be reduced by the assumption of an initially stable planetary boundary layer, which would better represent average diurnal conditions.

Staggered grid configurations have advantages over those in which all parameters are located at the same grid points. For example, the interlaced grid of Fromm (1964), has the advantage that velocity components are considered as average inflow and outflow rates on horizontal and vertical elements of a cube. In addition, recent studies, [e.g., Clarke (1970a), Taylor and Delage (1971), and Hanna (1971)], have discussed the superiority of variable grid spacing in achieving high resolution near the surface and near horizontal boundaries.

A large pseudo-viscosity is associated with the widely used "upstream" differencing scheme for the advective terms [see Pandolfo *et al.* (1971)]. However, the "donor cell method" of advection has smaller amounts of pseudo-viscosity, and is mass conservative, as the advective terms are evaluated in a flux form [see Bornstein (1972a)]. When a flux is advected to a grid point, it is added to that grid, and an equal amount is removed from the grid supplying the flux.

Three-dimensional models are superior to two-dimensional slab-symmetric models, as cities are not infinitely wide. However, fully three-dimensional models can only be run on the very largest computers. The ultimate goal of urban boundary modelling is real time forecasting. This might be achieved if real time synoptic forecasts could be used as boundary conditions for mesoscale models.

(d) Suggested work

The ideal future boundary layer model might be a fully three-dimensional time dependent model, which had as its upper boundary conditions real time synoptic forecasts. Lower boundary conditions would be obtained from the solutions to the balance equations for heat and moisture, including time dependent anthropogenic heat and moisture terms. Topography-induced vertical velocities would be included, and eddy-diffusion coefficients would be derived from solutions to the turbulent energy equation. Finite difference techniques, coupled to variable, staggered grids, would be selected to insure maximum reliability of the numerical finite difference equations.

This model would then be coupled, via a radiative flux divergence term in the energy equation, to a similar finite difference mesoscale air pollution transport and diffusion model, which included photochemical transformations, and other natural removal processes. The system of models would then be applied to various mesoscale urban/rural regions.

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